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A global wave parameter database for geophysical applications. Part 3: Improved forcing and spectral resolution

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Abstract :

Numerical wave models are used for a wide range of applications, from the global ocean to coastal scales. Here we report on significant improvements compared to the previous hindcast detailed in Part 2 of the present study by Rasclé and Arduin (2013). This result was obtained by updating forcing fields, adjusting the spectral discretization and retuning wind wave growth and swell dissipation parameters. Most of the model calibration and performance analysis is done using significant wave heights (Hs) from the recent re-calibrated and denoised satellite altimeter data set provided by the European Space Agency Climate Change Initiative (ESA-CCI), with additional verification using spectral buoy data. We find that, for the year 2011, using wind fields from the recent ERA5 reanalysis provides lower scatter against satellite Hs data compared to historical ECMWF operational analyses, but still yields a low bias on wave heights that can be mitigated by re-scaling wind speeds larger than 20 m/s. Alternative blended wind products can provide more accurate forcing in some regions, but were not retained because of larger errors elsewhere. We use the shape of the probability density function of Hs around 2 m to fine tune the swell dissipation parameterization. The updated model hindcast appears to be generally more accurate than the previous version, and can be more accurate than the ERA5 Hs estimates, in particular in strong current regions and for Hs > 7 m.

Highlights

► An update on the Rasclé and Arduin (2013) wave hindcast is presented. ► ERA5 wind forcing gives best results but requires a bias correction for high winds. ► Satellite altimeter data was used to adjust swell dissipation parameters.

Keywords : Wind-generated waves, WAVEWATCH III

11 1. Introduction

12 Spectral wave models are routinely used for many applications in Earth
13 sciences and ocean engineering. Global and regional wave datasets gener-
14 ated through models such as WAM (WAMDI Group, 1988; Bidlot, 2005) or
15 WAVEWATCH III[®] (The WAVEWATCH III[®] Development Group, 2019)
16 have helped to improve our understanding of the wind-generated wave dy-
17 namics, estimate ocean-atmosphere interactions (e.g. surface drift and air-sea
18 fluxes), analyze extreme events occurrences, define operational conditions for
19 shipping, offshore and port activities, and assess wave energy resources, just
20 to name a few examples. New applications, for example in seismology (e.g.
21 Lecocq et al., 2019) or infrasound monitoring (De Carlo et al., 2021) are
22 made possible by the ever increasing quality of modeled wave spectra and
23 associated parameters.

24 The global hindcasts presented in Part 1 (Rascle et al., 2008) and Part 2
25 (Rascle and Ardhuin, 2013), and the Arctic hindcast of Stopa et al. (2016b)
26 are unique in providing wave parameters in an "Earth System" context, in-
27 cluding wave-related fluxes of momentum and energy between the ocean,
28 atmosphere and sea ice. These hindcasts have been used in a wide range of
29 applications, including as a source of boundary conditions for coastal models
30 (Roland and Ardhuin, 2014; Boudière et al., 2013), air-sea fluxes and upper
31 ocean mixing (Wunsch and Ferrari, 2009), surface drift of kelp or plastics
32 (Fraser et al., 2018; Onink et al., 2019; Dobler et al., 2019), and the investi-
33 gation of microseisms (e.g. Nishida and Takagi, 2016; Retailleau et al., 2017).
34 For most open ocean regions, the accuracy of significant wave height (H_s) es-
35 timates is typically better than 10%, with great benefits for the safety of life
36 at sea, but for some regions, in enclosed seas, regions of strong currents, and
37 near the sea ice, H_s errors typically exceed 20%, and other parameters can
38 be much less accurate, in particular the shape of the frequency spectrum, the
39 height of swells or the directional spreading (Stopa et al., 2016b). The rea-
40 sons for these errors, and some first steps to reduce them, are the main topic
41 of the present paper. In general the quality of numerical wave model out-
42 put is a function of at least three factors, in decreasing order of importance.
43 First, the accuracy of forcing fields (e.g. Cavaleri and Bertotti, 1997), sec-
44 ond, the realism of parameterization of processes representing spectral wave
45 evolution (e.g. Ardhuin et al., 2010) and third, the numerical choices made
46 to integrate the Wave Action Equation, namely discretization and numerical
47 schemes (e.g. Tolman, 1995; Roland and Ardhuin, 2014).

48 The present paper presents the effect of adjustment to model parameter-
49 izations in section 2, the impact of forcing field choices in section 3, and the
50 influence of model discretization in section 4. We briefly discuss in section 5
51 alternative parameterizations that can lead to clear improvements for some
52 parameters most sensitive to the higher frequencies of the wave spectrum but
53 that, so far, have not led to improvements in H_s estimates and will probably
54 require further adjustments and have thus not yet been used for the hind-
55 cast presented here. The global validation presented in section 6 shows a
56 clear improvement on sea state parameters produced by [Rascle and Ardhuin](#)
57 (2013) and, for specific conditions, also an improvement on the H_s estimates
58 in the ERA5 reanalysis. Conclusions follow in section 6.

59 2. Model setup

60 2.1. Forcing fields

61 Because waves are forced by the wind, are damped by sea ice, and are
62 strongly modified by currents, any improvement in the knowledge of these
63 three forcing fields should result in better wave model results.

64 One of the main features in the generation of the wave hindcast analyzed
65 in the present study, is the utilization of the wind fields from the fifth genera-
66 tion ECMWF atmospheric reanalyses of the global atmosphere, ERA5 ([Hers-
67 bach et al., 2020](#)), and the introduction of satellite-derived merged surface
68 current product that combines geostrophic and Ekman currents, as produced
69 by the Copernicus Marine Environment Monitoring System (CMEMS). The
70 ERA5 reanalysis was developed using 4D-Var data assimilation from the Inte-
71 grated Forecast System (IFS) model cycle 41r2. The number of observations
72 assimilated from different measurement sources goes from 0.75 million per
73 day in 1979 to approximately 24 million in 2018. The hourly output wind
74 fields with a 31 km horizontal grid resolution, represents a clear increase in
75 detail compared with some of its predecessors, like ERA-Interim ([Dee et al.,
76 2011](#)). Still, the limited horizontal resolution makes the ERA5 wind fields
77 less well resolved than those of recent ECMWF operational analyses that
78 use a T799 Gaussian grid with an equivalent resolution of 25 km. [Rivas and
79 Stoffelen \(2019\)](#) showed that ERA5 winds have a root mean square differ-
80 ence with the ASCAT winds that is 20% lower compared to ERA-Interim.
81 Still, at wind speeds above 20 m/s, ERA5 biases may be as large as -5 m/s
82 ([Pineau-Guillou et al., 2018](#)), which should have a very important impact on
83 waves modeled with ERA5 winds.

84 The surface current fields were taken from the CMEMS-Globcurrent prod-
 85 uct (Global Ocean Multi Observation Product, MULTIOBS_GLO_PHY_RE-
 86 P_015_004), with a resolution of 3 hour in time, and 0.25 degrees in latitude
 87 and longitude. This current field is the sum of geostrophic and Ekman com-
 88 ponents based on the method of Rio et al. (2014), using an updated mean
 89 dynamic topography (MDT) from CNES-CLS (Mulet et al., 2021), which is
 90 key for the reconstruction of the ocean absolute dynamic topography from
 91 altimetry data. With the geostrophic approximation, the MDT is used to
 92 estimate surface currents.

93 Finally, the ice concentration is taken from the Ifremer SSMI-derived daily
 94 product (Girard-Ardhuin and Ezraty, 2012). For ice thickness, that matters
 95 most near the ice edge where it is poorly known, we have used a constant
 96 1 m ice thickness. Partial blocking of waves by icebergs is represented fol-
 97 lowing Ardhuin et al. (2011) using the Ifremer-Altiberg icebergs distribution
 98 database Tournadre et al. (2015).

99 2.2. Adjusted parametrizations and parameters

100 Atmosphere-wave interactions include both wave generation as parametri-
 101 zed by Janssen (1991) with modifications by Bidlot et al. (2005, 2007) and
 102 swell damping caused the air-sea friction effect described by Ardhuin et al.
 103 (2009). The details and adjustments of these parametrizations are described
 104 in Ardhuin et al. (2010), and Leckler (2013). Here we only recall equations
 105 where the parameters that we have tuned in the present work are included.
 106 A more comprehensive description can be found in The WAVEWATCH III®
 107 Development Group (2019).

108 In particular, the wind input source term was reduced by using a modified
 109 friction velocity u_* with a frequency dependent term u'_* , similar to what was
 110 done by Chen and Belcher (2000). Eqs. (20) in Ardhuin et al. (2010) is

$$111 S_{\text{atm}}(f, \theta) = S_{\text{out}}(f, \theta) + \frac{\rho_a}{\rho_w \kappa^2} \frac{\beta_{\text{max}}}{\exp(Z) Z^4 \left(\frac{u_*}{C}\right)^2} \quad (1)$$

$$112 \times \max\{\cos(\theta - \theta_u), 0\}^p \sigma F(f, \theta) \quad (2)$$

113 where: S_{out} is the energy flux from the ocean to the atmosphere (swell dissi-
 114 pation term), $Z = \log(\mu)$, with μ the dimensionless critical height as given by
 115 Janssen (1991, eq. 16). ρ_a is the air density, ρ_w the water density and κ is
 116 von Kármán's constant. C is the wave phase speed, θ the wave direction, θ_u

115 the wind direction, and σ the wave relative frequency ($2\pi/f_r$, observed from
116 a reference frame moving with the mean current).

117 In eq. (1) β_{\max} is a non-dimensional wind-wave growth coefficient that
118 has been used as a tuning parameter to calibrate for wind strength biases
119 (e.g. [Stopa et al., 2019](#)). We will revisit this tuning for ERA5 winds in the
120 present paper.

121 The swell dissipation parameterization is based on observations of ocean
122 swell evolution from satellite data ([Ardhuin et al., 2009](#)). It includes expres-
123 sions to take into account the effects of the transitions from (linear) viscous
124 boundary layer to (non-linear) turbulent boundary layer. The smoothing
125 between these two regimes accounts for the Rayleigh distribution of wave
126 heights ([Perignon et al., 2014](#)). The negative part of the wave-atmosphere
127 interaction, is thus parameterized as follows,

$$S_{\text{out}}(k, \theta) = r_{\text{vis}} S_{\text{out,vis}}(k, \theta) + r_{\text{tur}} S_{\text{out,tur}}(k, \theta), \quad (3)$$

128 where the two weights give the relative importance of viscous and turbu-
129 lent attenuation, and are controlled by the ratio of the significant Reynolds
130 number $\text{Re} = 2u_{\text{orb,s}}H_s/\nu_a$ and its critical value Re_c .

$$r_{\text{vis}} = 0.5 [1 - \tanh((\text{Re} - \text{Re}_c)/s_7)] \quad (4)$$

$$r_{\text{tur}} = 0.5 [1 + \tanh((\text{Re} - \text{Re}_c)/s_7)]. \quad (5)$$

131 Based on the analogy with oscillatory bottom boundary layers, Re_c was ini-
132 tially set to 1.5×10^5 .

133 Wave energy loss to the ocean is dominated by wave breaking, and param-
134 eterized following the saturation-based breaking ideas of [Phillips \(1985\)](#). An
135 ad hoc "cumulative term" was added to enhance the dissipation of relatively
136 short waves ([Banner and Morison, 2006](#); [Ardhuin et al., 2010](#)). Alternatives
137 are discussed in section 5.

138 Finally, to reduce computational costs, we have used the Discrete Interac-
139 tion Approximation (DIA [Hasselmann and Hasselmann, 1985](#)), to represent
140 the 4-wave nonlinear interactions. This rather crude parameterization in-
141 duces errors that are partly corrected by the other adjusted source terms in
142 the Wave Action Equation ([Banner and Young, 1994](#)).

143 *2.3. Spectral and spatial discretization*

144 The wave spectrum is discretized in 24 directions, equivalent to a 15°
 145 directional resolution, and 36 exponentially spaced frequencies from 0.034 to
 146 0.95 Hz, with a 1.1 increment factor from one frequency to the next. The
 147 selected frequency range represents a departure from our previous hindcasts
 148 (Rascle and Ardhuin, 2013), in which a narrower frequency range was em-
 149 ployed, from 0.037 to 0.71 Hz. Although the parameterizations used here
 150 are not very accurate for frequencies above 3 times the wind sea peak (e.g.
 151 Peureux et al., 2018), the extension to higher frequencies allows to better
 152 capture the variability of the wave spectrum for very low wind speeds or
 153 very short fetches. The lower frequencies are there to let the spectrum de-
 154 velop for the most severe storm cases (Hanafin et al., 2012). We have used
 155 the third order Upwind Quickest advection schemes (Leonard, 1991) for both
 156 spatial and spectral advection, and the correction for the Garden Sprinkler
 157 Effect proposed by Tolman (2002).

158 All the model testing and tuning presented in section 2 was performed
 159 over a near-global grid with a spatial resolution of 0.5°, from 78° S to 83° N
 160 in latitude. However, all the other results, including the final hindcast, use
 161 a multi-grid system (Tolman, 2008; Chawla et al., 2013) in which regional
 162 grids provide a refinement near the coasts, the ice edge, and in regions of
 163 strong currents. A total of 7 nested grids were placed within the global
 164 grid, 6 regular grids and 1 curvilinear grid for the Arctic region. Details of
 165 the nested grids are provided in table 1 and Fig. 1. As shown in Fig. 1,
 166 the boundaries of the high resolution domains (in color) generally follow the
 167 coast at 500 km distance, including regions around Hawaii and the Tuamotus
 168 for the East Pacific grid, and the Azores for the North-East Atlantic grid.
 169 The regions in white are only covered with the global 0.5 degree resolution.
 170 The boundary conditions from a lower rank grid are taken at the edges of
 171 the colored regions in Fig. 1, and the higher rank grid results are spatially
 172 averaged to give the lower rank grid solution where these overlap (Tolman,
 173 2008).

174 The benefits of the multi-grid system are particularly discussed in section
 175 4.1. Compared to Rascle and Ardhuin (2013), including the Arctic grid
 176 allowed to provide a truly global wave hindcast.

177 *2.4. Model tuning*

178 The value of β_{\max} in eq. (1), s_7 and Re_c in eqs. (4) and (5) have been
 179 adjusted to minimize the model differences against satellite altimeter mea-

Sub-Grid Name	Region	Grid type	Spatial resolution	Rank
ATNE-10M	North-East Atlantic	regular	1/6°	2
ATNW-10M	North-West Pacific	regular	1/6°	3
AFRICA-10M	Africa	regular	1/6°	3
PACE-10M	East Pacific	regular	1/6°	2
CRB-3M	Carribean Sea	regular	1/20°	3
NC-3M	New Caledonia and Vanuatu	regular	1/20°	3
ARC-12K	Arctic Ocean	curvilinear	12 km	4

Table 1: Nested grids characteristics. Global grid is defined as rank 1.

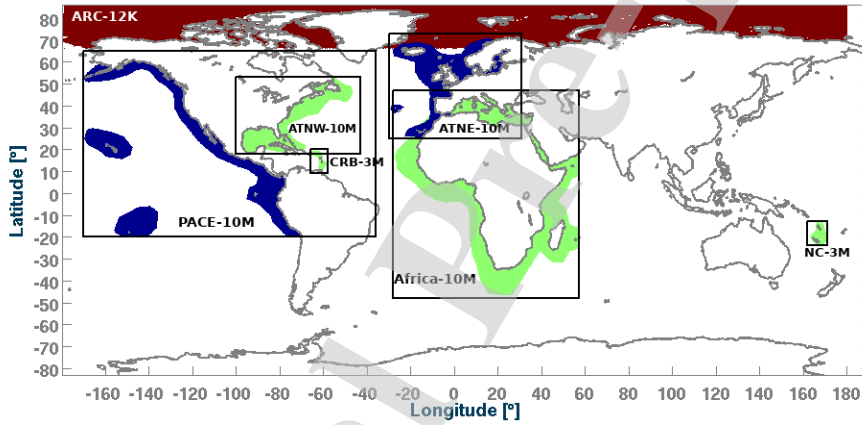


Figure 1: Sub-Grids nesting layout for multi-grid tests. Colors indicate areas where computations are performed and grids' rank in the nesting scheme: Blue is rank 2, Green is rank 3, and Red is rank 4.

180 surements of H_s by the Jason-2 mission for the year 2011, using the European
 181 Space Agency Climate Change Initiative data set (Dodet et al., 2020). We
 182 use a full year for calibration to properly sample all types of sea states in
 183 all seasons, and the year 2011 has been chosen because it had the highest
 184 wave heights ever recorded Hanafin et al. (2012), and this allows a sampling
 185 of the most extreme conditions. The variable used is the "denoised" signifi-
 186 cant wave height, at 1 Hz (approximately 7 km) resolution. The model tests
 187 performed and associated parameter values are listed in table 2. All test sim-
 188 ulations are 1-year hindcasts with data output frequency of 3 hours. These
 189 tests also include some wind bias correction. This correction is defined as a

190 piece-wise linear correction, with modeled wind speeds above U_c multiplied
 191 by a factor x_c as follows,

$$U_{10,\text{corr}} = U_{10,\text{raw}} + x_c \max\{U_{10,\text{raw}} - U_c, 0\}. \quad (6)$$

Name for set of parameters	β_{\max}	s_7	Re_c	U_c (m/s)	x_c
T471f	1.33	3.60×10^5	1.50×10^5	–	–
T471	1.43	3.60×10^5	1.50×10^5	–	–
Bm1.5	1.50	3.60×10^5	1.50×10^5	–	–
Bm1.65	1.65	3.60×10^5	1.50×10^5	–	–
Bm1.7	1.70	3.60×10^5	1.50×10^5	–	–
Bm1.75	1.75	3.60×10^5	1.50×10^5	–	–
Bm1.65-W01	1.65	3.60×10^5	1.50×10^5	20	1.05
Bm1.65-W02	1.65	3.60×10^5	1.50×10^5	21	1.05
Bm1.65-W03	1.65	3.60×10^5	1.50×10^5	23	1.08
Bm1.65-W04	1.65	3.60×10^5	1.50×10^5	22	1.05
Bm1.7-W02	1.70	3.60×10^5	1.50×10^5	21	1.05
Bm1.7-W03	1.70	3.60×10^5	1.50×10^5	23	1.08
Bm1.7-W04	1.70	3.60×10^5	1.50×10^5	22	1.05
Bm1.75-W02	1.75	3.60×10^5	1.50×10^5	21	1.05
Bm1.75-W03	1.75	3.60×10^5	1.50×10^5	23	1.08
Bm1.75-W04	1.75	3.60×10^5	1.50×10^5	22	1.05
Bm1.75-W02-s7-01	1.75	3.96×10^5	1.50×10^5	21	1.05
Bm1.75-W02-s7-02	1.75	4.14×10^5	1.50×10^5	21	1.05
Bm1.75-W02-s7-03	1.75	4.32×10^5	1.50×10^5	21	1.05
Bm1.75-W02-s7-03-s4-01	1.75	4.32×10^5	1.35×10^5	21	1.05
Bm1.75-W02-s7-03-s4-02	1.75	4.32×10^5	1.20×10^5	21	1.05
T475	1.75	4.32×10^5	1.15×10^5	21	1.05

Table 2: Models parameters and their adjustments, in bold, leading to run T475. All parameters not specified here correspond to the default T471 parameterization (Rascle and Ardhuin, 2013; The WAVEWATCH III[®] Development Group, 2019). Variables β_{\max} , s_7 , Re_c , U_c and x_c correspond to namelist parameters BETAMAX, SWELLF7, SWELLF4, WCOR1 and WCOR2 in the WW3 input files (see Appendix A for the full set of parameters).

192 The normalized root mean square difference (NRMSD), scatter index (SI)
 193 and normalized mean difference (NMD) were employed to assess the model
 194 - satellite discrepancy and its change when model parameterizations, forcing
 195 or discretization are modified. These statistical parameters were calculated
 196 for the entire domain and over a set of specific ocean regions (defined in table

197 3), for each 1-year test in table 2. They are defined as follows,

$$\text{NRMSD}(X) = \sqrt{\frac{\sum (X_{\text{mod}} - X_{\text{obs}})^2}{\sum X_{\text{obs}}^2}} \quad (7)$$

$$\text{SI}(X) = \sqrt{\frac{\sum [(X_{\text{mod}} - \overline{X_{\text{mod}}}) - (X_{\text{obs}} - \overline{X_{\text{obs}}})]^2}{\sum X_{\text{obs}}^2}} \quad (8)$$

$$\text{NMD}(X) = \frac{\sum (X_{\text{mod}} - X_{\text{obs}})}{\sum X_{\text{obs}}} \quad (9)$$

198 where X_{obs} and X_{mod} are the altimeter significant wave heights (denoised) and
 199 the modelled H_s respectively. In particular for the tuning process, X_{obs} is the
 200 along-track data from the altimeter, and X_{mod} is obtained by interpolating
 201 the model output in space and time from the closest 4 grid points, into the
 202 position of the altimeter measurement.

203 We note that other normalizations could be used (Mentaschi et al., 2015),
 204 and in particular a larger scatter index is not always the indication of a
 205 poorer model performance, in particular in the presence of large biases or
 206 large fluctuations.

207 We particularly looked at differences for different ranges of observed val-
 208 ues of H_s , binning all the model output as a function of the satellite values.
 209 In general, for the model's performance assessment, attention was only paid
 210 to H_s larger than 1.0 m because H_s smaller than 0.75 m is not very accurate
 211 due to limited sampling of the signal associated with the radar bandwidth
 212 (Smith and Scharroo, 2015; Arduin et al., 2019).

213 Previous parameter settings defined as "T471" were used as a starting
 214 point. After gradual increases of β_{max} without changing the other parameters
 215 (sets T471f to Bm1.75 as defined in table 2), a persistent negative NMD for
 216 H_s values larger than 7 m is found, as illustrated in Fig. 2.

217 This behavior is expected to be related to an underestimation of wind
 218 speeds in excess of 25 m/s in ECMWF IFS model results, including the
 219 ERA5 data set, as analyzed by Pineau-Guillou et al. (2018). This wind-
 220 speed dependent bias, which is not found with CFSR winds, was the main
 221 motivation for introducing the wind speed correction in eq. (6).

222 After setting $\beta_{\text{max}} = 1.75$, wind speed corrections with the parameters
 223 Bm1.75-W02 helped to reduce the wave heights underestimation in the 8-
 224 14 m range (Fig. 3).

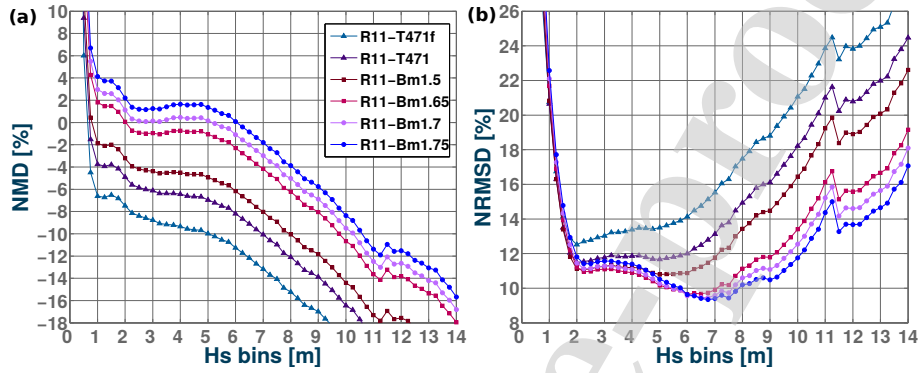


Figure 2: Error statistics for H_s for the β_{\max} sensitivity runs (a) Normalized mean difference between model runs – with parameters given in Table 2 – and the Jason-2 altimeter data, (b) normalized root mean square difference.

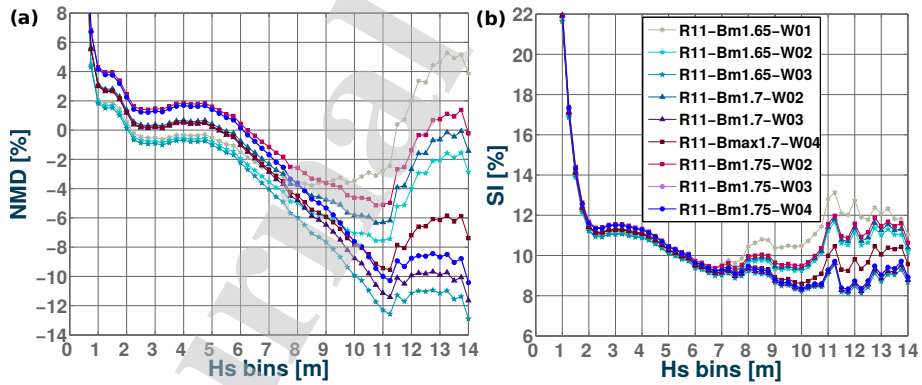


Figure 3: Error statistics for H_s for the wind correction sensitivity runs (a) Normalized mean difference between model runs – with parameters given in Table 2 – and the Jason-2 altimeter data, and (b) scatter index.

225 The wind speed U_c at which the correction kicks in is consistent with the
 226 analysis of models and in situ wind data by Pineau-Guillou et al. (2018),
 227 where it was demonstrated that typically strong winds above 20 m s^{-1} are
 228 underestimated by the ECMWF models.

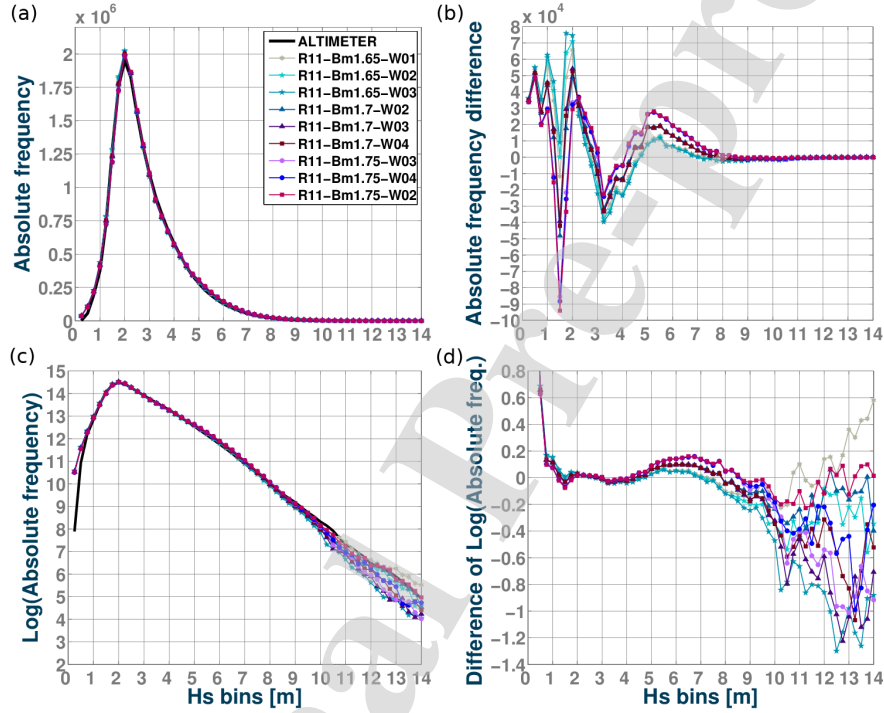


Figure 4: (a) Histogram of H_s values in the Jason-2 and co-located model simulations. (b) Differences between the model and altimeter histograms. Plots shown are from wind correction tests only. (c) Same as (a) but with a logarithmic scale. (d) Difference of logarithm of the modeled and measured H_s histograms.

229 Once the NMD and NRMSD were reduced, particular attention was paid
 230 to the distribution of H_s . The applied changes in β_{\max} and wind correction
 231 lead to more intense waves in storms and swells radiated from these storms.
 232 As a result the swell dissipation necessarily needs further tuning, which is
 233 done here by adjusting s_7 and Re_c . This adjustment can be done using
 234 wave spectra measurements from buoys, but also using the distribution of
 235 H_s . Indeed, the smoothing of swell dissipation was introduced in eq. (3) by

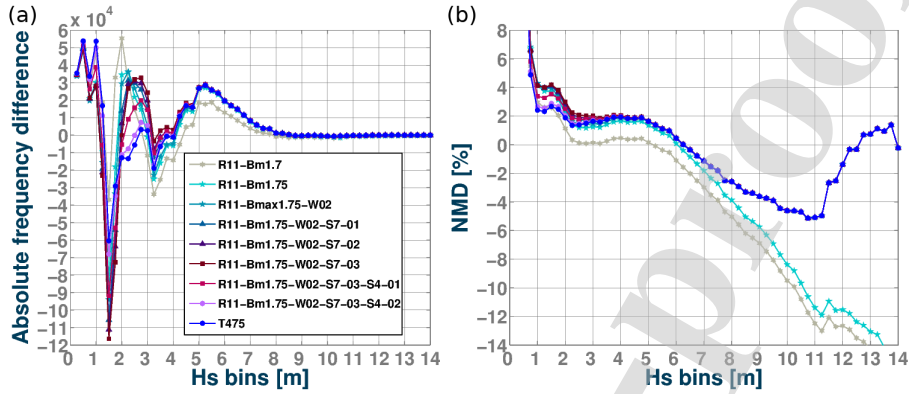


Figure 5: (a) Histogram of H_s values in the Jason-2 and model simulations absolute frequency of occurrence difference (WW3 - altimetry data). (b) Normalized mean bias. Plots shown are from s_7 and Re_c sensitivity tests.

236 Leckler et al. (2013) to correct the sharp jump around 2 m in the distribution
 237 of modeled H_s that was first noted by D. Vandemark (personal communi-
 238 cation, 2012). It was only later rationalized as an effect of the Rayleigh
 239 distribution of wave heights with turbulent boundary layers over the largest
 240 waves in a group and viscous boundary layers over the lowest waves in a
 241 group (Perignon et al., 2014; Stopa et al., 2016b). Fig. 4 shows the distri-
 242 bution of H_s in the model and observations. With panel b showing the
 243 difference between model and observation to make the differences more visi-
 244 ble for wave heights smaller than 8 m, and in panel d the difference of the log
 245 of frequency of occurrence to see the deviations for larger H_s . Augmenting
 246 s_7 from 3.6×10^5 with the parameters s7-01 to 4.32×10^5 with s7-03 spreads
 247 the transition from viscous to turbulent dissipation over a wider range of
 248 H_s and tends to smooth the histogram of H_s . This corrects the bias in the
 249 distribution around $H_s = 2.0$ m but makes things worse around 1.5 m. To
 250 correct those errors requires also shifting the transition Reynolds number Re_c
 251 to lower values in runs s4-01, s4-02 and s4-03 as shown in Fig. 5.a. These
 252 later adjustments made it possible to match the occurrence of the highest
 253 values of H_s , up to 14 m, as shown in Fig. 5.b.

254 Although H_s gives a very limited description of the sea state, the great
 255 benefit of H_s altimeter data is their global coverage, and the differences
 256 between model and observation over different regions of the world ocean can

257 also be revealing due to the different types of sea states found in these regions
 258 (Chen et al., 2002), but also due to different forcing by winds, currents and
 259 sea ice. Table 3 defines the different ocean regions for which we have looked
 260 at regional H_s statistics. Further analyses on effects over the directional
 261 spreading and other wave parameters based on in-situ measurements, are
 presented in section 5 and 6.3 respectively.

Region (basin)	Minimum Longitude [°]	Maximum Longitude [°]	Minimum Latitude [°]	Maximum Latitude [°]
North Atlantic	-80	-5	10	50
South Atlantic	-68	20	-54	-2
North Pacific	125	-100	5	60
South Pacific	150	-73	-54	-2
Indian Ocean	50	100	-30	25
Southern Ocean	-179.98	180	-70	-55
NO SOUTH	-179.98	180	-55	66

Table 3: Regions definition for performance analysis.

262 The adjustments of β_{\max} and wind intensities corrections showed particu-
 263 larly good improvements in the North and South Pacific. By only augmenting
 264 the β_{\max} value (for example in tests R11-Bm1.7 and R11-Bm1.75), an im-
 265 portant decrease of the H_s occurrences is obtained around 2 m, especially in
 266 the South Pacific, but this comes at the price of an excess of H_s values in
 267 the 1–1.5 m range (Fig. 6).
 268

269 Higher values of β_{\max} also reduced the overall negative bias in wave heights
 270 within the range of 1.5–7 m, with a further reduction of the negative NMD
 271 when the selected wind correction is applied. This specially improves the
 272 NMD for H_s of 7 to 11 m in the North Atlantic and South Pacific (Fig. 7).
 273 The South Pacific stands out as a region of high positive bias (Fig. 8).
 274

275 Although it is possible that winds in the Southern Ocean may have spe-
 276 cific biases due to a limited set of data used for assimilation, the state of the
 277 atmosphere is very much controlled by remote sensing data, including ra-
 278 diometers and scatterometers that are assimilated globally (Hersbach et al.,
 2020).

279 Another peculiarity of the Southern Ocean is the importance of the cir-
 280 cumpolar current that generally flows from West to East. Not taking it into
 281 account is known to produce a large positive bias of the order of 20 cm

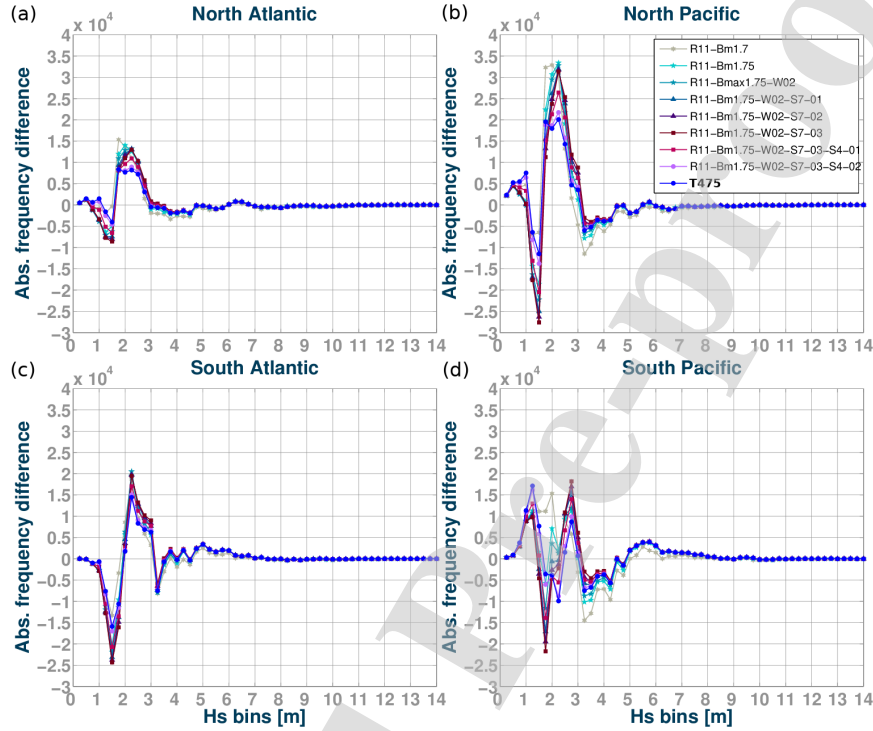


Figure 6: H_s absolute frequency of occurrence difference (WW3 - altimetry data) from Atlantic and Pacific basins.

282 in wave heights due to the relative wind effect (Raschle et al., 2008; Rapizo
 283 et al., 2018), and large gradients in H_s associated to refraction (Quilfen and
 284 Chapron, 2019). Indeed, the relevant wind speed for wave generation is the
 285 wind velocity minus the surface current velocity. However, these previous
 286 estimates use numerical models that are not very reliable for surface current
 287 estimates (ESA, 2019). Another effect specific to the Southern Ocean is the
 288 presence of both sea ice and icebergs, with a very large impact on wave heights
 289 (Ardhuin et al., 2011). The year 2011 has a rather large anomaly in iceberg
 290 numbers, although not as large as in 2009 (Tournadre et al., 2016). Finally,
 291 the details in sea ice concentration near the ice edge and the parameteriza-
 292 tions of wave-ice interactions are another important source of uncertainties
 293 at latitudes south of 55°S (Doble and Bidlot, 2013; Ardhuin et al., 2020).

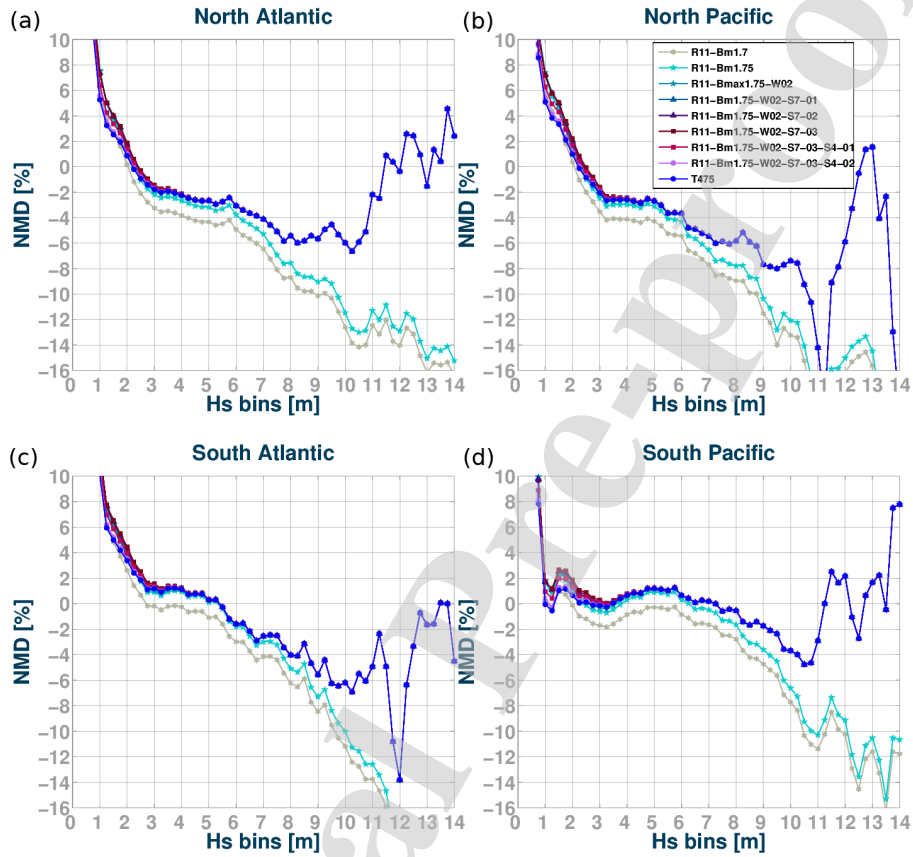


Figure 7: H_s NMD within Atlantic and Pacific basins as a function of observed wave heights. H_s bins' range is 0.25 m.

294 For these reasons, we now investigate alternative forcing fields for winds, ice
 295 and currents, and their impact on the model results.

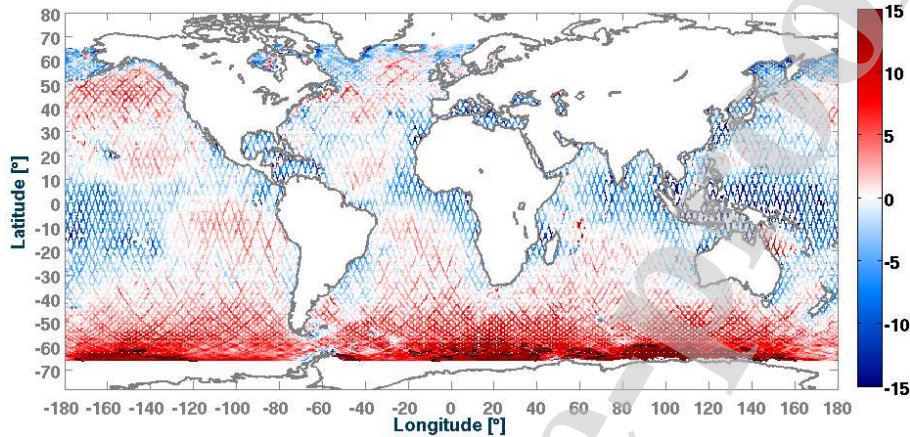


Figure 8: NMB for 1-year averaged H_s using ERA5 winds. Modelled year: 2011. Parameter settings from test T475. Colorbar indicates NMD values in %. Black lines represent positive 10 % contours.

296 3. Influence of forcing field choices

297 As we did for the choice of model parameters, forcing set-up and model
 298 adjustment was done over the year 2011, with a complete validation on other
 299 years described in section 6. Whereas we had used Jason-2 data only for
 300 the model calibration, we now use the full ESA Sea State Climate Change
 301 Initiative merged altimeter data set, using the denoised 1-Hz data for the
 302 significant wave height (Dodet et al., 2020). For the year 2011 this includes
 303 data from the following satellite missions: Jason-1, Envisat, Jason-2 and
 304 Cryosat-2. Using the model with parameters T475, our baseline model run
 305 uses ERA5 winds, Ifremer sea ice and iceberg concentrations, and CMES-
 306 Globcurrent surface currents.

307 3.1. Choice of forcing wind field

308 We now look at three alternative wind fields. These include the opera-
 309 tional ECMWF IFS winds which, for the year 2011, was obtained with IFS
 310 cycle 37r2, an earlier and less accurate version of IFS compared to the 41r2
 311 used for ERA5. We also considered the CFSR winds (Saha et al., 2010) that
 312 were used by Rasche and Arduin (2013). Finally we tested the Ifremer CER-
 313 SAT Global Blended Mean Wind Fields (Bentamy et al., 2018), from here
 314 on just named "Ifremer". Other wind fields like ERA-Interim and MERRA2

315 (Gelaro et al., 2017) have also been considered in other hindcasts such as
 316 Sharmar et al. (2021), with analyses focused on inconsistencies and trends of
 317 the different atmospheric forcing.

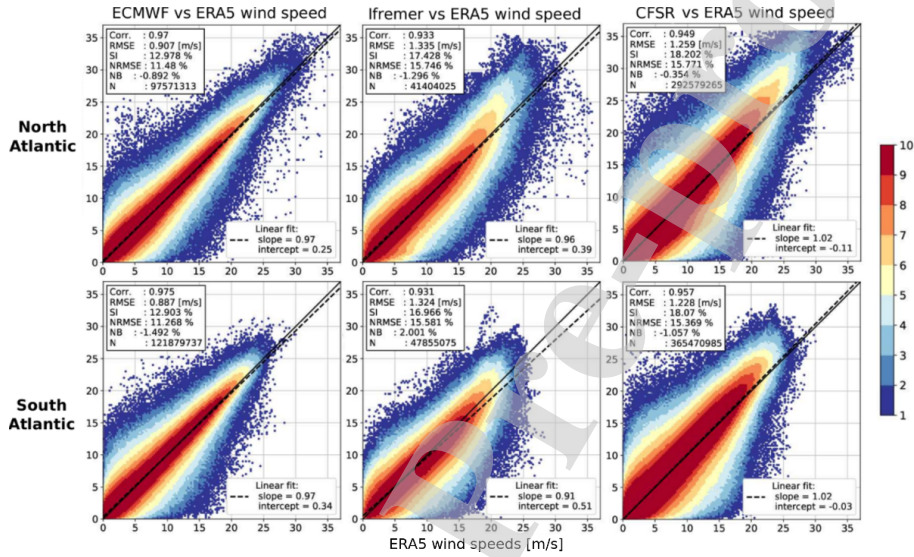


Figure 9: Scatter plot of wind speed for the months of January to July 2011. ERA5 intensity bins along x-axis. Top panels: ECMWF operational product vs ERA5, Middle panels: Ifremer vs ERA5. Bottom panels: CFSR vs ERA5. Colors give the logarithm of the number of data points in each $0.25 \text{ m/s} \times 0.25 \text{ m/s}$ wind speed bin.

318 The main difference between the Ifremer winds and the 2 other data sets,
 319 is that the Ifremer 6-hourly surface wind fields are estimated mainly from
 320 scatterometer wind vector observations, merged with wind magnitude mea-
 321 surements from radiometer data (SSM/I, SSMIS, WindSat) and the ERA-
 322 Interim atmospheric wind reanalyzes. Further details on the product and
 323 methods can be found in Bentamy et al. (2012, 2013).

324 As discussed by Rasclé and Ardhuin (2013) and Stopa et al. (2019), dif-
 325 ferent wind fields are biased relative to one another. This is true for the
 326 average values around 7 m/s, and biases are even larger for high speeds over
 327 20 m/s (Pineau-Guillou et al., 2018). This is shown again here in Fig. 9.
 328 The NCEP operational GFS model (not shown here) and CFSR hindcast
 329 both have wind speeds higher than those produced by the ECMWF models
 330 (operational IFS results and ERA5 results), leading to higher wave heights

331 when using NCEP winds. Because the Ifremer blended wind product uses
 332 ERA-Interim as a background "filler" when and where observations are too
 333 far in space or time, these data sets were homogenized to have the same low
 334 bias for average conditions (slope of 0.91 for the Ifremer wind vs the ERA5
 335 winds in the South Atlantic) but higher values for wind speeds above 20 m/s
 336 that are more frequent in the North Atlantic.

337 There is also a clear indication that ECMWF operational winds give
 338 higher values for wind speeds above 20 m/s compared to ERA5, probably
 339 due to the higher resolution of the operational IFS model (25 km approx. and
 340 hourly output for 2011). The consequences of these wind field properties on
 the wave height biases are shown in Fig. 10.

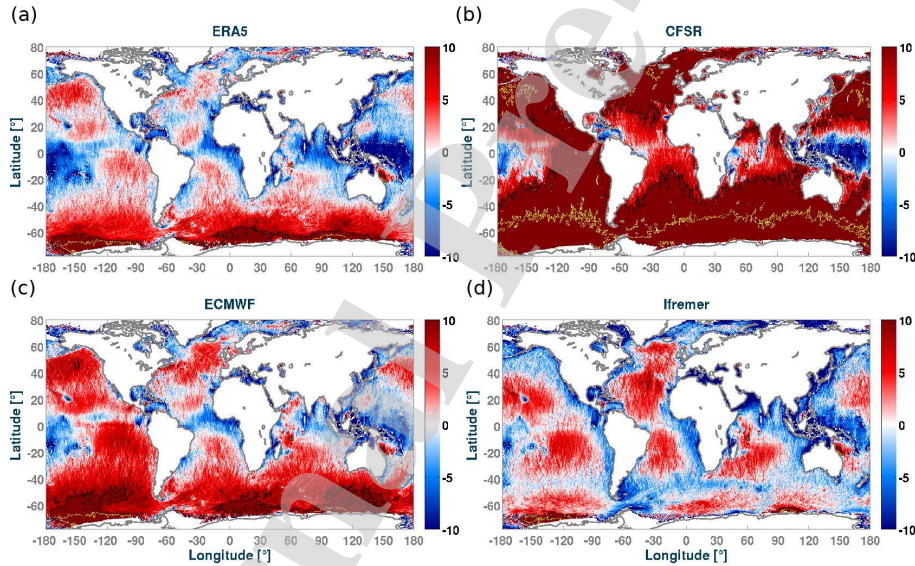


Figure 10: Normalized Mean Difference of modelled H_s minus Sea State CCI Altimeter data, averaged over the year 2011, using (a) ERA5, (b) CFSR, (c) ECMWF operational deterministic products and (d) Ifremer winds. The model was run with the set of parameters T475 as given in Table 2. Colorbar indicates NMD in percent. Black and yellow lines mark the +10 and +20 % contours.

341
 342 Given the relative biases of the different wind datasets, it is not surprising
 343 that, without any retuning, the T475 set of parameters gives large H_s biases
 344 when used with other wind forcing than ERA-5. In particular the CFSR

345 winds give positive biases larger than 15% over most of the oceans.

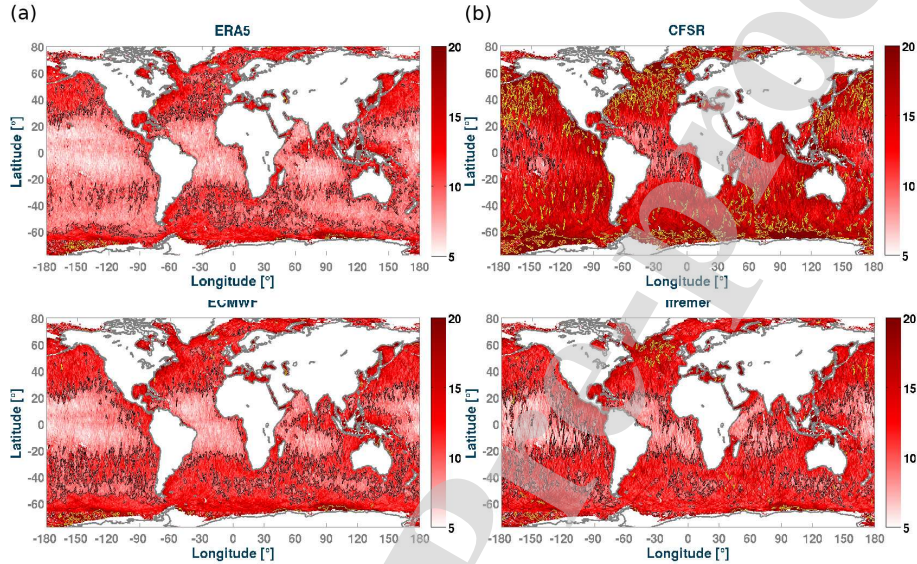


Figure 11: Scatter Index of modelled H_s minus Sea State CCI Altimeter data, averaged over the year 2011, using (a) ERA5, (b) CFSR, (c) ECMWF operational deterministic products and (d) Ifremer winds. The model was run with the set of parameters T475 as given in Table 2. Colorbar indicates SI in percent. Black and yellow lines mark the +10 and +20 % contours.

346 The Ifremer winds have interesting properties and are probably more
 347 realistic in some regions, where they give lower scatter index (Fig 11.d),
 348 including the southern ocean where the bias is also lower and significantly
 349 different (Fig. 10.d). This difference between Ifremer and ERA5 winds is
 350 possibly due to the fact that the remote sensing data used in the Ifremer
 351 product generally measures a wind that is relative to the current and not an
 352 absolute wind (Quilfen et al., 2004). There is also probably a contribution
 353 to the generally low bias of the ERA-Iterim product that is used to fill in
 354 between the different satellite passes.

355 3.2. Effects of wave-ice parametrizations and forcing fields

356 Much work has been done on the interactions of waves and sea ice in the
 357 recent years, with a large emphasis on pancake ice (Thomson et al., 2018),

358 that is particularly relevant near the ice edge and during the freeze-up period
359 (Doble et al., 2003). Here we have rather used a parameterization associated
360 to the presence of larger floes and their possible break-up induced by waves.
361 In particular the formulation we have used in our baseline simulation was
362 developed by Boutin et al. (2018) and adjusted by Ardhuin et al. (2020) to
363 2 months of waves measured in the sea ice of the Ross sea. That parameter-
364 ization combines both wave scattering in sea ice with a wave-induced ice
365 break-up (IS2) and dissipation below ice plates including a smooth laminar
366 to rough turbulent flow as a function of the boundary layer Reynolds number
367 (IC2, Stopa et al., 2016b). Given uncertainties on ice thickness, in particu-
368 lar in the Southern Ocean (Williams et al., 2014) and around the ice edge
369 where it matters for wave-ice interactions, we have chosen a crude and simple
370 constant thickness of 1 m. This parameterization is compared to the old
371 default WW3 parameterization that is a 40 km exponential decay of wave
372 energy proportional to the ice concentration (IC0 parameterization). The
373 new IC2+IS2 parameterization gives a much weaker attenuation near the ice
374 edge, and thus a larger value of H_s in the open ocean where we have data
375 for validation (Fig. 12a,b). We have not attempted to validate the predicted
376 wave parameter and maximum floe size in the ice-covered regions. We note
377 that the scatter index is generally reduced around the ice, especially around
378 Greenland and in the Ross sea. These areas typically require more valida-
379 tion, and the model resolution (0.5°) is probably marginal for the Southern
380 Ocean, whereas the 12 km resolution in the Arctic allows a more detailed
381 investigation of wave-ice interactions.

382 Much less work has been devoted to the effect of icebergs, so we use
383 here the parameterization proposed by Ardhuin et al. (2011). We verify that
384 including icebergs has a very positive effect on reducing the bias and scatter
385 index where the icebergs are present. For the year 2011, a large concentration
386 of icebergs was found in both the South-East of the Pacific and the South
387 of the Indian ocean, giving a bias reduction up to 10 percentage points and,
388 locally, a very large reduction in scatter index up to 6 percentage points
389 (Fig. 12c,d). The concentration of icebergs in the South Pacific in 2011 is
390 associated with two large icebergs, C19a and B15j, that drifted northward
391 and eastward within the Antarctic Circumpolar Current (Tournadre et al.,
392 2015, 2016), later breaking up into hundreds of smaller icebergs. These small
393 icebergs are much more effective in reducing the wave energy flux, compared
394 to a single parent iceberg, as they have a much larger cross section.

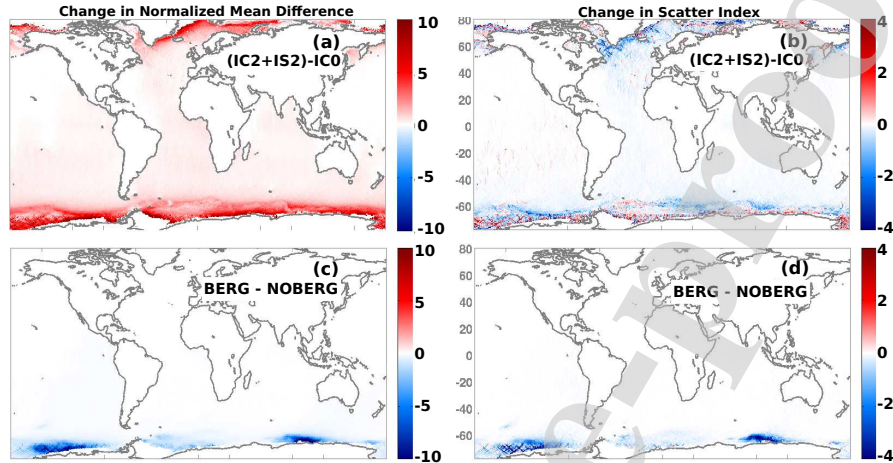


Figure 12: (a,b) using dissipation, scattering and ice break-up (IC2, IS2) or partial ice blocking (IC0) Differences in NMB and SI in percentage points for the T475 parameterization variations when using: (c,d) iceberg forcing or no iceberg forcing.

395 3.3. Effect of currents

396 Ocean surface currents can have large influences on the wave field ei-
 397 ther locally through the relative wind effect and advection, or down-wave of
 398 current gradients, due to refraction, with larger effects associated to larger
 399 current magnitude (Ardhuin et al., 2012). An important difficulty for prop-
 400 erly taking currents into account at global scales is that there are no global
 401 observations of the Total Surface Current Velocity (TSCV) that matters
 402 for wind waves, and the only proper surface measurements are made with
 403 High Frequency radar near the coasts (Barrick et al., 1974; Roarty et al.,
 404 2019). Instead, the closest global proxy is given by the drift velocity around
 405 15 m depth provided by instruments of the Surface Velocity Program (Elipot
 406 et al., 2016; Lumpkin et al., 2017), with only about 1500 drifters globally
 407 giving a 500 km resolution. We note that at the Equator and a few other
 408 places of interest, the 15-m depth drift is often in the opposite direction of
 409 the surface drift. Most importantly, finer spatial resolution is needed, typ-
 410 ically down to 30 km, to represent most of the refraction effects (Ardhuin
 411 et al., 2017a; Marechal and Ardhuin, 2020). As a result, surface current
 412 estimates are often taken from numerical models, or, which is the case of
 413 the CMEMS Globcurrent product used here, derived from combined obser-

414 vations of sea surface height anomaly, mean dynamic topography and surface
 415 winds, assuming a quasi-geostrophic equilibrium of the Coriolis force asso-
 416 ciated to the surface current with the combination of the wind stress and
 417 the pressure gradient associated to sea surface height. Except possibly for
 418 western boundary currents such as the Gulf Stream or the Agulhas, this ap-
 419 proach does not work very well, in particular around the equator and in mid-
 420 latitudes where currents are dominated by near-inertial currents as illustrated
 in Fig. 13. The CMEMS Global Ocean Multi Observation Products (MUL-

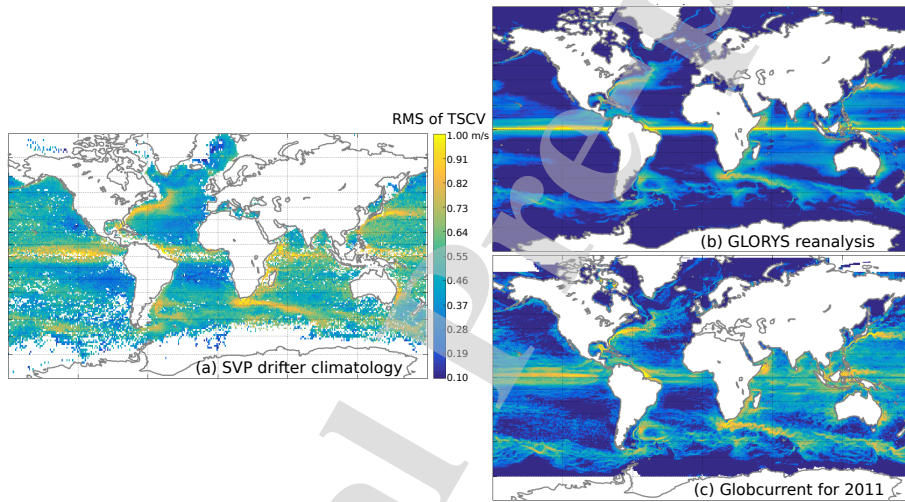


Figure 13: Root mean square current velocity (a) at 15 m depth using in situ drifter data from the Surface Velocity Program (SVP) processed by Elipot et al. (2016) with rms velocity computed over 30-day long trajectories and attributed to the center of that trajectory and white ocean pixels corresponding to 1 by 1 degree squares in which no data was available, (b) as given by the CMEMS GLOPYS reanalysis, (c) as given by the CMEMS-Globcurrent product based on altimeter sea level anomalies, mean dynamic topography inferred from satellite gravimeters and ocean drifters, and "Ekman currents" estimated from ECMWF wind analyses.

421 TIOBS_GLO_PHY_REP_015_004) has an average current that is closer to the
 422 SVP drifter climatology than the CMEMS Global Ocean Reanalysis (GLO-
 423 RYS) product GLOBAL-REANALYSIS-PHY-001-031, in particular around
 424 the Equator, which is why we have chosen to use the former product as our
 425 TSCV forcing.

427 Given all these limitations it is not specially surprising that the TSCV is

428 seldom used at global scale. Including the TSCV forcing can indeed increase
 429 errors in some regions due to errors in the forcing field, but it generally cor-
 430 rects part of the bias and gives lower scatter index for wave heights compared
 431 to altimeter data, as illustrated in Fig. 14. Comparing our simulation with
 432 parameters T475 with and without currents, we find a clear lower bias along
 433 the Equator and in the Southern ocean when currents are used, as already
 434 reported by Rascle et al. (2008). This is probably associated with the relative
 435 wind effect, with wave generation given by the difference between the wind
 436 vector and the TSCV and not the wind vector alone. We know that this
 437 approach can overestimate the current effect when the atmosphere model is
 438 not coupled with an ocean model (Hersbach and Bidlot, 2008; Renault et al.,
 439 2016), however, we also expect that the TSCV is generally underestimated
 440 by the CMEMS-Globcurrent product.

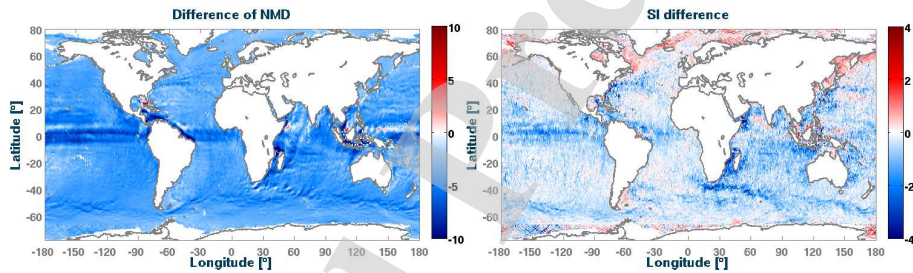


Figure 14: Left: Change in Normalized Mean Difference (NMD in percentage points)
 for H_s with currents and the T475 parameterization versus the same simulation without
 current. For both simulations the reference is the Sea State CCI H_s for the year 2011.
 Right: same for difference in SI, with the dark blue corresponding to a reduction of 4
 percentage points (e.g. from 14% to 10%) when TSCV forcing is used.

441 The reduction of the scatter index against altimeter H_s that is brought
 442 by the current (blue regions in Fig. 14.b) clearly corresponds to the regions
 443 of strong currents where the variability of incoming waves can cause a large
 444 variability of the wave heights around the current: this is the case in the Ag-
 445 ulhas current, in the Gulf Stream, the Kuroshio, the Mozambique channel,
 446 the Somali current. However, as shown in Fig. 11, these regions are still
 447 places where the models error are relatively large, possibly due to a combi-
 448 nation of factors, including errors in the TSCV fields, insufficient directional
 449 resolution of our wave model (Marechal and Ardhuin, 2020), and insufficient
 450 spatial resolution in the TSCV field and/or the wave model. We note that

451 the scatter index is generally increased for latitudes above 50° N, probably
452 due to an insufficient resolution of the altimetry where the Rossby radius
453 of deformation is less than 50 km (Ballarotta et al., 2019). Given the im-
454 portance of the spectral and spatial discretizations, we now discuss these
455 aspects.

456 4. Model discretization

457 The choice of spatial and spectral discretizations can have a large impact
458 on the model solutions, and it also has a direct and clear impact on the cost
459 of the model, the time needed to perform the simulations. As a result, the
460 particular choices we made for the discretizations are a compromise between
461 the computational cost and the accuracy benefits. The 28-years hindcast
462 used around 500,000 cpu hours distributed over 504 processors, distributed
463 in 18 nodes that each hold 28 CPUs and 75Gb of memory.

464 4.1. Spatial resolution

465 Using higher resolution grids is critical for resolving smaller scale varia-
466 tions in the sea state that are caused by the time-varying forcing fields (wind,
467 current, sea ice) or fixed features (shoreline, water depth, bottom sediment
468 type and grain size). In practice, small scale gradients in wave heights are
469 dominated by the distance to the coast and the presence of strong currents
470 (Quilfen and Chapron, 2019). Because some important current system are
471 located close to coasts, we have chosen to define nested grids that cover the
472 relatively shallow waters of the coastal regions and, where possible, extend
473 over strong current regions (Fig. 1). As a result, our North-West Atlantic
474 grid covers the Grand Banks and the Gulf Stream, as well as the entire gulf
475 of Mexico. In a similar fashion, the Africa grid was extended to the south to
476 cover the Agulhas current retroflexion. Using different grids also allows to
477 tune the model parameters locally.

478 Because the wind-wave growth tuning that corresponds to T475 is very
479 similar to T471, it tends to give an underestimation of the wave height for
480 short fetches (Stopa et al., 2016a). This effect is more pronounced with higher
481 resolution grids, which explains the general reduction in wave height for
482 enclosed seas and eastern coasts (stronger negative bias, in blue in Fig. 15.a).
483 We also find that the explicit higher resolution of shorelines and islands gives
484 larger H_s values compared to the subgrid treatment of complex shorelines and
485 islands in a coarser grid (Chawla and Tolman, 2008), explaining the more

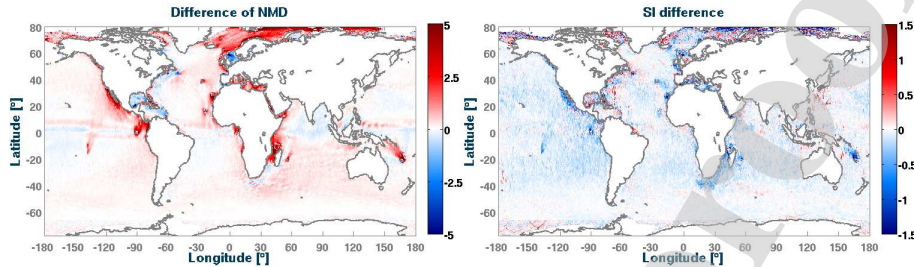


Figure 15: NMD and SI variations in percentage points for the year 2011: values for Multi-grid minus values for Single grid setup, both using the same T475 parameters. Left panel: Difference in NMD values, in this case red values represent a reduction of the negative NMD.

486 positive bias around 140E 10S, downwave of the Tuamotus, or around the
 487 Galapagos, Azores etc. The presence of the full Arctic ocean thanks to the
 488 Arctic grid also adds wave energy that was otherwise missing in the near-
 489 global grid that stopped at 83°N.

490 Overall, the scatter index is reduced over most of the ocean with the
 491 strongest reduction in regions of strong currents like the Agulhas current, or
 492 along complex coastlines such as the Baja California peninsula (blue regions
 493 in Fig. 15.b).

494 4.2. Spectral grid and resolution

495 However, to converge to the true solution of the wave action equation,
 496 increasing only the spatial resolution is not enough, and a finer spectral res-
 497 olution is also needed, in particular for parameters sensitive to numerical
 498 diffusion like the directional spread (Ardhuin and Herbers, 2005). Although
 499 we know that current effects on wave heights would be better resolved with 48
 500 directions instead of only 24 (Ardhuin et al., 2017b; Marechal and Ardhuin,
 501 2020), we have stuck to 24 directions only because of the much lower CPU
 502 cost, and because differences in wave heights when using 24 or 36 directions
 503 were fairly limited. Fig. 16.b shows a change in the Normalized Mean Differ-
 504 ence that is mostly limited to the tropical regions, especially around coasts
 505 and islands for which the finer directional resolution must have an impact
 506 on swell propagation, but the change in scatter index is typically much less
 507 than 1 percentage point (Fig. 16.d).

508 Compared to the costly increase of directional resolution, we found a
 509 higher benefit in terms of H_s accuracy in increasing the spectral range with

510 a maximum frequency of 0.95 Hz instead of the 0.72 Hz used by Rascle and
 511 Arduin (2013). This higher frequency gives a better response, in particular
 512 for the short fetch and low wind conditions in which the peak of the wind
 513 sea would otherwise not be well resolved.

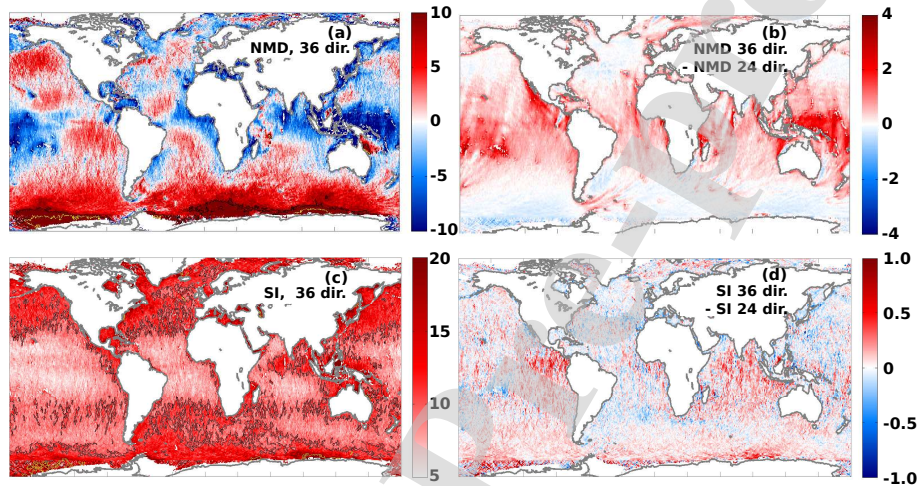


Figure 16: (a)NMD for 1 year averaged H_s using T475 with 36 directions and (b) differences in NMD for T475 with 36 directions with respect to 24 directions (Fig. 10a). Black lines mark the positive 10 % contours. (c) SI for 1 year averaged H_s using T475 with 36 directions and (d) SI difference for T475 with 36 directions with respect to 24 directions. Analyzed year: 2011. Black and yellow lines mark the positive 10 and 20 % contours respectively

514 5. Wave directionality and alternative dissipation parameteriza- 515 tions

516 As noted by Stopa et al. (2016b), the directional spread (Kuik et al.,
 517 1988) is the least well predicted parameter among the most common metrics
 518 used to define the shape of the wave spectrum. Whereas the mean direction
 519 is well controlled by the wind evolution and the time scale of adjustment
 520 of the wave field, the directional spread is probably influenced by details
 521 of the wave generation and dissipation parameterizations. Here we use 3-
 522 hour averaged data from WMO buoy 46436 in the North East Pacific as an
 523 example (see table 4 and Fig. 17), which is the station 166 of the Coastal

524 Data Information Program and is maintained by Thomson et al. (2013). The
 525 correlation coefficient for $\sigma_\theta(f)$ falls below 0.7 for frequency above 0.3 Hz.
 526 Indeed, the model has no skill in predicting $\sigma_\theta(f)$ for $f > 0.5$ Hz, and the
 527 shape of the modeled spectral tail is given by the shape at frequency f_m with
 528 an energy level decreasing like $(f_m/f)^5$, where f_m is a dynamically adjusted
 529 maximum prognostic frequency, set to 2.5 times the mean frequency of the
 530 wind sea part of the spectrum.

531 We note that the directional spread at low frequencies is, close to coasts,
 532 very sensitive to shoreline reflections (Ardhuin and Roland, 2012). Whereas
 533 this has a limited impact on most wave parameters, it is a critical contribution
 534 to microseism and microbarom sources (Stutzmann et al., 2012; De Carlo
 535 et al., 2021). In the present hindcast we have not used the slope-based
 536 reflection coefficient proposed by Ardhuin and Roland (2012) because of the
 537 difficulty of defining the proper slope and mixed results when validating
 538 modeled microseisms. Instead, we have used constant reflexion coefficients of
 539 5%, 10% and 20% for the resolved shorelines, subgrid shorelines and icebergs,
 540 respectively. Clearly that parameterization will have to be tested and further
 541 improved upon using buoy directional spreads together with microseism and
 542 microbarom data.

543 The T475 parameterization is thus still fairly poor for the frequency range
 544 0.4 to 1 Hz when the waves are developed (when the wind sea peak frequency
 545 is below 0.15 Hz), in particular for the directional distribution (Fig. 17.d),
 546 which is critical for the ratio of crosswind to downwind mean square slope
 547 (Munk, 2009), wave breaking statistics (Romero et al., 2017) and the sources
 548 of microseisms and microbaroms at seismic or acoustic frequencies above
 549 0.8 Hz (Farrell and Munk, 2010; Peureux and Ardhuin, 2016; De Carlo et al.,
 550 2020). Recent work have suggested that the shape of the dissipation func-
 551 tion could be better described by Romero (2019), giving the T700 set of
 552 parameters in the WAVEWATCH III model, available in versions 7.0 and
 553 above. In T700, the ad hoc and not very effective cumulative term of Ard-
 554 huin et al. (2010) is replaced with a cumulative term that could be explained
 555 by the straining of short waves caused by long waves (Peureux et al., 2020).
 556 Preliminary tests reveal an interesting behavior for the shape of the high fre-
 557 quency spectrum (Fig. 18), which allows to remove the imposed diagnostic
 558 tail for $f > f_m$ thanks to a completely local (in the spectral sense) parame-
 559 terization of the breaking probability, and the added cosine-squared angular
 560 dependence in the parameterization of the cumulative effect. Possibly this
 561 imposed shape of the cumulative term will have to be revised, as for example

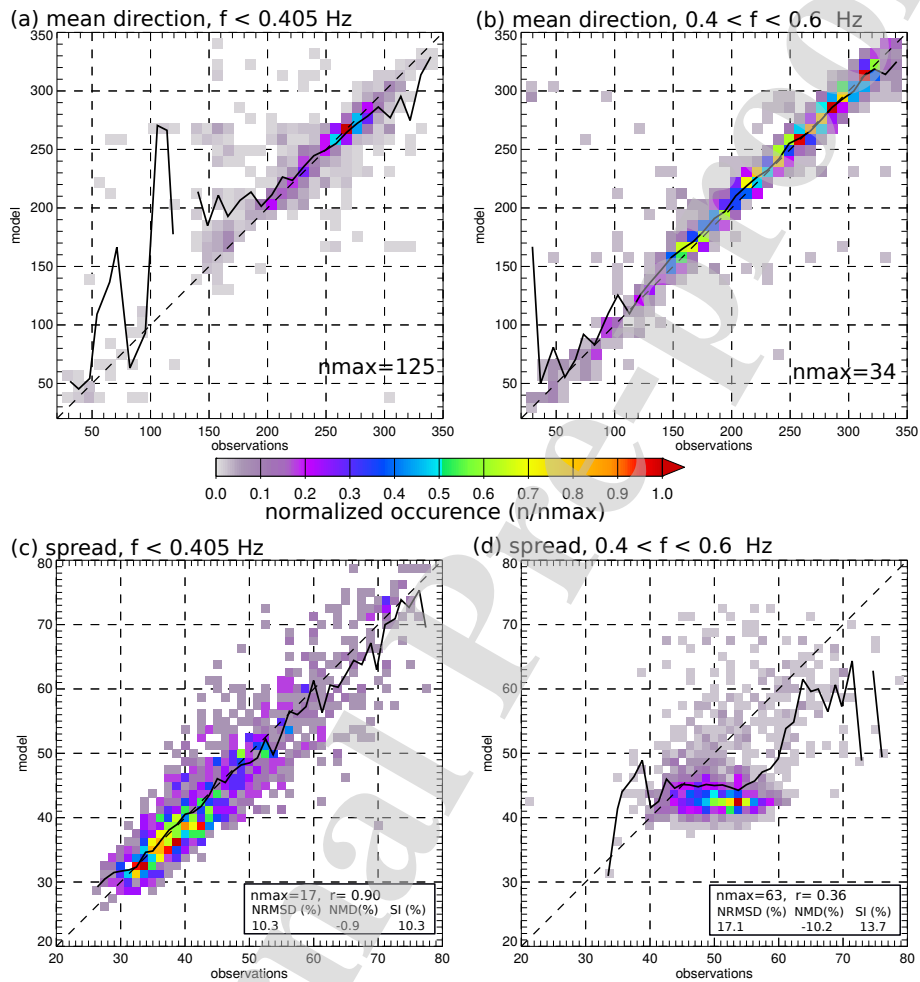


Figure 17: Modeled spread and mean direction for low frequencies ($f < 0.4$ Hz) and high frequencies ($f > 0.4$ Hz) at buoy 46246 for the year 2018. Colors show the number of 3 hour records for which the model-buoy pair falls in one bin, as normalized by the maximum value n_{max} . The solid lines gives the mean modeled value for each observation bin.

562 an isotropic spectrum of long waves should produce an isotropic effect unless
 563 it is a joint effect of the long and short waves. However, Romero (2019) has
 564 produced the first parameterization that is able to produce larger cross-wind

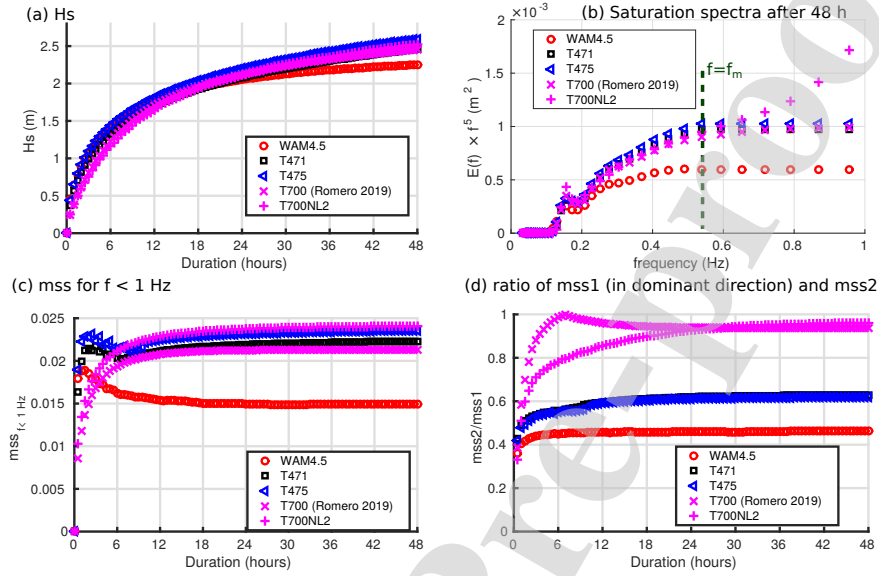


Figure 18: Differences in model results for an academic case considering a uniform ocean and a constant wind speed of 10 m/s starting from no waves. The WAM4.5 parameterization is close to the one used in the ERA5 reanalysis, and the T700NL2 corresponds to the parameterization of Romero (2019) with the non-linear interactions computed with the exact Webb-Resio-Tracy method van Vledder (2006).

565 slopes than down-wind slopes for wavelengths around 1 m (after 7 hours in
 566 Fig. 18.d, the dominant direction for mss1 in T700NL2 is indeed the cross-
 567 wind direction), which are critical to explain the first of the inconvenient sea
 568 truths highlighted by Munk (2009).

569 Taken "out of the box" without the present retuning, the Romero (2019)
 570 parameterization performs similarly to T471 in terms of scatter index but
 571 has a 2 to 6% higher value of wave height (Fig. 19) that will also require an
 572 adjustment of the swell dissipation. The benefits of such a parameterization
 573 will probably be most important for the model parameters that are most
 574 sensitive to the high frequencies, including the mean square slope, and will
 575 require an important upgrade of the wave model in the way these shorter
 576 wave components are treated, so that the wave model result can be validated
 577 with radar back-scatter data (e.g. Nouguier et al., 2016). This effort is beyond
 578 the scope of the present paper and will be discussed in Part 4.

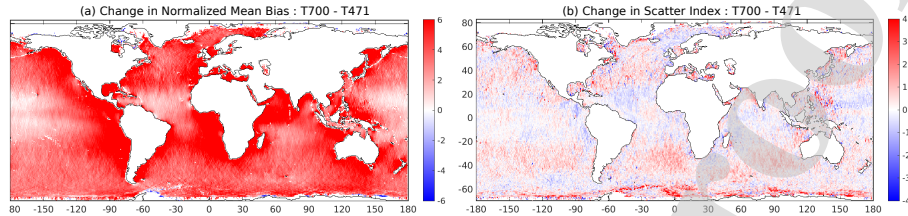


Figure 19: Change in NB and SI from the T471 to T700 change in parameterization for the year 2018. These simulations did not include ocean currents.

579 6. Validation

580 6.1. Validation with altimeter data

581 An important concern about numerical wave model hindcasts for all ap-
 582 plications is their consistency in time which can be compromised by the
 583 time-evolving error statistics of the forcing fields (winds, currents, sea ice)
 584 and/or of the assimilated data which may both introduce time varying biases
 585 and jumps, possibly requiring the statistical adjustment of the forcing fields
 586 (e.g. [Stopa et al., 2019](#)) or the correction of the model results. It is thus nec-
 587 essary to verify the consistency of the model output over time. This requires
 588 validation data that are stable in time. Here we use the satellite altimeter
 589 H_s measurements of [Dodet et al. \(2020\)](#) that were especially designed for
 590 this purpose, and we look at the evolution of the NMB and SI over the years
 591 1997 to 2018 (Fig. 20). We find a general agreement over the years, with
 592 lower variations of the mean difference than was found by [Rascle and Ard-
 593 huin \(2013\)](#) when using CFSR winds, and which had to be corrected in later
 594 hindcasts ([Stopa et al., 2019](#)). Still, the changes from -1 to 2% for the bulk
 595 of the data ($1.5 < H_s < 4$ m) suggest a systematic drift in either the ERA5
 596 wind speeds or the altimeter data, with relatively flatter biases as a function
 597 of H_s for the years 2011-2018 (but still a decrease in the mean model val-
 598 ues or an increase in the altimeter values), and steeper H_s -dependent biases
 599 for the years 1997-2010. The scatter index shows a general reduction of the
 600 random differences that can be caused by a reduction in the random noise
 601 of satellite altimeter data for the more recent missions and an improvement
 602 in the quality of the ERA5 wind fields thanks to the assimilation of a richer
 603 set of data ([Hersbach et al., 2020](#)).

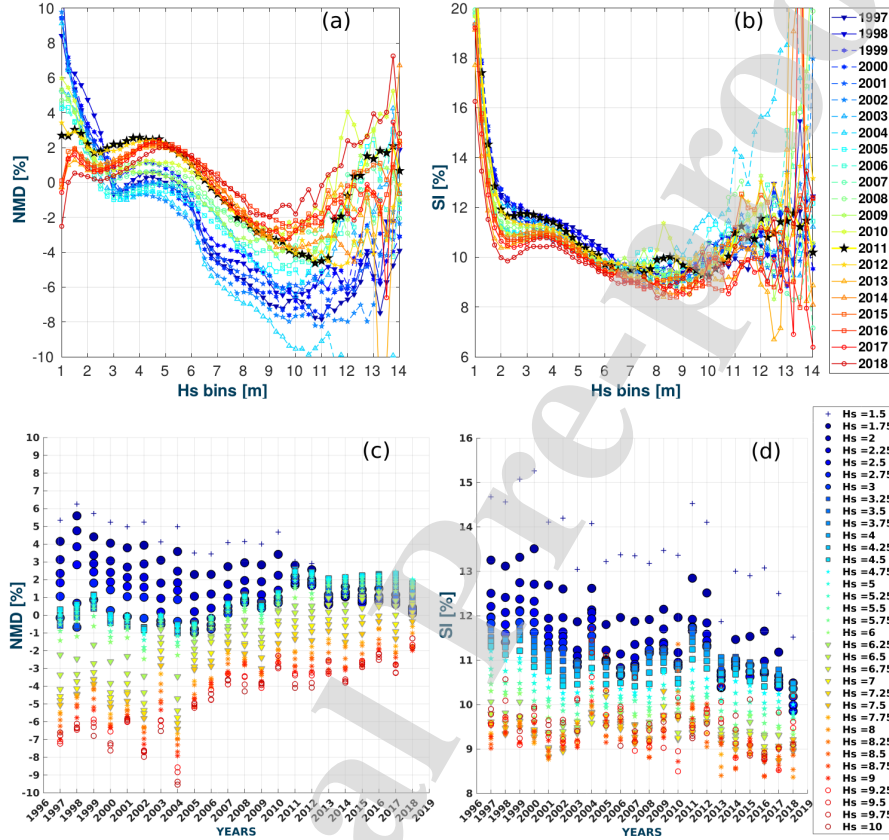


Figure 20: Performance parameters for 22 years hindcast using T475. (a) H_s NMD curves and (b) SI curves, the reference year (2011) used for model tuning has been highlighted with a black star. (c) and (d) are the NMD and SI time series of 1.5 to 10 m H_s bins. Bin size is 0.25 m. Altimeters used for validation: Topex (1997-2002), Envisat (2003-2010), Jason-2 (2011-2012), Saral (2013-2018).

604 6.2. Comparison to ERA5 wave heights

605 Because the ERA5 reanalysis also included a wave model it is questionable
 606 that our efforts have any added value, especially because the ERA5 wave
 607 model assimilates altimeter wave heights and uses a wind forcing at the 10
 608 minutes time step of the atmospheric circulation model to which it is coupled.
 609 However, we know (J.R. Bidlot, personal communication) that the same

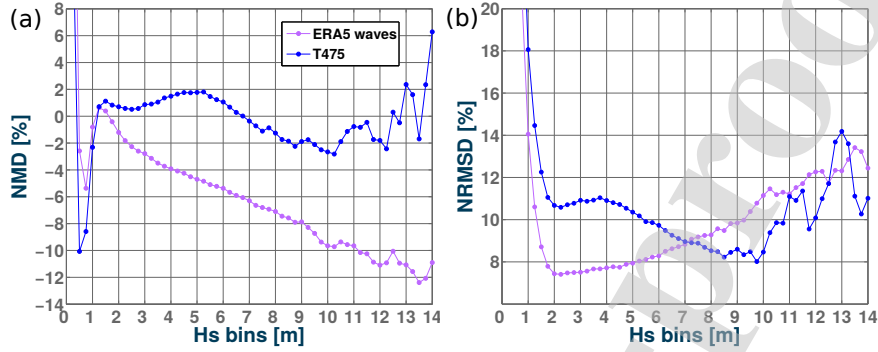


Figure 21: Performance parameters curves for test T475 and ERA5 wave product with respect to Jason-3 altimeter data. (a) H_s NMD, and (b) NRMSD. Analyzed year: 2018. H_s bin size is 0.25 m.

610 ECMWF wave model that uses improved wave generation and dissipation
 611 parameterization in the IFS cycle 46R1 that is operational as of June 6, 2019
 612 (ECMWF, 2019) and is similar to T471, already gives better results than the
 613 ERA5 wave heights at buoy locations. It is thus interesting to look at the
 614 differences between the ERA5 wave heights and the results of the present
 615 hindcast. We note that our model uses different forcing, in particular for
 616 currents, sea ice and icebergs, includes some shoreline and iceberg reflexion
 617 and produces different output parameters, including fluxes of energy between
 618 the ocean and atmosphere, in addition to the parameters that can be derived
 619 from the wave spectrum. Here we only compare the two simulations using
 620 the Jason-3 data for 2018, which has not been assimilated in ERA5.

621 Fig. 21 shows a very strong negative bias in the ERA5 wave heights that,
 622 combined with a much lower random errors, gives larger rms differences for
 623 $H_s > 7$ m. Looking at the spatial distribution of these errors we typically find
 624 larger random errors in the Southern ocean with T475 compared to ERA5
 625 wave heights (Fig. 22), possibly a benefit of the assimilation of the other
 626 satellite missions where the satellite tracks are most dense, and we find lower
 627 random error in a few specific areas with T475, including in the Agulhas
 628 current, which shows again the benefit of properly including ocean surface
 629 currents in a wave model.

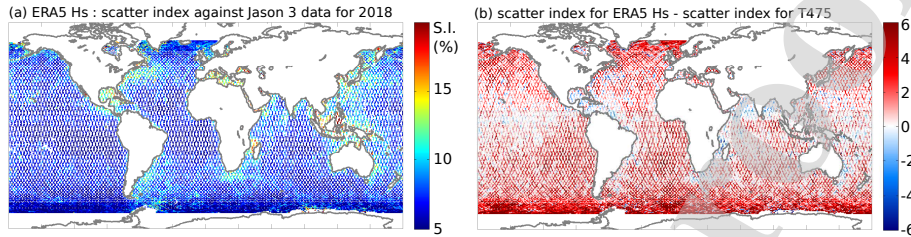


Figure 22: (a) Scatter Index for 1 year (2018) averaged ERA-5 H_s with respect to Jason-3 altimeter data. (b) Difference in scatter index between T475 and ERA-5 waves product.

630 6.3. Validation with buoy data

631 So far all of our analysis, except for a brief discussion of mean direction
 632 and directional spread, has been based on wave heights alone, whereas
 633 our model hindcast is based on the simulation of ocean wave spectra and
 634 produces a wide range of spatially gridded parameters as well as spectra at
 635 selected locations: around 10,000 points all along the world coastline plus the
 636 locations of moored buoys and a few additional offshore points. Even though
 637 the model was only marginally changed compared to the version validated
 638 by [Stopa et al. \(2016a\)](#), it is interesting to look at errors on the shape of
 639 spectra and wave period and directions parameters.

640 These comparisons are not simple because of the large response differences
 641 of different buoy types for wavelenghts shorter than 10 m ($f \simeq 0.4$ Hz)
 642 in particular 3 m diameter discus buoys tend to filter frequencies above 0.4 Hz
 643 which are well reproduced, up to 0.6 Hz by 0.8 m diameter Waverider buoys
 644 (e.g. [Ardhuin et al., 2019](#)). We thus focus on the 0.05 to 0.4 Hz frequency
 645 band. Another difficulty is that most Waverider buoys are located in coastal
 646 areas. We have particularly selected 5 buoys that are representative of different
 647 wave climates, as listed in Table 4. The buoy heave spectra were averaged
 648 over 3 h intervals.

649 Fig. 23 shows different validations of the spectral content of the wave
 650 spectrum. Away from the coasts, at station Papa (buoy 46246), the average
 651 wave spectra in Fig. 23.a reveal a general good behavior of the model compared
 652 to Datawell buoy measurements with mean differences under 10% in the
 653 frequency range 0.05 to 0.4 Hz. The deviation at low frequencies can be
 654 due to the presence of infragravity waves in the buoy measurements which
 655 were not included in our model simulation, but could have been added and
 656 have a typical height of 1 cm in the open ocean ([Ardhuin et al., 2014](#)). That

WMO code	latitude	longitude	depth	shore distance	buoy type
46246	50.0N	145.2 W	4252 m	900 km	Datawell WR
51208	22.285 N	159.574 W	200 m	5 km	Datawell WR
51004	17.53 N	152.25 W	5183 m	300 km	3-m discus
42097	25.7 N	83.65 W	81 m	130 km	Datawell WR
44098	42.8 N	70.17 W	77 m	37 km	Datawell WR

Table 4: List of buoys selected for detailed validation over the years 2018 and 2019. Note that data was missing before July 6, 2019 for buoy 46246.

657 deviation could also be the result of mooring line effects. At high frequen-
658 cies, the model underestimation for $f > 0.5$ Hz may be due to the buoy heave
659 resonance (Datawell, 2014).

660 The variability of the energy content at different frequencies is generally
661 well captured by the parameters H_s and mean periods $T_{m0,2}$ (which is more
662 sensitive to the high frequencies) and $T_{m-1,0}$ (more sensitive to the low fre-
663 quencies). With a bias for the mean periods at buoy 46246 under 1% and a
664 scatter index around 5%, the model is particularly accurate for the shape of
665 the wave spectrum.

666 For other buoys, differences between the model and the observations can
667 reveal errors in buoy measurements (e.g. the spectrum roll-off for $f > 0.52$ Hz
668 at 51004 is typical of 3-m discus buoys) and difficulties for the model to re-
669 solve coastal sea state variability and growth for relatively short fetches. In
670 particular, the energy for low frequencies ($f < 0.06$ Hz) is strongly underes-
671 timated in the Gulf of Mexico and the Gulf of Maine.

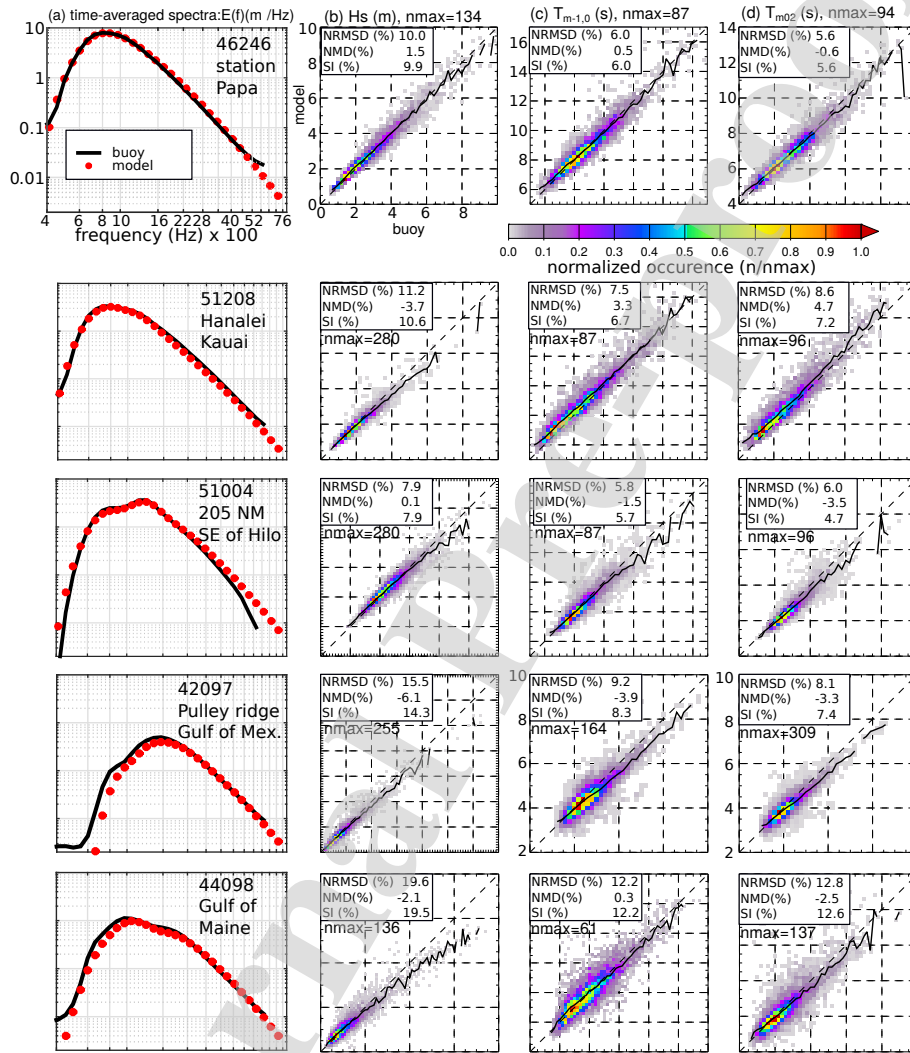


Figure 23: Modeled and measured mean spectra, scatter plots for H_s , and mean periods $T_{m-1,0}$, $T_{m0,2}$ at selected buoys listed in Table 4.

672 7. Conclusions

673 The present paper discusses the influence of forcing fields (winds, sur-
674 face current, sea ice concentration, iceberg concentration), parameterizations
675 (wind-wave generation and swell damping) and resolution (in physical and
676 spectral space) on the wave heights produced by a wave model hindcast, us-
677 ing the WAVEWATCH III modelling framework and satellite-derived wave
678 heights. It is unfortunately not practical to test all the possible combina-
679 tions of model settings, but we expect that the choice of forcing fields and
680 adjustment of parameters is generally robust, and the measurements shows
681 that the present hindcast, in the context of the Integrated Ocean Waves for
682 Geophysical and other Applications (IOWAGA) project, is generally super-
683 ior to the previous version described by [Rascle and Ardhuin \(2013\)](#), and in
684 some regions, for large wave heights, is superior to the ERA5 reanalysis wave
685 product.

686 For the forcing, we found that ERA5 winds, once corrected for a low bias
687 at wind speeds above 21 m/s, gave more accurate results than operational
688 ECMWF analyses or the CFSR reanalysis. Alternative merged satellite-
689 model products ([Bentamy et al., 2018](#)) gave interesting results. We also
690 found that the use of currents provided by CMEMS-Globcurrent generally
691 improved the model results. Probably because these current estimates are
692 missing a significant part of the Total Surface Current Velocity, they degraded
693 the model results at latitudes larger than 50° N. Finally, we confirmed the
694 importance of both sea ice and icebergs for Southern Ocean and Arctic wave
695 heights.

696 For the model parameterizations of air-sea interactions, we have shown
697 that the distribution of H_s around the global maximum of 2 m, could be
698 used to adjust the transition from a laminar to a turbulent boundary layers
699 above the waves, that is very important for the attenuation of swells, and is
700 probably the most sensitive part of the model parameterizations.

701 Regarding model discretizations, we have found a great benefit in includ-
702 ing the 0.7 to 1 Hz frequency range, even though the directionality in that
703 range is not yet well described by the model when waves are developed.

704 For all these tests, we have only performed limited validation for other
705 parameters besides the significant wave height. We expect that future adjust-
706 ments will particularly focus on the high frequencies ($f > 0.4$ Hz) with more
707 validation of the variables that are most sensitive to that frequency range,
708 starting with the mean square slope and its directional components. In this

709 respect, we expect to produce a Part 4 update on the present work based on
710 the parameterizations of Romero (2019) and a much better treatment of the
711 model high frequencies that would make it consistent with remote sensing
712 data, as analyzed by Nouguier et al. (2016) or Yueh et al. (2006), following
713 the work of Elfouhaily et al. (1997).

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724 have helped to improve the quality of this manuscript.

725 Data availability

726 The generated hindcast using test T475, will be available in the following
727 web site: <https://www.umr-lops.fr/Donnees/Vagues/sextant>

728 At the moment of the generation of the present study, the wave data
729 set covers 28 years, from 1993 to 2020 with 3-hourly outputted data. The
730 extension of the hindcast to prior and more recent years is an ongoing work.

731 Appendix A. Detailed model implementation

732 The wave model hindcast and tests presented here all use version 7.0 of
733 WAVEWATCH III. The hindcast uses a list of switches, which appears in all
734 NetCDF file products,

- 735 • physical parameterizations : LN1 ST4 STAB0 NL1 BT4 DB1 MLIM
736 TR0 BS0 IC2 IS2 REF1 RWND WCOR
- 737 • advection and GSE correction: PR3 UQ

738 • other numerical aspects: F90 NOGRB NC4 SCRIP SCRIPNC DIST
 739 MPI FLX0 XX0 WNT2 WNX1 CRT1 CRX1 TIDE TRKNC O0 O1
 740 O2 O2a O2b O2c O3 O4 O5 O6 O7

741 The model parameters are adjusted with the same parameters for all
 742 model grids, and except for default parameter values the T475 parameters
 743 use these adjustments

744 • air-sea interaction parameters (SIN4 namelist) BETAMAX = 1.75,
 745 SWELLF = 0.66, TAUWSHELTER = 0.3, SWELLF3 = 0.022, SWELL-
 746 F4 = 115000.0, SWELLF7 = 432000.00

747 • wave-ice dissipation parameters (SIC2 namelist) IC2DISPER = F, IC2-
 748 TURB = 1.0, IC2ROUGH = 0.001, IC2DMAX = 0.3, IC2REYNOLDS
 749 = 150000, IC2SMOOTH = 200000., IC2VISC = 2.

750 • wave-ice scattering and floe size effects including break-up and in-
 751 elastic dissipation (SIS2 namelist): ISC1 = 0.2, IS2C2 = 0., IS2C3
 752 = 0., IS2BACKSCAT = 1., IS2BREAK = T, IS2DUPDATE = F,
 753 IS2CREEPB = 0.2E8, IS2CREEPD = 0.5, IS2CREEPN = 3.0, IS2-
 754 BREAKF = 3.6, IS2WIM1 = 1.0, IS2FLEXSTR = 2.7414E+05, IS2-
 755 CREEPC = 0.4, IS2ANDISE = 0.55

756 • reflexion parameters (REF1 namelist): REF Coast = 0.05, REF-
 757 COSP_STRAIGHT = 4, REFFREQ = 1., REFICEBERG = 0.2, REF-
 758 MAP = 0., REFSLOPE=0., REFSUBGRID = 0.1, REFRMAX = 0.5

759 • other parameterizations (MISC namelist) ICEHINIT = 1., ICEHMIN
 760 = 0.1, CICE0 = 0.25, CICEN = 2.00, LICE = 40000., FLAGTR = 4,
 761 FACBERG = 0.2, NOSW = 6, WCOR1 = 21., WCOR2 = 1.05 /

762 • activation of 3D output fields (full spectra and seismic sources, OUTS
 763 namelist) P2SF = 1, E3D = 1, I1P2SF = 3, I2P2SF = 24

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- An update on the Rasche and Ardhuin (2013) wave hindcast is presented
- ERA5 wind forcing gives best results but requires a bias correction for high winds
- Satellite altimeter data was used to adjust swell dissipation parameters

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