

Highlights

A satellite altimetry data assimilation approach to optimise sea state estimates from vessel motion

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- Sea state properties, including the directional wave spectrum, are estimated from ship motion response applying the response amplitude operators
- Satellite altimeter observations are assimilated to minimize uncertainties in the estimation of the response amplitude operator.
- Estimated sea state conditions from ship motion during the Antarctic Circumnavigation Expedition compare well against concurrent marine radar measurements

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A satellite altimetry data assimilation approach to optimise sea state estimates from vessel motion

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Abstract

Estimates of directional wave spectra and related parameters can be obtained from ship motion data through the wave-buoy analogy approach. The fundamental input is the response amplitude operator (RAO), which translates ship response into a wave energy spectrum. While ship motion is routinely measured on ocean going vessels, the RAO is not directly available and it is approximated using ship hydrodynamic models. The lack of publicly available details of hull geometry and loading conditions can result in significant inaccuracy of this operator. Considering the reliability of remotely sensed wave height, here we propose an assimilation technique that uses satellite altimeter observations to calibrate the RAO and minimise its uncertainties. The method is applied to estimate sea state conditions during the Antarctic Circumnavigation Expedition by converting motion response of the icebreaker Akademik Tryoshnikov as recorded by the on-board inertial measurement unit. Comparison against concurrent sea state observations

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obtained from a marine radar device shows a good agreement for a variety of parameters including significant wave height, wave periods and mean wave direction.

Keywords: Response amplitude operator, sea-state reconstruction, WaMoS-II, satellite altimeter, wave spectrum

1. Introduction

Vessels across the world have surveyed the sea state through the voluntary observing ships (VOS) scheme for centuries, reporting visual estimates of wave height, period and mean direction for wind sea and swell systems (e.g. Gulev and Grigorieva, 2004; Grigorieva et al., 2017). Nowadays, satellite borne sensors scan the Earth surface twice a day, providing records of significant wave height (i.e. four times the standard deviation of the surface elevation; Holthuijsen, 2010) from altimeter sensors (Ribal and Young, 2019), and directional wave spectra from synthetic aperture radar (SAR; Collard et al., 2009) and surface waves investigation and monitoring (SWIM; Aouf et al., 2021) sensors. Despite the global coverage, a low observation density and contamination of sea ice in the satellite footprints at high latitudes result in significant under-sampling and high uncertainties of measured wave properties in polar regions (Takbash and Young, 2019).

Following the opening of new shipping routes in the Arctic ocean (e.g. Smith and Stephenson, 2013) and a renewed interest in the marginal ice zone for its contribution to the global climate, there has been a growing demand for reliable sea state measurements at high latitudes to enhance performance of Earth system models. To complement satellite data, ice-tethered and

20 drift buoys have been deployed during recent campaigns (e.g. Kohout et al.,
21 2014; Thomson et al., 2018; Vichi et al., 2019; Alberello et al., 2020a), but
22 these devices only endure the harsh polar environment for a short period of
23 time (usually few weeks). Considering the numerous expeditions taking place
24 every year (Schmale et al., 2019), ship based observations, whether through
25 visual or technological means, still remain a key source of wave data at high
26 latitudes (e.g. Derkani et al., 2021; Alberello et al., 2022; Løken et al., 2021).

27 Remote sensing techniques using the vessel’s marine radars are exten-
28 sively employed to measure the directional wave spectrum in the open ocean
29 with good accuracy (see comparison against buoy data in e.g. Hessner et al.,
30 2002; Borge et al., 2004). However, the strong back-scatter from sea ice is
31 a significant source of uncertainty in the marginal ice zone (Derkani et al.,
32 2021). Stereo imaging have recently been applied effectively to detect wave
33 properties from moving vessels (Alberello et al., 2022), but the technique
34 is limited to daylight hours. A promising approach based on bow-mounted
35 altimeter sensors have recently been proposed by Løken et al. (2021), but no
36 directional properties can be extrapolated. An alternative and inexpensive
37 method that uses standard ship sensors consists in converting the vessel six-
38 degree-of-freedom motion response from the on-board inertial measurement
39 units (IMU) into an incident directional wave spectrum through the response
40 amplitude operator (RAO; Newman, 2018), i.e. a transfer function that de-
41 termines the behaviour of the vessel at sea. The method is referred to as the
42 wave-buoy analogy (WBA) and it has been demonstrated to produce reliable
43 measurements of integrated wave parameters such as significant wave height
44 and mean periods (e.g. Nielsen, 2007; Nielsen and Stredulinsky, 2012; Nielsen,

2017; Nielsen and Dietz, 2020). Nevertheless, the reconstruction of the full wave spectrum remains challenging. One of the reasons is the low-pass filtering effect of the vessel itself. This prevents accuracy in the upper tail of the reconstructed wave spectrum and hampers applications related to nonlinear wave dynamics (e.g. estimates of wave height distributions and probability of occurrence of large waves) and wave-structure interaction (Fadaeiazar et al., 2018, 2020; Decorte et al., 2021). Other reasons include the lack of publicly available information of ship geometry and draft, which lead to inaccuracies in the estimates of the RAO.

Assuming that vessel motion response depends linearly on the incident wave properties and considering that navigation and sea state conditions are stationary, the analogy builds on a comparison between the measured response and its calculated counterpart in the spectral domain, according to the following equation:

$$R_j(\omega_e) = \int_0^{2\pi} |Z_j(\omega_e, \vartheta)|^2 E_e(\omega_e, \vartheta) d\vartheta, \quad (1)$$

where $R_j(\omega_e)$ is the measured response spectrum for the j -th degree of freedom, $Z_j(\omega_e, \vartheta)$ is the vessel's RAO associated to the j -th degree of freedom and $E_e(\omega_e, \vartheta)$ is the encounter, unknown, directional wave spectrum (Faltinsen, 1993); ω_e is the encounter angular frequency and ϑ is the wave direction. Note that $E_e(\omega_e, \vartheta)$ is the spectral density detected by a moving body and, hence, it is a distortion of the true spectrum due to a Doppler shift (Faltinsen, 1993). A Bayesian approach is used to estimate the response spectrum $Z_j(\omega_e, \vartheta)$ in order to obtain the best convergence between the measured (left-hand side of the equation (1)) and the calculated (right-hand side) response spectrum, which includes a parametric technique based on analytical spec-

69 tral shapes (e.g. Nielsen, 2007). The RAO is an intrinsic feature of the vessel
 70 and, if unknown, it can be estimated by solving the equation of motion of
 71 a solid body (Newman, 2018) with different methodologies, which are im-
 72 plemented in commercial and open source software (Gourlay et al., 2015).
 73 Calculations require comprehensive knowledge of the hull geometry, loading
 74 conditions (which define the draft), forward speed and heading. While the
 75 latter two parameters are routinely recorded and made available with posi-
 76 tion and meteorological data, ship’s drawings are not publicly accessible and
 77 actual loading conditions are unknown, constituting a significant source of er-
 78 ror. Machine learning techniques based on e.g. convolutional neural networks
 79 (Scholcz and Mak, 2020; Duz et al., 2019) attempt to eliminate uncertainties
 80 due to the RAO by directly predicting the incident wave spectra. However,
 81 these methods require extensive records of concurrent ship response and sea
 82 state to train the algorithms that are not commonly available.

83 Here we propose a data assimilation approach (a technique used to en-
 84 hance performance of numerical weather predictions, e.g. Aouf et al., 2021),
 85 where satellite observations in the open ocean are combined with sea state
 86 estimations from vessel motion. The aim is to optimise the RAO through an
 87 iterative process that minimises the error in the approximation of significant
 88 wave height. The method is applicable when key data for the estimate of
 89 the RAO are unavailable. Herein we use observations of ship motion (with
 90 minimal information of ship characteristics) that were acquired during the
 91 Antarctic Circumnavigation Expedition (ACE, December 2016 – March 2017;
 92 Schmale et al., 2019; Derkani et al., 2021; Landwehr et al., 2021) and freely
 93 accessible satellite altimetry data (Ribal and Young, 2019, 2020). Recon-

94 structured sea state conditions (wave spectra and integrated parameters) are
 95 validated against concomitant records of directional wave spectra from the
 96 marine radar, which are assumed to be the benchmark (ground truth) for
 97 this study. In the next section, field and satellite data are presented. The
 98 sea state estimation approach is introduced in Section 3 and its validation is
 99 discussed in Section 4. Concluding remarks are reported in the last section.

100 **2. Field data**

101 *2.1. Records of ship motion and sea state*

102 Concurrent observations of ship motion and sea state were acquired dur-
 103 ing ACE, which consisted in a voyage around the Southern Ocean between
 104 34° and 74° South aboard the research vessel *Akademik Tryoshnikov* from
 105 20 December 2016 to 19 March 2017 (Schmale et al., 2019; Derkani et al.,
 106 2021). A schematic of the expedition is presented in Fig. 1. Significant wave
 107 height along the route as measured by ship borne instruments and remotely
 108 sensed averages of sea ice extent from the satellite borne advanced microwave
 109 scanning radiometer 2 (AMSR2) sensor are also reported in the figure.

110 Vessel accelerations and rotational motion were acquired continuously
 111 with the ship’s inertial measurement unit (IMU) at a sampling rate of 1Hz.
 112 Data were automatically processed on board to obtain a vertical transla-
 113 tional motion (heave), a rotation about the longitudinal axis (pitch) and a
 114 rotation about the transverse axis (roll). Only these three degrees of freedom
 115 were made available. Note that other approaches using the wave-buoy anal-
 116 ogy are also limited to heave, pitch and roll (e.g. Nielsen and Jensen, 2011;
 117 Stredulinsky and Thornhill, 2011) and so are other techniques to extrapo-

late directional wave spectrum from floating devices such as buoys (Young, 1994). The database was complemented with forward speed, ship heading and GPS positions (details to freely access the database can be found in Alberello et al., 2020b). Basic information on ship geometry (beam and length, 23 m and 133.6 m respectively) and class are available on the internet. However, details of hull type (i.e. drawings) and loading conditions (draft) are not publicly available, noting that this is a common situation when accessing data from past expeditions.

Simultaneous sea state measurements were recorded with the wave and surface current monitoring system (WaMoS-II), which uses the ship’s marine X-band radar to acquire high-definition radar images of the surrounding ocean surface and derive the directional wave energy spectrum at a sampling rate of 175 s (see Alberello et al., 2020c, to access the data). A running average over a 20-minute window was applied to minimise the effects of natural variability of waves within the sampling interval (World Meteorological Organization, 1998). For each spectrum, integrated parameters such as significant wave height, mean wave periods, spectral bandwidth and directional spreading, among many others, were computed automatically (see Derkani et al., 2021). It is worth noting that WaMoS-II has an internal quality control protocol that automatically discards unreliable data (Hessner et al., 2019). Most of the observations in heavy rainfall did not pass this quality control (cf. Chen et al., 2020; Huang et al., 2017) and, hence, were excluded from any further analysis. Data recorded in the marginal ice zone were also excluded due to uncertainties related to strong radar back-scatter from sea ice (see Derkani et al., 2021). Overall, 8990 data points were available for this study.

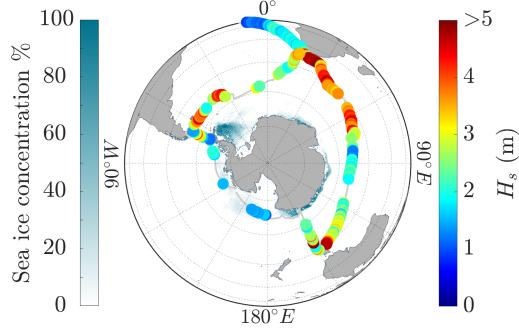


Figure 1: Significant wave height (H_s) recorded underway and average sea ice concentration from satellite sensors during the Antarctic Circumnavigation Expedition (ACE).

143 The significant wave height was, on average, ≈ 2.6 m, which is consistent
 144 with the 50th percentile of wave height that is expected during the austral
 145 summer. The maximum wave heights were recorded in excess of 5 m during
 146 most of the storms Derkani et al. (2021) and exceeded the 90th percentile
 147 expected in the austral summer. Wave periods were generally long, with
 148 values over 8 s (wavelengths > 100 m).

149 2.2. Satellite altimetry data

150 Remotely sensed observations of significant wave height from altimeter
 151 sensors (H_s^{sat}) mounted on Jason-2, Jason-3, Sentinel-3A, Sentinel-3B, Saral
 152 and Cryosat-2 satellites were extracted from the Australian Ocean Data Net-
 153 work (an overview of the database and access links can be found in Ribal
 154 and Young, 2019). Records were quality controlled, calibrated and validated
 155 against available buoy data (see Ribal and Young, 2019, 2020) and only ob-
 156 servations flagged as accurate were used herein. To avoid interference of sea
 157 ice, observations at latitudes higher than 60° South are neglected. Similarly,
 158 observations too close to the coastline (i.e. less than 50 km from the shore)

159 were omitted to avoid contamination of land in the satellite footprint.

160 3. Sea state estimation from vessel motion

161 3.1. Method

162 The governing equation (1) can be reformulated to model the encounter
163 wave energy spectrum as

$$S_e(\omega_e) = \left\langle \frac{R_j(\omega_e)}{|\zeta_j(\omega_e)|^2} \right\rangle, \quad (2)$$

164 where $\langle \cdot \rangle$ denotes the average over the degrees of freedom (heave, pitch and
165 roll),

$$\zeta_j(\omega_e) = \frac{1}{2\pi} \int_0^{2\pi} Z_j(\omega_e, \vartheta) d\vartheta \quad (3)$$

166 is the RAO in the frequency domain and

$$S_e(\omega_e) = \int_0^{2\pi} E_e(\omega_e, \vartheta) d\vartheta \quad (4)$$

167 is the encounter frequency wave spectrum. Although a weighted average
168 approach based on the actual encountered wave direction is desirable (e.g.
169 Nielsen and Dietz, 2020), this requires knowledge of the wave spectrum, and
170 specifically the wave direction, *a priori*. The only source for wave direction
171 along the ship route, in this respect, is model hindcast, which would add fur-
172 ther uncertainties to the process (see biases of model hincast in the Southern
173 Ocean in Derkani, 2021). Therefore, standard averaging is applied. The sig-
174 nificant wave height (H_s) can be calculated from the zero-th order moment
175 (m_0) of the wave spectrum as

$$H_s = 4 \sqrt{m_0} = 4 \sqrt{\int_0^\infty S_e(\omega_e) d\omega_e}. \quad (5)$$

176 Note that the Doppler shift distorts the spectral shape but it affects neither
 177 the energy content nor the zero-th order moment.

178 The incident wave spectrum is the unknown function in the governing
 179 equation. However, if the geometry of the vessel and its draft are not acces-
 180 sible, also the RAO remains unknown. Whereas the hull geometry can be
 181 approximated using publicly available drawings of similar vessels, the draft
 182 is not routinely recorded. As the latter determines the submerged portion of
 183 the hull and, consequently, affects the mass and moments of inertia (New-
 184 man, 2018), its absence represents a significant constraint for the estimation
 185 of the RAO. Herein we use a trial-and-error approach to identify the draft
 186 by minimising errors in the evaluation of the significant wave height, relative
 187 to satellite observations. The strategy is to (i) estimate the RAO based on
 188 guessed parameters of hull geometry and draft, (ii) model the unknown wave
 189 spectrum with (2) and H_s with (5), (iii) compare the latter against a mea-
 190 sured counterpart (H_s^{sat}), and (iv) calibrate the RAO by adjusting the draft
 191 until $|H_s - H_s^{sat}|$ is minimised (see §3.2.1). We remark that this iterative
 192 process is limited to open ocean, where satellite data is reliable. The RAO
 193 corresponding to the minimal error is selected and applied to equation (2) to
 194 evaluate the frequency wave spectrum $S_e(\omega_e)$ anywhere along the ship route.

195 This approach allows the direct estimation of the encounter spectral den-
 196 sity function in the frequency domain. Therefore, a directional spreading
 197 function, $D_e(\omega_e, \vartheta)$, has to be applied to obtain the directional spectrum:

$$E_e(\omega_e, \vartheta) = D_e(\omega_e, \vartheta) S_e(\omega_e). \quad (6)$$

198 The directional function, which has to satisfy the condition

$$\int_0^{2\pi} D_e(\omega_e, \vartheta) d\vartheta = 1, \quad (7)$$

199 is often expressed as a Fourier series

$$D_e(\omega_e, \vartheta) = \frac{1}{\pi} \left\{ \frac{1}{2} + \sum_{n=1}^{\infty} [a_n(\omega_e) \cos(n\vartheta) + b_n(\omega_e) \sin(n\vartheta)] \right\} \quad (8)$$

200 where a_n and b_n are the Fourier coefficients (see Young, 1994; Kuik et al.,
 201 1988; Ewans, 1998; Pettersson et al., 2003, for details). At each frequency, a_n
 202 and b_n are estimated from a combination of ship's elevation (heave), surface
 203 slope in the longitudinal direction (the pitch angle) and surface slope in the
 204 transversal direction (the roll angle) using the Maximum Entropy Method
 205 (MEM, Lygre and Krogstad, 1986). It should be noted that MEM, similarly
 206 to other techniques such as the maximum likelihood method, tends to over
 207 predict the directional spreading, especially at high frequencies (e.g. Young,
 208 1994)

209 The encounter spectra (and frequencies) need to be corrected to eliminate
 210 the Doppler shift and thus restore the original spectral shape of the sea state.
 211 The true frequency ω (i.e. the frequency that would be measured by a fixed
 212 device) relates to ω_e through the following expression (Faltinsen, 1993):

$$\omega_e = \omega - \frac{\omega^2}{g} V \cos(\vartheta) \quad (9)$$

213 where V is the forward speed and ϑ is the wave direction. The true wave
 214 spectrum can be derived by performing the change of variable:

$$E(\omega, \vartheta) = E_e(\omega_e, \vartheta) \frac{d\omega_e}{d\omega} \quad (10)$$

215 where $d\omega_e/d\omega$ is the Jacobian of the transformation (Holthuijsen, 2010).

216 3.2. Application to field data

217 3.2.1. Response amplitude operator

218 Starting from ship geometry, $Z_j(\omega_e, \vartheta)$ can be computed using different
219 software packages such as WAMIT, NEMOH and Hydrostar (a model inter-
220 comparison is reported in Gourlay et al., 2015; Parisella and Gourlay, 2016;
221 Nelli et al., 2021). Herein we use NEMOH, a model based on linear free
222 surface potential flow theory (Babarit and Delhommeau, 2015) which has
223 the advantage of being open source. In normal operating conditions (small
224 wave steepness and broad directional spreading), nonlinearity is weak (e.g.
225 Onorato et al., 2009) and the linear approximation is sufficient to model re-
226 sponse motion, with results being consistent with second order solutions (see
227 Nelli et al., 2021). It is worth noting that, while the motion of heave and
228 pitch can be described reasonably well by the potential flow theory, this is
229 not the case for the rolling motion, which is complicated by viscous effects
230 (Chakrabarti, 2001). Therefore, roll was estimated including additional com-
231 ponents to account for frictional damping, eddy damping and lift (Himeno,
232 1981; Newman, 2018; Quality Systems Group of the 28th ITTC, 2011).

233 The input condition for the model is the geometry of the submerged part
234 of the vessel. Dimensions (length and beam) are available and specified in
235 §2. The hull type is unknown and, hence, it is approximated by a similar
236 Arc7 ice-class design, which is then fitted to the Akademik Tryoshnikov’s
237 dimensions. The approach of fitting a general hull to known dimensions is
238 shown in the literature to produce minimal uncertainties (see, e.g. Gourlay
239 et al., 2015) and thus it does not require further adjustment. The loading
240 conditions, which determine the draft, are also unspecified. Although the

241 Akademik Tryoshnikov's standard summer draft can be found on the internet,
 242 there is no assurance that this was the actual draft during the expedition.
 243 Uncertainties in the draft are more significant than those associated to hull
 244 type. Fig. 2, in this respect, reports the sensitivity of the RAO to different
 245 loading conditions, with drafts ranging from 7 m (under-loaded condition) to
 246 9.5 m (over-loaded condition). Variations are evident at high frequencies for
 247 any degree of freedom. These are only slightly noticeable for heave, but are
 248 moderate to prominent for pitch and roll.

249 The forward speed of the vessel affects added mass and damping (Faltin-
 250 sen, 1993) and it needs to be considered in the computation of the RAO.
 251 As the focus is the open ocean, we make the bold assumption that the ship
 252 is at cruising speed, which, for the Akademik Tryoshnikov, is 16 knots (or
 253 8.3 m/s). Although a real-time speed adjustment is preferable, the cruise
 254 speed was preferred due to gaps in the time series of GPS positions and
 255 forward speed.

256 3.2.2. *Response spectra*

257 In order to satisfy the hypothesis of stationarity, records of heave, pitch
 258 and roll motion are subdivided in segments of 20 minutes. The response spec-
 259 tra for each degree of freedom, $R_j(\omega_e)$, are calculated as ensemble averages
 260 of the Fourier transform of 256 s intervals of the 20-minute time series with
 261 50% overlap. As the vessel filters high frequency modes, components with
 262 encounter frequencies $> 1.5 \text{ rad/s}$ (i.e. 0.24 Hz) are excluded to minimise
 263 noise.

264 3.2.3. Calibration with satellite data

265 The response spectra, $R_j(\omega_e)$, and RAO, $Z_j(\omega_e, \vartheta)$, are applied to esti-
 266 mate the significant wave height from motion response (H_s^{IMU}), equations (2)
 267 and (5), which is then compared against H_s^{sat} . Response motion and satellite
 268 data, however, do not have a uniform spatial and temporal distribution: the
 269 sea state estimations are available along the ship route every 20 minutes,
 270 while satellite observations are scattered in space and time. To allow com-
 271 parison, records are consolidated into clusters centred on ship positions and
 272 extending in space over a radius of a quarter of a degree and in time over a
 273 window of ± 1.5 hours. All response spectra and satellite observations within
 274 a cluster are averaged to obtain consistent values.

275 The difference between H_s^{IMU} and H_s^{sat} is measured over the entire voy-
 276 age through the root mean square error ($RMSE$). The variation of $RMSE$
 277 as a function of the draft is shown in Fig. 3a. Error is ≈ 0.62 m for an initial
 278 draft of 7 m and weakly decreases with the increase of the draft, reaching a
 279 minimum of 0.56 m for a draft of 8.5 m (coincidentally, this is the standard
 280 summer draft of the vessel). It is worth noting that the minimum $RMSE$ is
 281 consistent with the one associated to global wave hindcasts (Janssen et al.,
 282 2005). A similar result can be observed in Fig. 3b, where the corresponding
 283 values of the correlation coefficient (r) and scatter index (SI) are plotted.
 284 Both values reach optimal conditions for the draft value of 8.5 m (i.e. maxi-
 285 mum r and minimum SI), while any further increase of the draft produces
 286 a notable degeneration of error metrics, corroborating that even small over
 287 estimations of the draft can result in notable errors. The scatter of modelled
 288 versus observed significant wave height for this optimal draft is reported in

289 Fig. 4. The agreement is good with scatter index $SI \approx 0.33$ and correlation
 290 coefficient $r = 0.88$

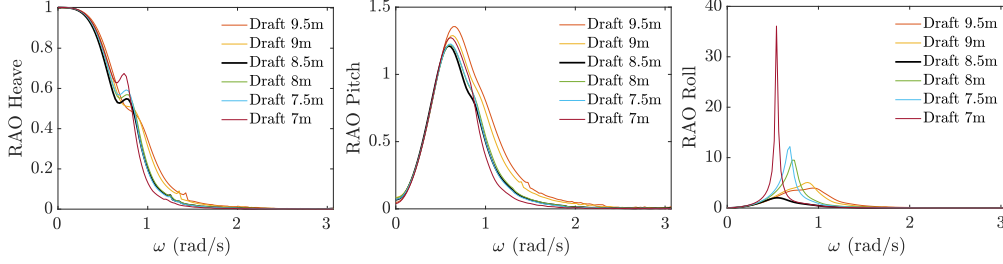


Figure 2: Response amplitude operator (RAO) for heave, pitch and roll calculated for variable drafts; the thick black line indicates the optimal draft.

291 4. Validation

292 Sea state estimations derived with the optimal RAO are compared against
 293 collocated measurements (close in space and time) from the marine radar
 294 (WaMoS-II) to validate the method. Although acquisition rates are different,
 295 we recall that the radar smooths records with a 20-minute running average
 296 window. Thereby, collocated data points are consistent with the sea state
 297 estimations, which also represent an average over a 20-minute window. It
 298 is worth noting that a comparison with in-situ devices such as wave buoys
 299 would be preferable. However, only a few buoys are operated in the Southern
 300 Ocean (Young et al., 2020; Derkani et al., 2021) and none of them were
 301 nearby the ship route. Marine radar provides high quality observations that
 302 are comparable with buoys (Borge et al., 2004) and has the advantage to
 303 record data concurrently with the ship motion device, making it an ideal
 304 benchmark for validation.

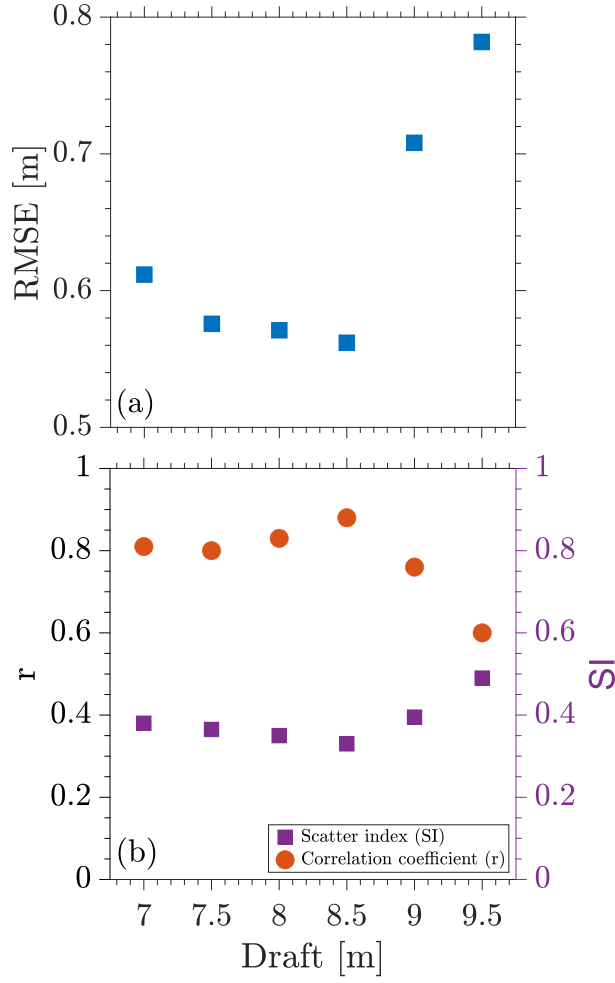


Figure 3: Panel (a): Root Mean Square Error (RMSE) showing differences between H_s estimated from motion response and satellite observations as a function of draft. Panel (b): Scatter Index (SI) and Correlation Coefficient (r) relative to data in panel (a)

305 4.1. Frequency spectrum and integrated parameters

306 Figure 5 shows six examples of collocated frequency spectra estimated
 307 from the vessel motion (IMU) and the marine radar (WaMoS-II). Overall,
 308 the spectral density function is reproduced well (e.g. Fig. 5a,c,f), also in the

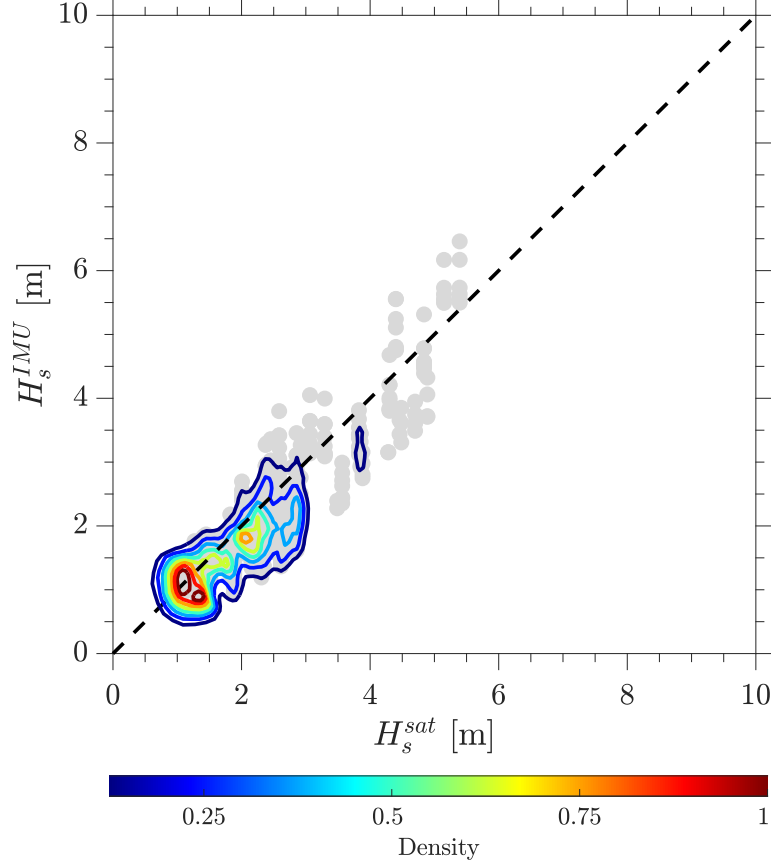


Figure 4: Estimated significant wave height from motion response (H_s^{IMU}) versus satellite observations (H_s^{sat}).

309 presence of multiple systems (mixed seas with peaks at 0.6 and 1.0 rad/s;
 310 Fig. 5e). There is, nonetheless, evidence of uncertainties in the frequency
 311 bandwidth, which can results in distortions of the spectral tails (see Fig. 5d).
 312 These uncertainties are partially attributed to the reconstruction method.
 313 However, they are further exacerbated by the fact that the vessel acts as
 314 a filter for high frequency components and the low sampling rate (1 Hz),

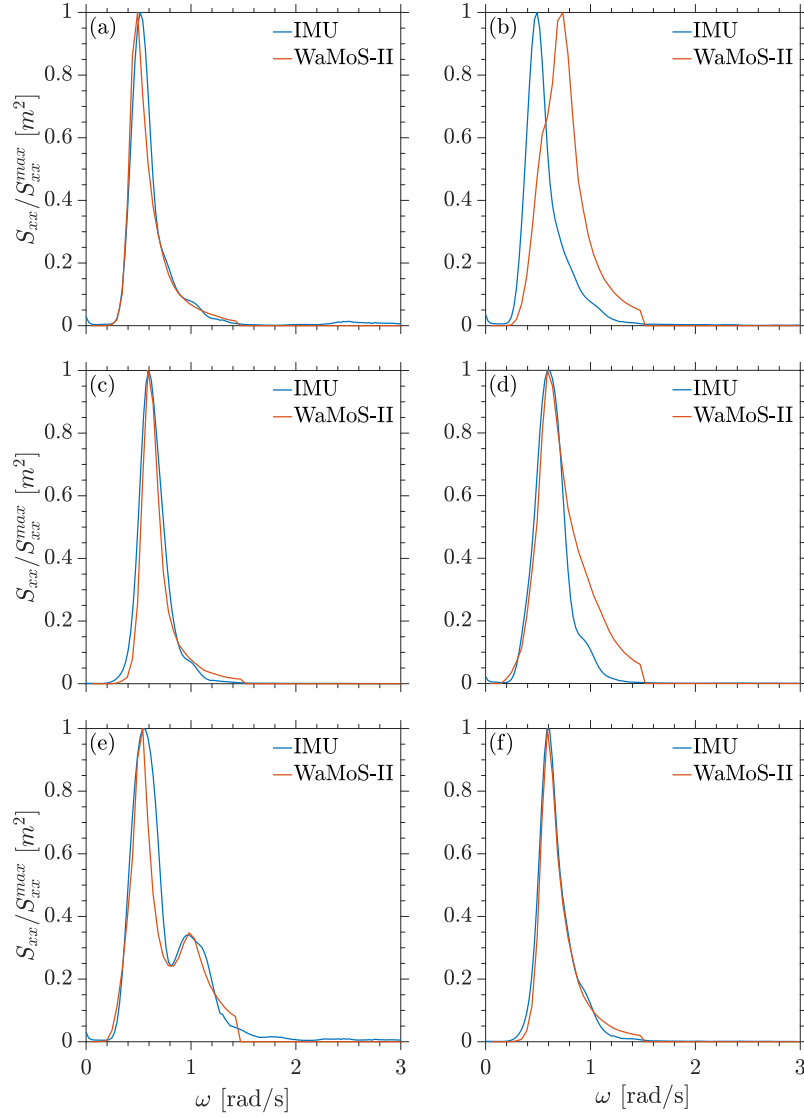


Figure 5: Example of frequency wave spectra S_{xx} estimated from ship motion response (IMU) and concurrent observations from the marine radar (WaMoS-II). Spectra are normalised by their peak S_{xx}^{max}

315 which affect resolution of high frequencies modes. In some instances, the
 316 spectrum is misrepresented, with estimates from ship motion being down- or

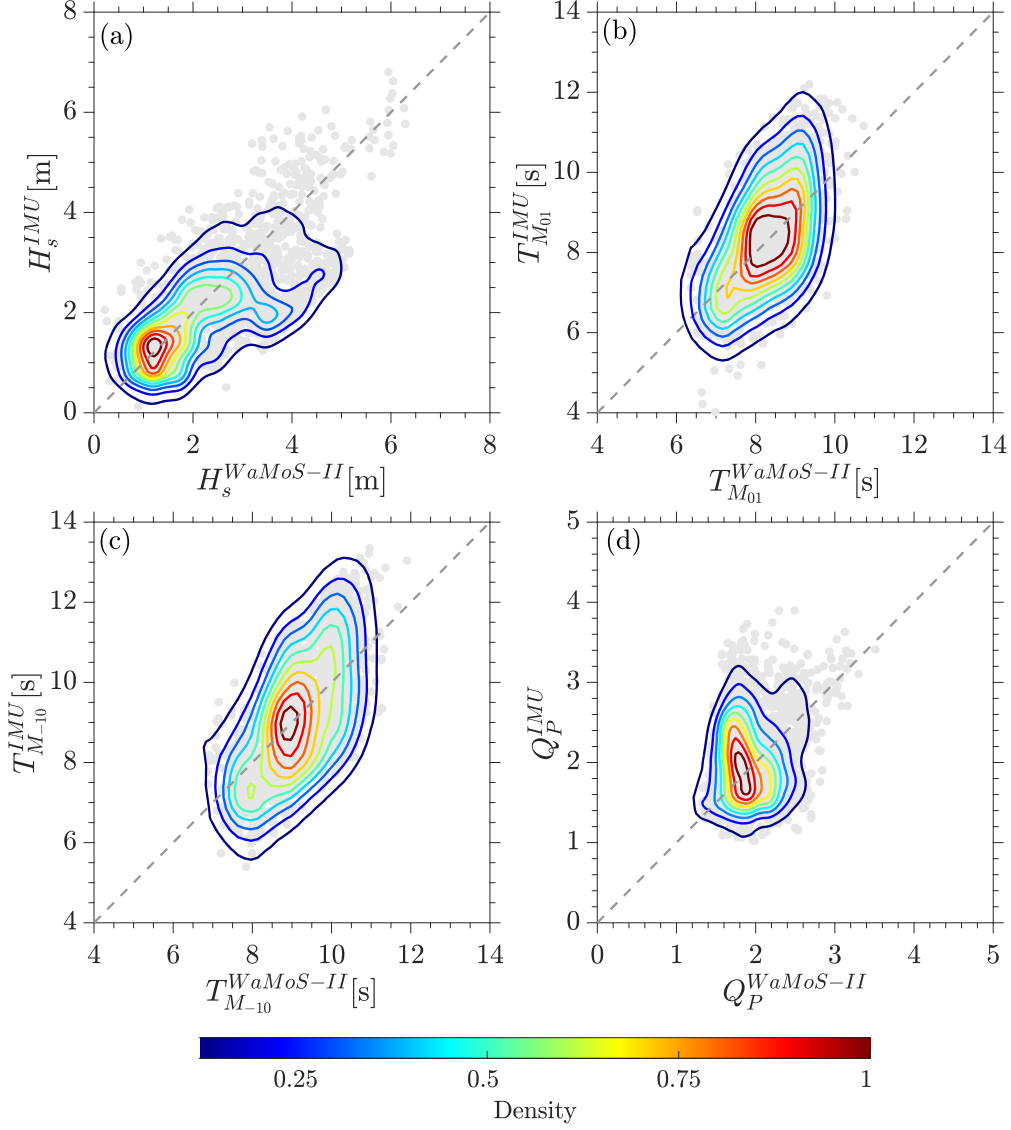


Figure 6: Integrated parameters estimated from motion response (IMU) versus observations based on the marine radar (WaMoS-II): (a) significant wave height H_s , (b) mean period $T_{M_{01}}$, (c) mean period $T_{M_{-10}}$ and (d) peakedness parameter Q_P .

up-shifted relative to those from the marine radar (an example of down-shift is reported in Fig. 5b). These large errors are attributed to uncertainties in the assumption of stationary forward speed and/or ship heading within the 20-minute time window.

Integrated parameters are used to assess accuracy across the entire database (Fig. 6). Estimations of significant wave height, i.e. the overall energy content of the wave spectrum, compare well with measurements from the marine radar (Fig. 6a) with $RMSE \approx 0.6$ m, $SI \approx 0.3$ and $r \approx 0.8$. There is, however, the tendency to slightly underestimate benchmark observations (mean bias error $MBE = 0.7$ m), which is an issue also reported for Bayesian and parametric approaches (see Nielsen, 2017). The mean periods $T_{M_{01}} = m_1/m_0$ and $T_{M_{-10}} = m_{-1}/m_0$, where m_i are the i -th order moments of the wave spectrum, weight the contribution of high (the former) and low (the latter) frequency components to the overall spectral density, respectively. The comparison with marine radar is reasonable, but its metrics is less accurate than the one for significant wave height: $RMSE \approx 1$ s, $SI \approx 0.1$, and $r \approx 0.6$ for $T_{M_{01}}$ (Fig. 6b); $RMSE \approx 1.3$ s, $SI \approx 0.15$, and $r \approx 0.6$ for $T_{M_{-10}}$ (Fig. 6c). Both periods shows a weak over prediction bias ($MBE = 1$ s and 1.5 s for $T_{M_{01}}$ and $T_{M_{-10}}$, respectively). The frequency bandwidth is evaluated via the peakedness parameter

$$Q_P = \frac{2}{m_0^2} \int_0^\infty \omega S(\omega)^2 d\omega, \quad (11)$$

which is a more stable approximation than other bandwidth parameters available in the literature (Goda, 2010). Error metrics $RMSE \approx 0.45$, $SI \approx 0.22$, $r \approx 0.3$ and $MBE = 0.3$ indicate inconsistencies with benchmark observations and confirm extent of uncertainties in the upper tail of the spectral

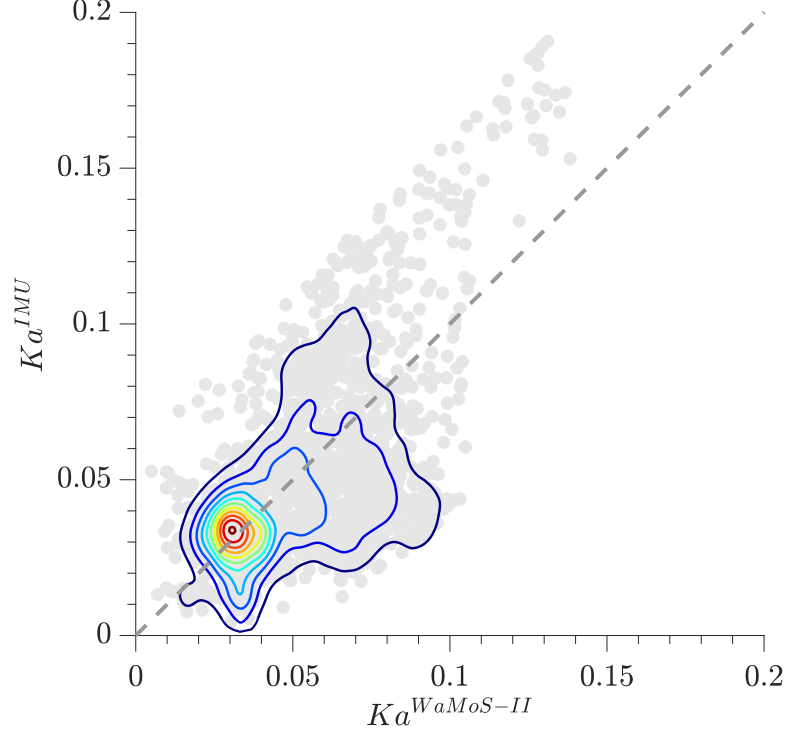


Figure 7: Wave steepness $Ka = K_m H_s / 2$ estimated from motion response (IMU) versus observations based on the marine radar (WaMoS-II)

341 density function across the database. Nonetheless, the scatter (Fig. 6d) re-
 342 mains centred on the 1:1 correlation line and data points are within a $\pm 50\%$
 343 error. This is a notable achievement for the proposed method, though, con-
 344 sidering that the spectrum has been reconstructed from (very) limited initial
 345 conditions.

346 The wave steepness defined as $Ka = K_m H_s / 2$, where $K_m = 2\pi / (1.56 T_{M01}^2)$,
 347 is presented in Fig. 7. Correlation is reasonable with error metrics $RMSE \approx$
 348 0.027 , $SI \approx 0.4$, $r \approx 0.6$ and $MBE = 0.011$ (SI and r are comparable to
 349 those of significant wave height). Nevertheless, there is a notable bias for

the ship motion to return higher wave steepness than the marine radar for $Ka > 0.12$. This is an indication that the linear approximation used to derive the RAO is not thorough when the nonlinearity of the sea state is not negligible (cf. Nelli et al., 2017). Note, however, that $Ka > 0.12$ is limited to relative small wave heights during early stages of wave growth.

4.2. Directional properties

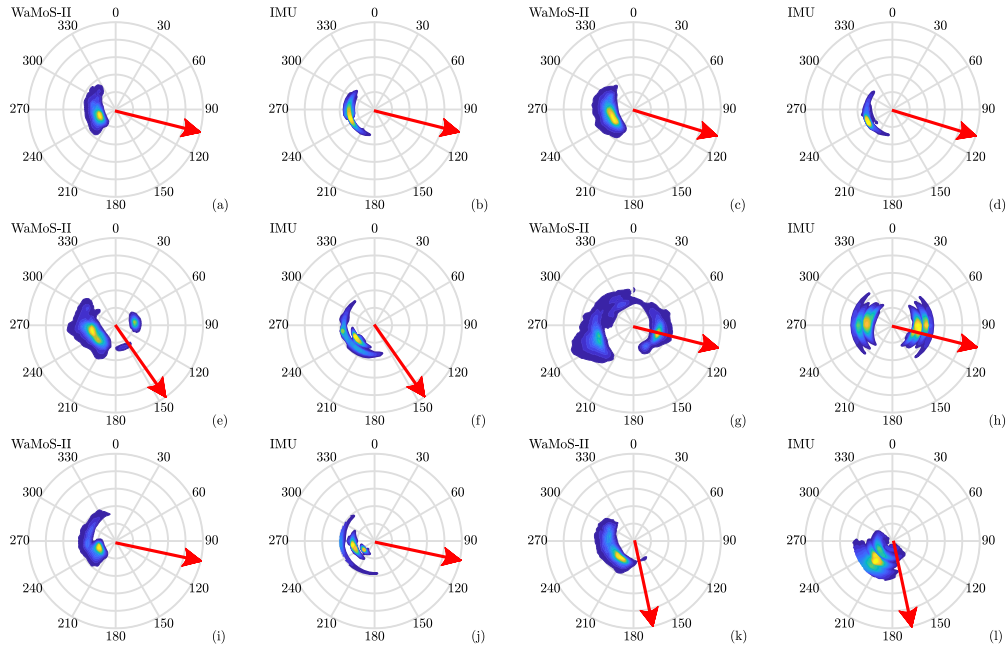


Figure 8: Examples of directional wave spectra as reconstructed from the motion response (IMU) and the marine radar (WaMoS-II); the arrow indicates the ship heading.

Examples of directional wave spectra as derived from the vessel motion and the marine radar are reported in Fig. 8. The comparison shows that the overall feature of the directional distribution is captured reasonably well, even though reconstructed spectra are generally narrower, both in frequency

and direction, than the marine radar counterpart (e.g. panels c-d, e-f and i-j). Despite some noise, which results in spurious peaks, multiple wave systems are generally detected (e.g. Fig. 8g-h). In this respect, it is worth noting that multi modality is a distinctive feature of Southern Ocean waves as reported by SAR/SWIM spectra (which also compare well with marine radar observations shown herein) and three-dimensional imaging of the ocean surface (e.g. Derkani et al., 2021; Aouf et al., 2021; Alberello et al., 2022).

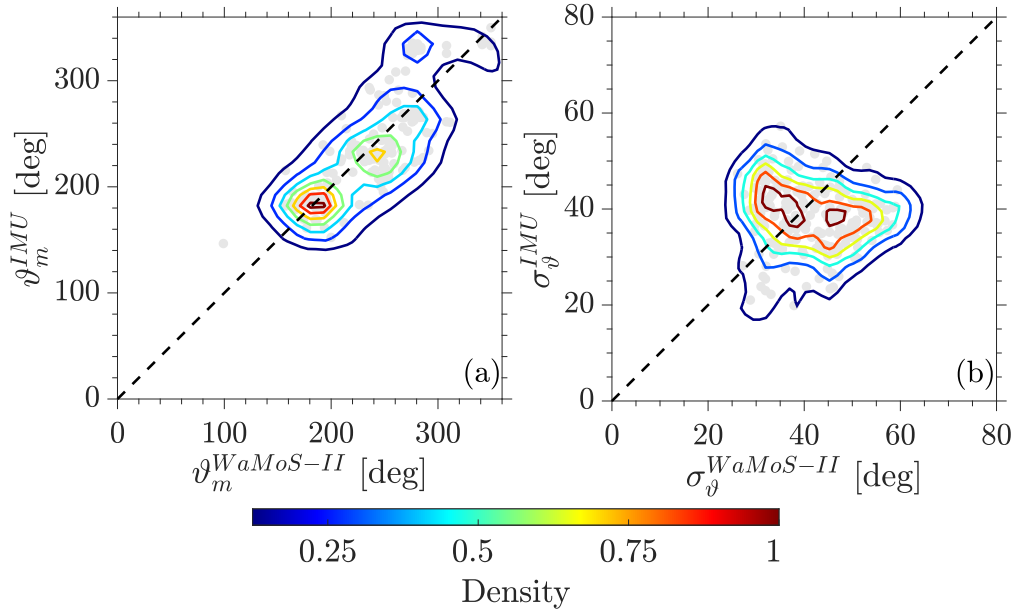


Figure 9: Directional properties estimated from motion response (IMU) versus the marine radar (WaMoS-II): (a) mean wave direction; and (b) wave directional spreading.

In order to validate the directional properties more comprehensively, estimations of the mean wave direction

$$\vartheta_m = \arctan\left(\frac{b}{a}\right) \quad (12)$$

369 and the directional spreading

$$\sigma_{\vartheta} = \frac{180}{\pi} \sqrt{2 \left[1 - \sqrt{\frac{a^2 + b^2}{m_0^2}} \right]}, \quad (13)$$

370 where

$$a = \int_{\omega=0}^{\infty} \int_{\vartheta=0}^{2\pi} \cos(\vartheta) E(\omega, \vartheta) d\omega d\vartheta \quad (14)$$

371 and

$$b = \int_{\omega=0}^{\infty} \int_{\vartheta=0}^{2\pi} \sin(\vartheta) E(\omega, \vartheta) d\omega d\vartheta, \quad (15)$$

are compared against those obtained from the marine radar (see Fig. 9).

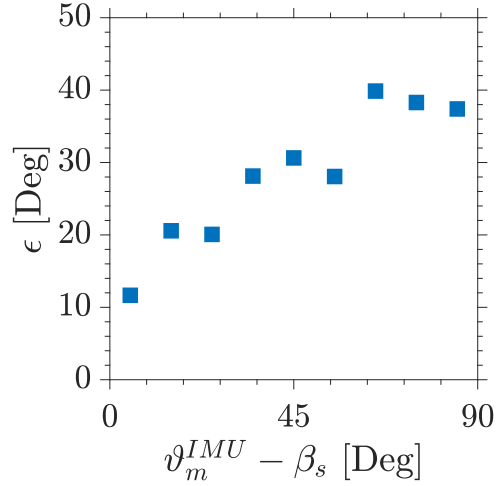


Figure 10: Mean wave direction error (ϵ) versus the difference between the incident mean wave direction (ϑ_m^{IMU}) and ship heading (β_s). Following/head seas correspond to $|\vartheta_m^{IMU} - \beta_s| = 0^\circ$; beam seas corresponds to $|\vartheta_m^{IMU} - \beta_s| = 90^\circ$.

372

373 Estimates from ship motion replicate ϑ_m well: $RMSE \approx 28^\circ$, $SI \approx$
 374 0.1 and $r \approx 0.85$. The mean wave direction error is minimum ($\approx 11^\circ$) for
 375 following/head seas, which occur when waves propagate along/against the

376 ship heading (i.e. $|\vartheta_m^{IMU} - \beta_s| \approx 0^\circ$, where β_s is the ship heading). For
 377 increasing $|\vartheta_m^{IMU} - \beta_s|$, the error rises monotonically, reaching a maximum
 378 of $\approx 38^\circ$ in beam seas for which $|\vartheta_m^{IMU} - \beta_s| \approx 90^\circ$ (see Fig. 10). This
 379 suggests the error is dominated by uncertainties in the rolling motion, which
 380 exacerbates in conditions of beam seas.

381 The agreement for σ_ϑ is less clear. Although estimates from ship motion
 382 are consistent with records from the marine radar ($RMSE \approx 15^\circ$), there is
 383 a notable scatter ($SI \approx 0.5$ and $r \approx 0.08$) with estimated spectra showing
 384 a directional spreading up to 20 degrees broader/narrower than the marine
 385 radar counterpart. Such uncertainties in the estimates of the directional
 386 spreading are not surprising though, as the parameters is one of the most
 387 challenging to measure (cf. Young, 1994; Ewans, 1998). The RMSE reported
 388 herein, in this respect, is consistent with measurement errors of $\approx 10^\circ - 20^\circ$,
 389 which are reported in the literature (for example, Kuik et al., 1988; Rogers
 390 and Wang, 2007; Young et al., 1996).

391 5. Conclusions

392 Techniques for deriving sea state parameters from ship motion are a valu-
 393 able source of data in regions where measurements based on buoys or satellite
 394 sensors are challenging (e.g. polar regions). Conversion of vessel motion into
 395 sea state can be performed through a ship specific response amplitude op-
 396 erator, which acts as a transfer function. If unknown, the operator can be
 397 estimated numerically with commercial or open source software, using ship
 398 geometry, such as dimensions, hull type and draft, as input. Detailed infor-
 399 mation on geometry, however, are often unavailable (particularly drawings

400 and draft), forcing to rely on guessed ship variables and, hence, contributing
401 to uncertainties in the estimated response amplitude operator and, conse-
402 quently, wave properties. To enhance precision (in the case ship geometry
403 and draft are unavailable), the assimilation of satellite observations into the
404 conversion process has been proposed to calibrate the operator through an
405 iterative method. This minimises the error in the estimate of significant
406 wave height relative to observations from freely available altimeter sensors in
407 the open ocean. Once calibrated, the response amplitude operator allows the
408 direct reconstruction of the incident frequency spectrum from a measured re-
409 sponse spectrum, while directional properties are derived by combining three
410 components of motion (e.g. heave, pitch and roll) through the maximum en-
411 tropy method. The approach is applied herein to estimate directional wave
412 spectra during the Antarctic Circumnavigation Expedition, where measure-
413 ments of motion response from an inertial measurement unit were acquired
414 over a period of three months. Concurrent measurements of the directional
415 wave spectra from the on-board marine radar are used as benchmark for
416 validation.

417 Overall, the main features of the directional wave spectrum are captured,
418 including presence of multiple peaks in mixed seas. Comparisons of inte-
419 grated parameters such as significant wave height and mean period show a
420 satisfactory agreement relative to the marine radar. Nevertheless, there is a
421 slight under estimation bias for wave height and an over prediction bias for
422 the mean periods, which are, to some extent, consistent to biases reported
423 using Bayesian and parametric approaches. The mean wave direction is also
424 well represented. The spectral bandwidth, both in the frequency and direc-

425 tional domain, remains the least accurate parameter, evidencing the occur-
426 rence of some evident distortions in the reconstruction of the spectral shape,
427 especially in its upper (i.e. high frequency) tail. Uncertainties are related to
428 inaccuracy in the estimate of the response amplitude operator of rolling mo-
429 tion, which is more sensitive to nonlinear and viscous effects than the heave
430 and pitch counterparts and the inevitable low-pass filtering effects produced
431 by the large floating body. The low sampling rate for acquisition of ves-
432 sel motion can also contribute to uncertainties in resolving high frequency
433 modes, adding inaccuracy to spectral shape and bandwidth. The method
434 would benefit of further verification through analysis of reconstructed wave
435 conditions from other vessels to fully assess uncertainties.

436 It should be noted that key sea state parameters for metocean applications
437 (e.g. validation of theoretical and prediction models, stochastic prediction of
438 extremes) are the significant wave height, mean periods and mean wave direc-
439 tion, while the spectral bandwidth is a marginal feature. Relying on limited
440 and publicly available information, the proposed method can be applied to
441 past expeditions to the Arctic and the Antarctic to retrieve significantly large
442 data set of basic wave parameters, which can support validation of satellite
443 measurements and development of coupled model in the marginal ice zone.
444 The accuracy of the spectral shape (and thus its bandwidth), nevertheless,
445 remains critical and hampers applications related to nonlinear wave dynamics
446 (e.g. Onorato et al., 2009; Fadaeiazar et al., 2018, 2020) and wave-structure
447 interactions, including high frequency excitation, fatigue and sea ice drift and
448 break up (e.g. Alberello et al., 2021; Decorte et al., 2021; Passerotti et al.,
449 2022). Although the proposed technique can also be intended as an alterna-

450 tive, it can be applied in conjunction with Bayesian or parametric methods
451 to further enhance accuracy of the reconstructed spectral density function
452 for wave-structure interaction applications.

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