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1 **The role of density currents and gravity waves in the offshore propagation of**
2 **convection over Sumatra**

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16 ABSTRACT: The Maritime Continent experiences some of the world's most severe convective
17 rainfall, with an intense diurnal cycle. A key feature is offshore propagation of convection overnight,
18 having peaked over land during the evening. Existing hypotheses suggest this propagation is due to
19 the nocturnal land breeze and environmental wind causing low-level convergence; and/or gravity
20 waves triggering convection as they propagate. We use a convection-permitting configuration of
21 the Met Office Unified Model over Sumatra to test these hypotheses, verifying against observations
22 from the Japanese Years of the Maritime Continent field campaign. In selected case studies there
23 is an organized squall line propagating with the land breeze density current, possibly reinforced by
24 convective cold pools, at $\sim 3 \text{ m s}^{-1}$ to around 150–300 km offshore. Propagation at these speeds is
25 also seen in a composite mean diurnal cycle. The density current is verified by observations, with
26 offshore low-level wind and virtual potential temperature showing a rapid decrease consistent with
27 a density current front, accompanied by rainfall. Gravity waves are identified in the model with a
28 typical phase speed of 16 m s^{-1} . They trigger isolated cells of convection, usually further offshore
29 and with much weaker precipitation than the squall line. Occasionally, the isolated convection
30 may deepen and the rainfall intensify, if the gravity wave interacts with a substantial pre-existing
31 perturbation such as shallow cloud. The localized convection triggered by gravity waves does not
32 generally propagate at the wave's own speed, but this phenomenon may appear as propagation
33 along a wave trajectory in a composite that averages over many days of the diurnal cycle.

34 SIGNIFICANCE STATEMENT: The intense convection experienced by the Maritime Continent
35 causes high-impact weather in the form of heavy precipitation, which can trigger floods and
36 landslides, endangering human life and infrastructure. The geography of the region, with many
37 islands with complex coastlines and orography, means that the spatial and temporal distributions
38 of convection are difficult to predict. This presents challenges for operational forecasters in the
39 region; and introduces biases in weather and climate models, which may propagate globally. A key
40 feature of the convection is its diurnal cycle and associated propagation offshore overnight from the
41 islands. Although this phenomenon has been often investigated, there is no strong consensus in the
42 literature on the mechanism or combination of mechanisms responsible. Improving our knowledge
43 of these mechanisms and how they are represented in a convection-permitting model will assist
44 forecasters to understand how and when the propagation of intense convective storms occurs, and
45 allow model developers to improve biases in numerical weather prediction and climate models.

46 **1. Introduction**

47 The Maritime Continent (Ramage 1968) is the south-east Asian archipelago located in the oceanic
48 warm pool between the equatorial Indian and Pacific Oceans. The equatorial location, warm sea
49 surface temperature (SST) and presence of thousands of islands – where onshore sea breezes, driven
50 by the diurnal land-sea temperature contrast, cause strong low-level moisture flux convergence –
51 gives rise to some of the most intense convective storms on Earth. The associated latent heat
52 release is so large that errors in simulating the spatial and temporal distributions of convection
53 cause considerable errors in models on larger scales. These biases in convection over the Maritime
54 Continent can lead to global biases, through processes such as Rossby wave propagation (e.g., Jin
55 and Hoskins 1995).

56 The greatest form of variability in Maritime Continent convection is the diurnal cycle,
57 with convection tending to peak over islands during the afternoon and evening, while the
58 smaller-amplitude diurnal cycle over the sea has its peak in the early morning. Of specific interest
59 in this study is the offshore propagation of convection overnight (e.g., Wu et al. 2009; Marzuki
60 et al. 2022). This propagation is a crucial aspect of the distribution of convection, with Coppin
61 et al. (2020) finding that 80% of all Maritime Continent rainfall is “coastal” – that is, it falls in
62 a precipitation feature which intersects the coast. Peatman et al. (2021) showed how the nature

63 of offshore propagation depends on large-scale phenomena; here we investigate the causes of the
 64 propagation at a physical process level.

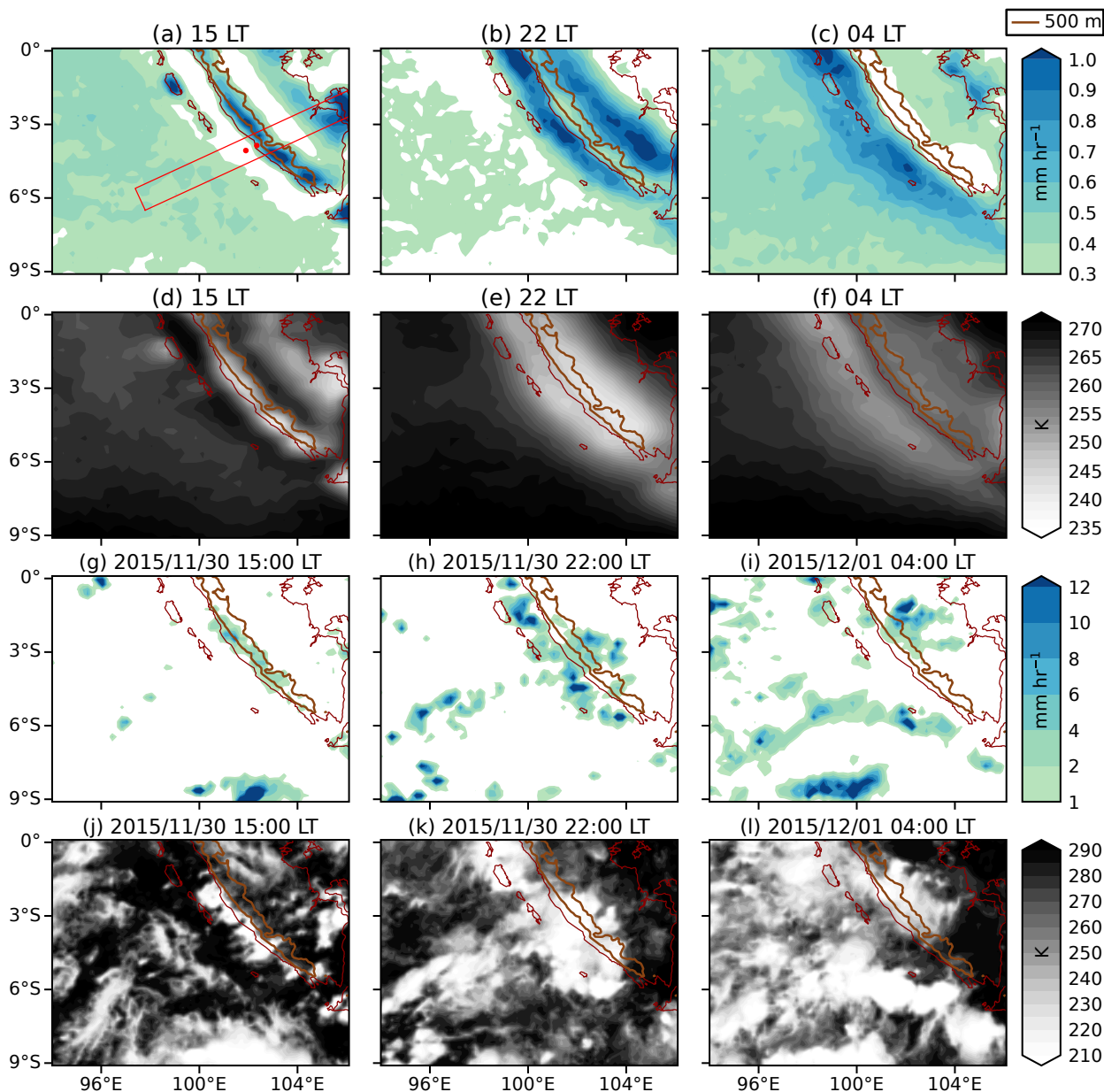


FIG. 1: (a–c) Composite diurnal cycle of observed precipitation rate over south Sumatra, from IMERG for November–April 2000/01–2020/21, showing selected times of day only. (d–f) As (a–c) but for observed 10.4 μm brightness temperature from the Himawari-8 satellite (channel 13) for November–April 2015/16–2019/20. (g–i) Observed precipitation rate for the same times of day as in (a–c) but for the 2015/11/30 case study. (j–l) As (g–i) but for observed 10.4 μm brightness temperature. Data are shown in local time (LT; defined as UTC+7). The Barisan mountains are shown by the 500 m orography contour (brown) from GLOBE. In panel (a) the red dots are the locations of the *R/V Mirai* during the 2015 field campaign and the town of Bengkulu; and the red box is used as the transect in figures 2 and 5.

65 The offshore propagation can be seen south-west of the island of Sumatra in figure 1. A
66 composite mean of the observed diurnal cycle for boreal winter (November to April) is shown
67 for selected times of day in figures 1a–c (precipitation rate) and 1d–f (brightness temperature, a
68 proxy for convection as it shows the location of cloud tops, although brightness temperature picks
69 out non-precipitating cirrus anvils so the relationship with precipitation is approximate). Over
70 Sumatra, convective rainfall tends to form over the Barisan mountains (near the coast, indicated by
71 the brown contour) during the afternoon. By late evening (figures 1b,e), the precipitation (although
72 not the cloud tops at this time) is on average suppressed over the mountain ridge, but has begun
73 to migrate both north-east and south-west. By early morning (figures 1c,f), there is rarely any
74 precipitation over the mountains, with the most intense rainfall and cloud occurring offshore.

75 Of course, a composite as shown here suggests very smooth fields and the reader may be
76 led to imagine a consistent behaviour in the timing and location of the convection day-to-day.
77 In fact, observations of individual cases show spatially noisy rainfall (figures 1g–i) and cloud
78 (figures 1j–l), much less coherent than implied by the composite even though this case has fairly
79 strong propagation, beginning on 2015/11/30 and continuing overnight to 2015/12/01. This case
80 study will be used throughout this paper.

81 Figure 2a shows a Hovmöller diagram of precipitation rate, for all times of day (with eight hours
82 repeated to show the whole of the propagating signal), for the same composite as in figures 1a–c. In
83 this composite the speed of the offshore propagation varies such that the envelope of precipitation
84 widens as it propagates, with the slow edge at around 2 m s^{-1} and the fast edge over 13 m s^{-1} .
85 The most intense rainfall occurs around 50–110 km offshore at 01 local time (LT; defined here as
86 UTC+7), implying a propagation speed of around 3 m s^{-1} from the coast.

87 The range of speeds in the composite may be due to multiple modes of propagation being
88 present on any given day (e.g., Vincent and Lane 2016, 2017; Yokoi et al. 2017; Coppin and
89 Bellon 2019a); or because the modes of propagation themselves can vary in speed from day to
90 day. Another possible cause is the inclusion of days with no propagation in the composite, so other
91 convection offshore may contribute to the averaging without being associated with propagation.
92 However, if we select only those days when the propagation occurs, using the method described
93 in the appendix, the composite precipitation intensifies but the widening of the envelope is still
94 apparent (figure 2b).

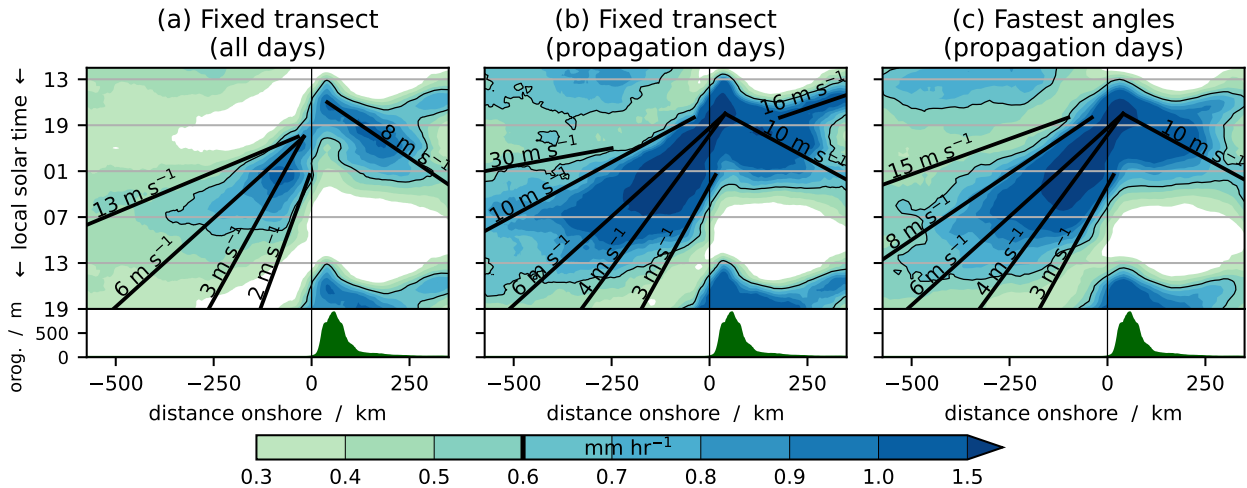


FIG. 2: (a) Onshore/offshore Hovmöller diagram of the composite diurnal cycle of observed precipitation rate also used in figures 1a–c, averaged over the red box in figure 1a. The 0.6 mm hr^{-1} contour is highlighted in black. Orography, averaged over the same box, is shown in green beneath the main plot. (b) As panel (a) but only for days when offshore propagation is diagnosed. (c) As panel (b) but using a variable transect, oriented in the direction of the propagation on each day. More details for (b) and (c) may be found in section 2c and the appendix. Propagation lines at selected phase speeds are overplotted in black.

95 Another possibility is that the direction of propagation varies between days, dependent on the
 96 environmental wind, so the apparent speed as projected onto the fixed box in figure 1a may not be
 97 the true speed. The impact of the varying direction of propagation is demonstrated by figure 2c, in
 98 which the transect for the Hovmöller diagram is not fixed, but is allowed to rotate about 102.05°E ,
 99 3.55°S . For each day with propagation, an angle is chosen to maximize the propagation speed; the
 100 algorithm for selecting this angle is also explained in the appendix. The propagation envelope still
 101 widens with distance from the coast, but it is less pronounced than in figure 2b. This widening
 102 implies there is variability in propagation direction and it does contribute to the shape of the
 103 composite propagation envelope, but even when this variability in direction is taken into account
 104 there is still a range of speeds present. Hence, there is a genuine variability in propagation speeds.

105 Many studies have investigated the physical processes involved in the offshore propagation
 106 over Sumatra and elsewhere, with a variety of mechanisms proposed but no clear consensus.
 107 Houze et al. (1981) found a regular diurnal cycle of rainfall over north Borneo and argued that
 108 the nocturnal offshore part of the cycle consisted of successive cloud clusters triggered due to
 109 low-level convergence between the land breeze from the coast and the boreal winter monsoon
 110 flow. However, studying offshore propagation over the Panama Bight, Mapes et al. (2003b,a)

111 argued against a similar mechanism due to the land breeze being too weak. They attributed the
112 propagation to gravity waves forced by the heat source of the diurnal mixed layer. Many other
113 studies since have also attributed the offshore propagation of convection to either gravity wave
114 processes; or low-level convergence due to the land breeze or other density currents impinging
115 on the environmental winds; or both. These mechanisms are discussed in the remainder of this
116 section.

117 Evidence for propagation due to convergence caused by the land breeze was presented by Hassim
118 et al. (2016) in a modelling study over New Guinea, and by Coppin and Bellon (2019a) in an
119 idealized modelling study of a generic island in the tropics. However, cold pools from the diurnal
120 convection can also cause density currents, and these similarly can lead to low-level convergence
121 and trigger a propagating convection signal (Mori et al. 2004; Wu et al. 2009; Dipankar et al.
122 2019). Moreover, several papers have indicated that the presence of orography, such as the Barisan
123 mountains close to the Sumatra coast, can cause or at least strengthen the boundary layer density
124 currents that converge with the environmental wind, for example due to downslope winds which
125 continue to propagate over the sea (Wu et al. 2009; Qian et al. 2012; Coppin and Bellon 2019b).

126 However, perhaps the most commonly cited mechanisms for the propagation are related to gravity
127 waves (e.g., Love et al. 2011; Hassim et al. 2016; Yokoi et al. 2017; Coppin and Bellon 2019a),
128 either due to direct triggering of convection by the wave or due to the wave destabilizing the
129 atmosphere ahead of the convection. However, in the latter case it is not necessarily clear what
130 controls the speed of the propagation behind the wave; and the gravity wave preconditioning also
131 occurs on days without propagation of convection, so is not alone a sufficient condition (Yokoi
132 et al. 2019).

133 Love et al. (2011) performed empirical orthogonal function analysis of vertical profiles of heating
134 in a model simulation over Sumatra to identify gravity wave modes, tracking the wave propagation
135 using the principal components. In a model simulation with convective parametrization on a 40 km
136 grid, gravity waves were found at speeds of 60 m s^{-1} and 31 m s^{-1} . Because this is considerably faster
137 than the propagation of convection, it was assumed that the waves precondition the atmosphere for
138 the convection through increasing instability. In a simulation with explicit convection on a 4 km
139 grid, the signals were diagnosed as having speeds of 40 m s^{-1} and 3 m s^{-1} , the latter being around

140 the same as the speed of the propagation of convection, but much slower than observed gravity
141 waves.

142 Yokoi et al. (2017) analysed radiosondes from the *R/V Mirai* stationed offshore of Sumatra
143 during a Years of the Maritime Continent (YMC; Yoneyama and Zhang 2020) field campaign
144 (the “pre-YMC” campaign) run by the Japan Agency for Marine-Earth Science and Technology
145 (JAMSTEC) in 2015. In a composite diurnal cycle of vertical profiles, a cooling and moistening
146 of the air just above the boundary layer was found in the evening. This was hypothesized to
147 be due to vertical advection caused by a gravity wave, with analysis showing that variations in
148 potential temperature and humidity at this time and location are approximately consistent with
149 vertical motion. However, from radiosondes alone it is not easy to confirm that gravity waves are
150 responsible for these motions.

151 In an idealized study, Coppin and Bellon (2019a) argued that a gravity wave with a speed of
152 30 m s^{-1} was responsible for offshore propagation of convection although, in a composite diurnal
153 cycle Hovmöller diagram, the correspondence between precipitation and a gravity wave trajectory
154 is difficult to recognize (see their figure 4b).

155 Recently, studies have drawn a distinction between the physical mechanisms close to and far
156 from the coast, consistent with figure 2. Vincent and Lane (2016) noted a slower propagation
157 speed ($3\text{--}5 \text{ m s}^{-1}$) 100–200 km from the coast and a faster speed ($\sim 18 \text{ m s}^{-1}$) further offshore.
158 Bai et al. (2021) argued that the land breeze converging with low-level background westerly winds
159 was responsible for slow (4.5 m s^{-1}) propagation within 180 km of the Sumatran coast, the range
160 of the radar data they analyzed, although the speed of rainfall propagation as observed by the radar
161 was typically slower than the propagation of the low-level convergence in a reanalysis. They also
162 suggested that the role of gravity waves may dominate further from the shore. A survey of the
163 global tropics, in addition to some extra-tropical locations, by Fang and Du (2022) also suggested
164 that propagation observed near to coastlines may depend on density current propagation.

165 Despite the considerable volume of literature on the subject, therefore, a lack of consensus
166 warrants further investigation. Here, we test the hypotheses that the offshore propagation of
167 convection within the diurnal cycle of precipitation over Sumatra is caused by the nocturnal land
168 breeze and/or gravity waves forced by land-based diurnal convection, over a range of distances from
169 the shore. To do this, we perform a modelling investigation of case studies of offshore propagation

170 from Sumatra, verifying the results using observations from the automatic weather station (AWS)
171 on the *R/V Mirai* during the pre-YMC field campaign (Yokoi et al. 2017). The model, and other
172 data and methods, are described in section 2. Results are presented in section 3 and conclusions in
173 section 4.

174 **2. Model, data and methods**

175 *a. Convection-permitting MetUM forecasts*

176 Reforecasts for 26 days of the JAMSTEC pre-YMC field campaign, from 2015/11/22 to
177 2015/12/17, were performed using the Met Office Unified Model (MetUM) version 11.1 at 0.02°
178 (approximately 2.2 km) grid spacing, with explicit convection and 80 vertical levels. Previous work
179 suggests that a fine horizontal resolution is necessary for research of this kind. Bhatt et al. (2016)
180 found that, in a simulation over the Maritime Continent using the Weather Research and Forecasting
181 (WRF) model with parametrized convection, a 10 km grid was not sufficient to capture all the
182 local interactions between density currents. However, Birch et al. (2013) successfully captured
183 the initiation of west African convective storms due to cold pools, using convection-permitting
184 simulations on a 4 km grid.

185 The Regional Atmosphere v1 in the Tropics (RA1T) science configuration was used in our
186 simulations (described in Bush et al. 2020 as RAL1-T). SST was fixed throughout each forecast
187 using data from the Operational Sea surface Temperature and Ice Analysis (OSTIA). The domain
188 is shown in figure 3 and the model was forced at the boundaries by the ECMWF global forecast
189 initialized at the same date and time. Forecasts were initialized at 00 UTC daily from the ECMWF
190 global 0.1° analysis, and the first 24 hours of each run were discarded to allow the model to spin
191 up. Output from $T + 24$ to $T + 54$ (07 LT to 13 LT the following day) was analysed for each forecast.
192 Diagnostics were output every 5 minutes.

193 Much of this paper focuses on a single forecast, initialized at 00 UTC on 2015/11/29. This is
194 referred to as the 2015/11/30 case study.

195 *b. Observations*

196 Observations of precipitation are taken from the Global Precipitation Measurement (GPM)
197 Integrated Multi-satellitE Retrievals for GPM (IMERG) product (Huffman et al. 2019). The

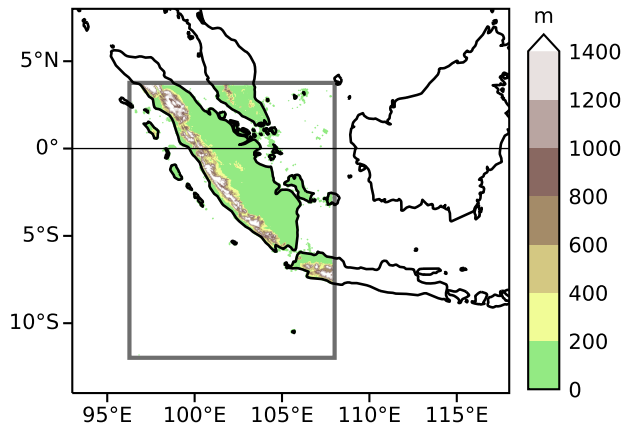


FIG. 3: Model orography for the convection-permitting MetUM experiment. The grey rectangle is the edge of the model domain and its thickness indicates the width of the lateral boundary forcing region.

198 IMERG algorithm merges precipitation estimates from microwave satellites, where available, and
 199 infra-red measurements, with calibration performed against data from a network of rain gauges.
 200 The data set is on a 0.1° grid at half-hourly temporal resolution.

201 During the pre-YMC field campaign, the *R/V Mirai* was stationed approximately 50 km offshore
 202 from the city of Bengkulu, on the south-west coast of Sumatra (red dots in figure 1a). An AWS
 203 recorded data every 10 minutes. In section 3c we use wind (measured at a height of 25 m); virtual
 204 potential temperature θ_v , which we compute from temperature and humidity (measured at 21 m);
 205 and precipitation rate.

206 Also on board the ship was a C-band polarimetric radar. We convert radar reflectivity Z in
 207 $\text{mm}^6 \text{m}^{-3}$ to precipitation rate R in mm hr^{-1} using the Marshall-Palmer relation (Marshall and
 208 Palmer 1948)

$$Z = \alpha R^\beta, \quad (1)$$

209 where the parameters α and β were derived empirically by Yokoi et al. (2017) as $\alpha = 216$ and
 210 $\beta = 1.28$, using the method of Yokoi et al. (2012). Calibration was performed against a rain gauge
 211 at Bengkulu during 2015/11/23–2015/12/14 (i.e., the pre-YMC field campaign period, except
 212 during an active MJO event). The case studies examined in this paper fall within this time period.

213 Observations of brightness temperature are taken from the Himawari-8 geostationary satellite
 214 at $10.4 \mu\text{m}$ (Himawari channel 13). Images are available every 20 minutes. We transform the

215 full-disc images to a Cartesian grid with a grid spacing of 2 km, using the `gdalwarp` command
216 from the Geospatial Data Abstraction Library (GDAL, 2022).

217 Orography is taken from the Global Land One-km Base Elevation (GLOBE) project (Hastings
218 and Dunbar 1998; Hastings et al. 1999).

219 *c. Methods*

220 To investigate gravity wave activity in the model, we compute spectra in wavenumber-frequency
221 space, as will be used in figure 9. A Hovmöller diagram of vertical velocity w , on a transect
222 running along the direction of the nocturnal offshore propagation of convection, is taken at a
223 selected vertical level. By analogy with Wheeler and Kiladis (1999), the fast Fourier transform
224 (FFT) algorithm is applied in the x -dimension (onshore distance) and in time, to obtain complex
225 coefficients in wavenumber k and frequency ω . Experiments showed that, although the w fields
226 are not periodic in time or distance, the edge effects when computing the Fourier transforms are
227 negligible. The modulus of these coefficients is computed. Following Peatman et al. (2018), a
228 background spectrum, which is a function of ω , is found by averaging over k . We divide through
229 by this background and plot \log_{10} of the resulting field. For ease of interpretation, the k and ω
230 axes are labelled with values of wavelength $|1/k|$ and period $1/\omega$, respectively.

231 The theoretical horizontal phase speed of a gravity wave of mode $n \in \mathbb{N}$ is (e.g., Lane and Reeder
232 2001)

$$\frac{\omega}{k} = \pm \frac{1}{n} \frac{NZ_T}{\pi}, \quad (2)$$

233 where Z_T is the tropopause height (taken to be 15.5 km) and N is the Brunt-Väisälä frequency, which
234 we diagnose from the model. We overlay these theoretical speeds on the spectra in (k, ω) space,
235 remembering the caveat that they are derived from dry theory so do not necessarily correspond
236 exactly to the speeds of the waves that are present.

237 Gravity wave activity is further demonstrated (see figure 10) by high-pass filtering w to obtain
238 w_{hp} . We use a Lanczos filter (Duchon 1979) with a cut-off frequency of $\omega_c = 1/(60 \text{ minutes})$ to
239 remove low-frequency variability associated with deep convection, which develops on time scales
240 of a few hours. Since gravity waves, according to theory, have the vertical profile of a standing
241 wave with $n + 1$ nodes, we use the tropospheric column average of $|w_{hp}|$ as a metric for gravity
242 wave activity, the absolute value ensuring that even-numbered modes are not averaged out.

243 3. Results

244 a. Forecast verification and choice of case studies

245 We evaluate the performance of the forecasts first by comparing the mean precipitation rate
246 over the entire 2015 field campaign period (figure 4). In IMERG observations (figure 4a), the
247 precipitation during this time was strongly focused offshore to the south-west of Sumatra. Notably,
248 this is in contrast to the November–April (NDJFMA) climatology in figures 1a–c, which has
249 intense precipitation over the land also. The MetUM forecasts (figure 4b) more closely resemble
250 the NDJFMA climatology, with intense precipitation over the mountains, especially in the southern
251 half of the island. However, the model does have considerable precipitation offshore, as in the
252 observations; and this is generally strongest in the south, again agreeing with IMERG.

253 Figure 4c shows the difference between the first two panels. There is mostly a positive bias over
254 land as already discussed, on the order 0.5 mm hr^{-1} . Over the sea to the west of Sumatra, we
255 find strong positive and negative biases. The spatial distribution of rainfall is not realistic on local
256 scales, with the offshore precipitation being too intense around $4\text{--}5^\circ\text{S}$ and too weak either side of
257 this. However, the domain-average rainfall rate is a good match with observations, as the mean
258 over the whole of figure 4c is close to zero (-0.07 mm hr^{-1}).

259 We further verify the model by considering the offshore propagation of precipitation south-west
260 of Sumatra. Figures 5a and c show a Hovmöller diagram of observed precipitation rate with time
261 running down the page. Overlaid in the black contour is the *R/V Mirai* radar (not available on
262 2015/11/22), interpolated onto the same fixed box as for IMERG except that it is limited in the
263 onshore-offshore direction due to the radar’s range. Where this range narrows, it is due to the
264 motion of the ship away from its nominal station. The blue dashed lines show the locations of the
265 *R/V Mirai* and Bengkulu. Figures 5b and d show the model precipitation rate from $T + 24$ to $T + 48$
266 (07 LT to 07 LT) for each forecast. Horizontally ruled lines are at 00 UTC (07 LT), where there
267 are some discontinuities due to forecasts being concatenated together.

268 The forecasts are mostly successful at capturing which days are wet or dry, and usually manage
269 to capture something of the offshore propagation. For example, on 2015/11/30 (see figures 1g–l
270 for observations) the precipitation is initiated over the mountains at approximately the right time
271 of day (in the afternoon), with propagation offshore which continues until the following morning.

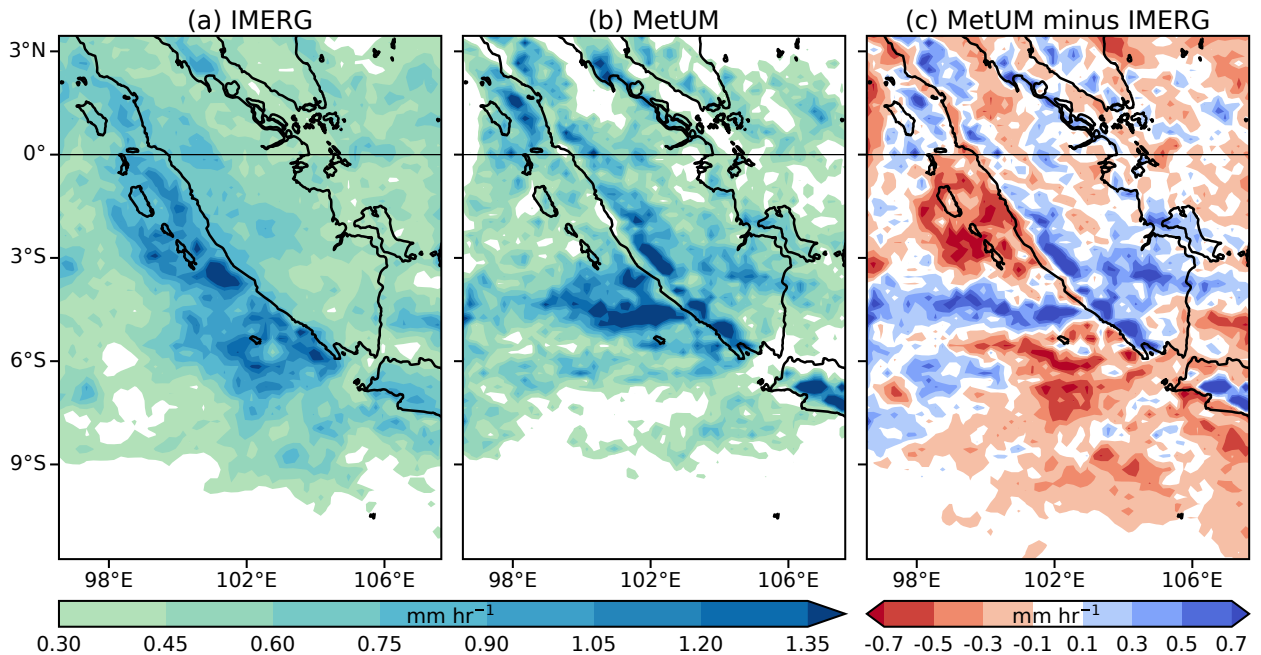


FIG. 4: (a) Observed precipitation rate from IMERG averaged over the 2015 field campaign (2015/11/22 to 2015/12/17). (b) As (a) but for the MetUM forecasts, regridded to the IMERG grid. (c) Bias in the MetUM (panel (b) minus panel (a)).

272 The propagation speed in the forecast is too slow, but it continues to around 99°E or beyond by
 273 07 LT on 2015/12/01. (On 2015/12/01, the subsequent forecast correctly has more intense rainfall
 274 beyond 99°E.) Further examples of realistic propagation are on 2015/11/25, although the model
 275 has too much larger-scale rainfall offshore during the morning; and 2015/12/03, although the
 276 propagation is again a little too slow. From 2015/12/13, an active MJO event occurred (in phase 4,
 277 using the Realtime Multivariate MJO (RMM) indices of Wheeler and Hendon 2004), associated
 278 with relatively large-scale rainfall which tends to propagate onshore. The transition to this new
 279 pattern of propagation is also well forecast.¹ Days on which the forecasts do not verify so well
 280 against observations include 2015/12/07, when intense offshore precipitation throughout the day
 281 was observed; and 2015/12/10, which does not capture the intense land-based precipitation or its
 282 propagation, instead having intense rain over the sea.

283 For the remainder of this section we focus on three case studies, namely 2015/11/25, 2015/11/30
 284 and 2015/12/03, all of which have offshore propagation in the observations which is well represented
 285 by the forecasts. As explained above, there are forecast biases for these three cases, but from this

¹The daily mean rainfall in the Sumatra region on these days was low compared with most of the field campaign, which is unusual for an active MJO event. This is explained by a dry air intrusion from the extra-tropical southern hemisphere, which occurred concurrently.

286 point we are interested in these forecasts as research tools, to understand the physical mechanisms
287 behind the offshore propagation of convection. Therefore, having established that the broad features
288 of these case studies are correctly forecast, we are not concerned with smaller discrepancies from
289 observations, which are typical of forecasts of convective activity.

290 Maps of precipitation rate are shown for each of these case studies for four selected times of day in
291 figure 6, starting during the afternoon (15 LT) and evening (21 LT), and following the propagation
292 through the early morning of the following day (01 LT and 05 LT). Overplotted maroon contours
293 show the magnitude of the horizontal gradient of θ_v ,

$$|\nabla_h \theta_v| \equiv \sqrt{\left(\frac{\partial \theta_v}{\partial x}\right)^2 + \left(\frac{\partial \theta_v}{\partial y}\right)^2},$$

294 at 45 m above sea level. This is plotted over sea only as it is a noisy field over land; and only
295 the 0.1 K km⁻¹ contour is shown. This indicates the presence of low-level fronts, which may be
296 caused by either the land breeze converging with environmental onshore wind; or cold pools due
297 to convection. In some places it is not easy to distinguish the two, but the orange shapes (drawn
298 subjectively) approximately indicate the land breeze location, with possible reinforcement from
299 cold pools. The red boxes, again drawn subjectively, follow the convection offshore and are used
300 for subsequent Hovmöller diagrams and vertical cross-sections (figures 7–12).

301 In all three cases, precipitation forms over the mountains near the coast by 15 LT, with rain
302 rates exceeding 15 mm hr⁻¹ in places. By 01 LT, the main rainfall has propagated from the island
303 to the coastal sea. Over land in the afternoon, precipitation is represented as a large number
304 of small-scale and intense convective cells, while the offshore precipitation overnight is more
305 organized, as a squall line. This squall line propagates away from the coast along with the land
306 breeze front. Note that neither the squall line nor the land breeze front necessarily preserve the
307 shape or orientation of the coastline. Instead, they may be curved (e.g., figure 6c) or split into
308 multiple features (e.g., figures 6g–h), which may be due to factors such as the environmental wind
309 or interactions with cold pools.

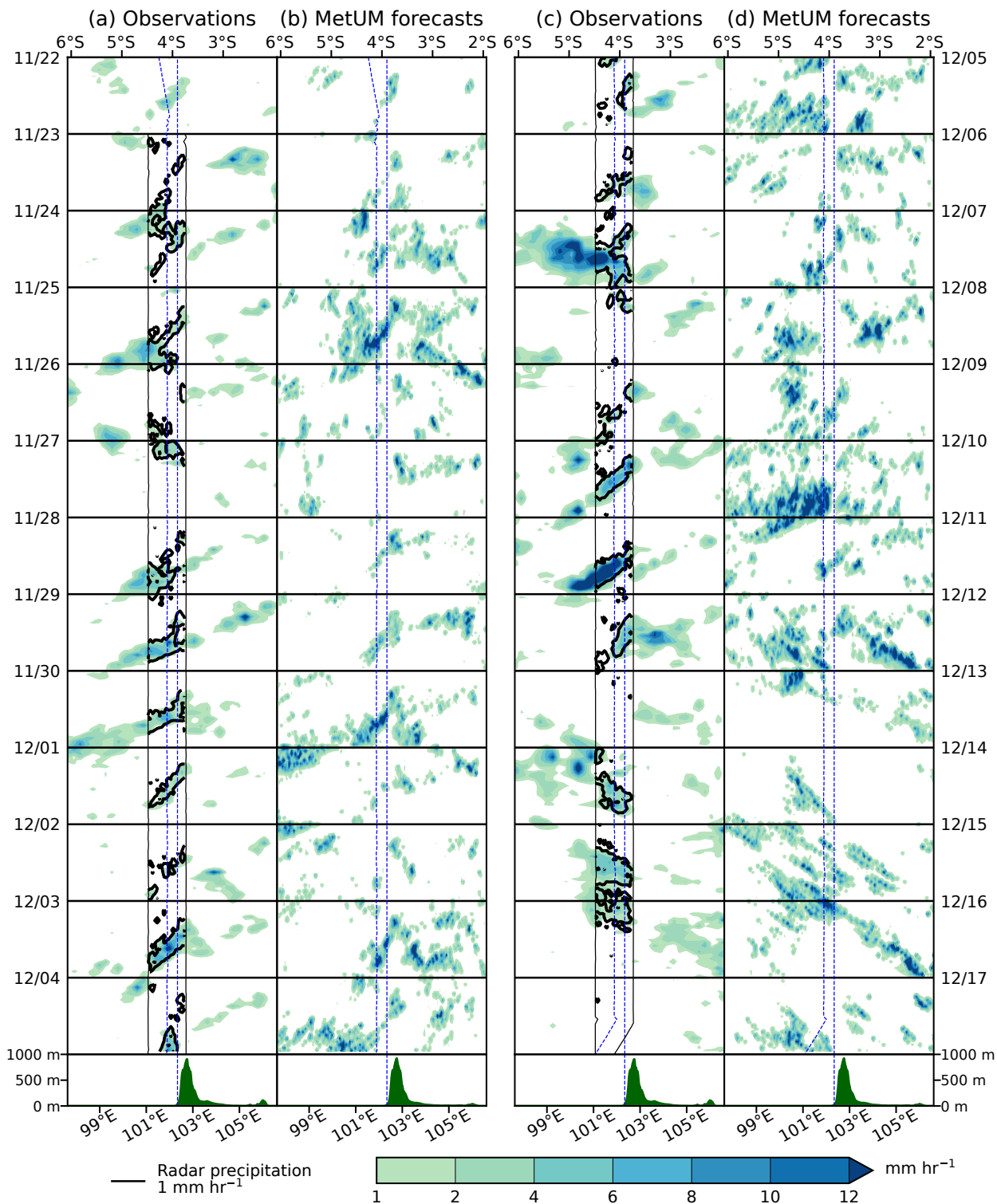


FIG. 5: Hovmöllers of precipitation rate during the 2015 field campaign taken along a transect passing through Bengkulu and the mean *R/V Mirai* position, averaged over a box of width 1° , from (a,c) GPM IMERG (shading) and the *R/V Mirai* radar (black contours); and (b,d) the MetUM forecasts. (a,c) Solid black vertical lines indicate the range of the radar transect. Dashed blue lines indicate the track of the *R/V Mirai* and the location of Bengkulu. Horizontal grid lines are at 00 UTC (07 LT); successive forecasts are concatenated at these times. Mean orography across the transect is shown in green below each panel.

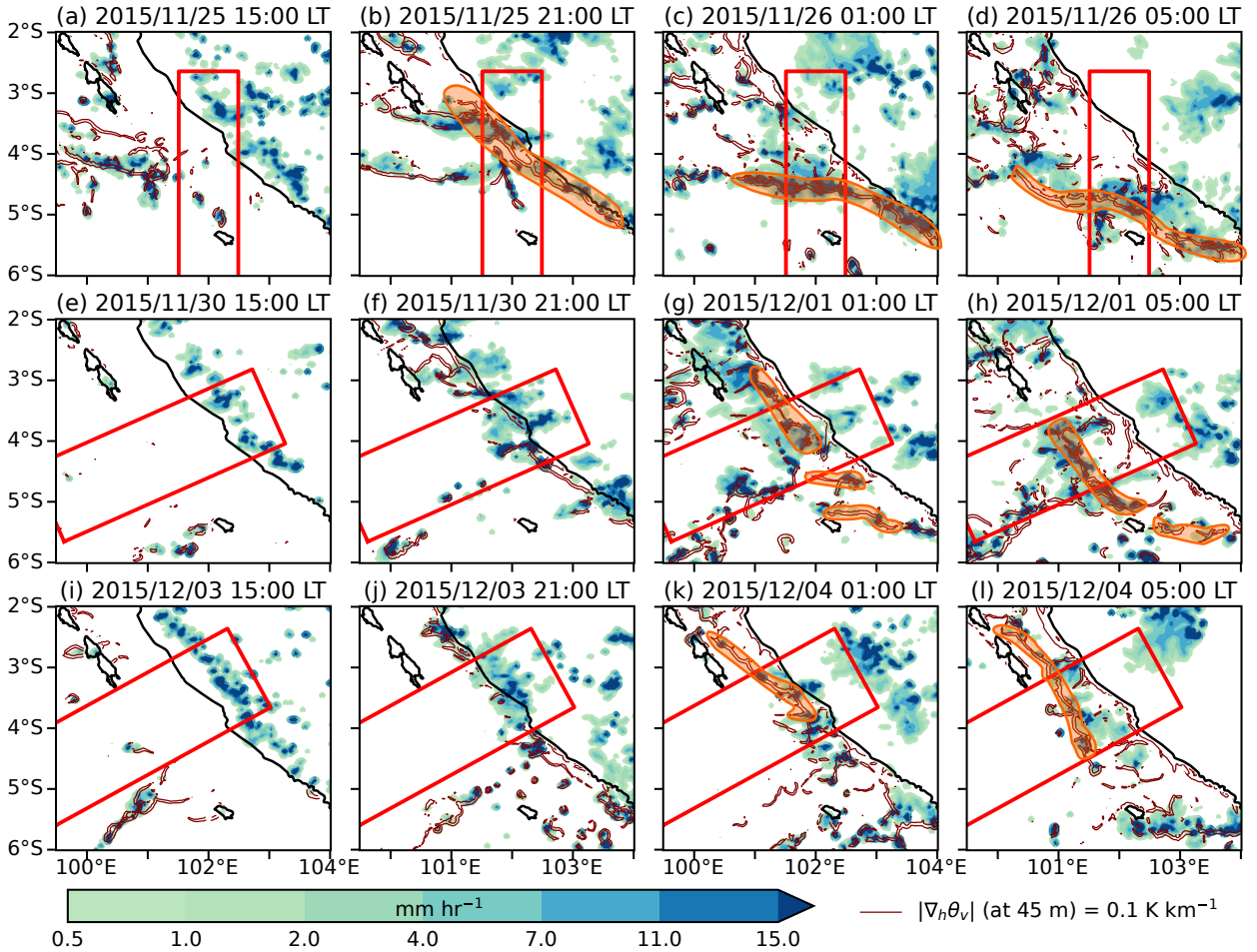


FIG. 6: Forecast precipitation rate at selected times from the (a) 2015/11/25, (b) 2015/11/30 and (c) 2015/12/03 case studies. The magnitude of $\nabla_h \theta_v$ at 45 m altitude is overplotted in maroon, over the sea only, to identify low-level fronts caused by the land breeze and cold pools. The approximate location of the land breeze front is indicated by the orange shapes. The red boxes are drawn subjectively to follow the propagation of convection and are the transects used for figures 7–12.

310 *b. Mechanisms of offshore propagation*

311 To investigate the physical mechanisms of the offshore propagation more closely, we consider
 312 vertical cross-sections of the 2015/11/30 case (figure 7). These are averaged over the red box in
 313 figures 6e–h, so the long edge of the box is the horizontal axes of the plots.

314 For this case study, the wind in the free troposphere blows offshore (coloured shading and vectors)
 315 throughout the day in almost all locations shown. Close to the ground, by local noon (figure 7a)
 316 there is an onshore sea breeze. This is driven by the land-sea temperature contrast, which peaks
 317 around this time at 6.3°C (not shown; taking the difference between the means up to 20 km either

318 side of the coast; but recall from section 2a that there is no diurnal cycle of SST in this model).
319 The sea breeze results in upslope flow and, at the peak of the mountains, there is moisture flux
320 convergence (MFC; red curve on lower panel). There are low- to-mid-level clouds (grey contours),
321 probably cumulus, at a fairly low concentration directly above the mountain peaks, but they are not
322 precipitating (blue curve on lower panel). The lack of rainfall is typical for this time of day.

323 By 16 LT (figure 7b), deep convection has been triggered over the mountain, forming
324 cumulonimbus clouds with associated intense precipitation, varying from around 5 to 13 mm hr⁻¹.
325 The sea breeze has intensified and extended further offshore. By 20 LT (figure 7c), the convection
326 has deepened further, reaching the tropopause. As the land surface cools, a katabatic flow begins
327 on the seaward side of the mountains, as seen by the blue shading (horizontal flow in the offshore
328 direction) around 5–40 km inland and the lower- θ_v air (black contours). As a result, there is
329 low-level convergence with the sea breeze just inland of the coast. The deep convection and high
330 precipitation rate (10–16 mm hr⁻¹) are collocated with this convergence, rather than being centred
331 near the mountain peaks.

332 Through the late evening and early morning (figures 7d–f), the density current initiated as the
333 katabatic wind continues to flow downhill and offshore as a land breeze current. During these
334 hours (22–03 LT), the land surface is cooler than the sea surface by about 4–5°C in the model. Air
335 behind the cold front has θ_v in the range 301–303 K, whereas in the well-mixed boundary layer
336 ahead of the front we typically have $\theta_v \sim 304$ K. The strongest ascent of air, low-level convergence
337 and precipitation propagate along with the front, with the precipitation rate gradually weakening
338 (13–22 mm hr⁻¹ at 22 LT and 4–17 mm hr⁻¹ at 03 LT). This is consistent with the mean diurnal
339 cycle, in which precipitation has propagated beyond the coast by 22 LT (figure 1b), weakening as
340 it propagates further through the night (figure 1c).

341 At a number of times and locations, gravity waves can be seen in the wind field. For example,
342 alternating ascending and descending air is visible above around 9 km altitude across the domain
343 in figures 7a and b; and around 3–6 km altitude at around 75 km and 125 km offshore in figure 7b.
344 Note also the ascent around 1–3 km altitude at around 150 km offshore in figure 7c. Note further
345 that there is convection and associated precipitation away from the land, not directly associated with
346 the initial convection over the mountains, around 190 km offshore (figures 7d and e). Inspecting
347 the maps of precipitation rate (figures 6f and g) reveals that this convection (around 101°E, 5°S) is

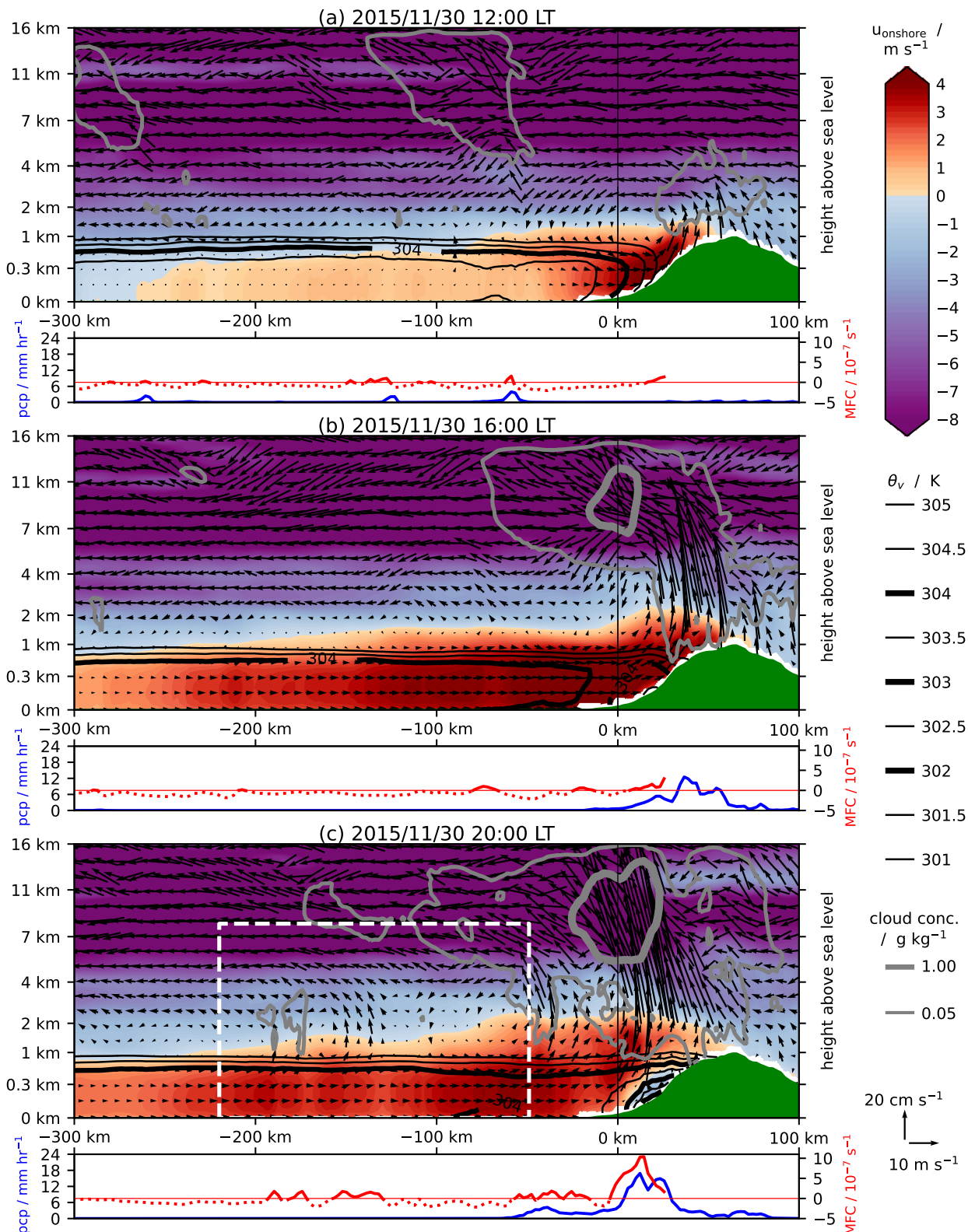


FIG. 7: Vertical cross-sections at selected times of the forecast of the 2015/11/30 case study, showing onshore and vertical wind (vectors, with onshore wind also in coloured shading); virtual potential temperature θ_v (black contours, shown up to 305 K only to illustrate the thermodynamics of the boundary layer, which is usually well-mixed); liquid+solid cloud concentration (grey contours); orography (green); precipitation rate (blue curve); and along-transect moisture flux convergence (MFC) averaged over 0–500 m above sea level (red curve; solid for convergence and dotted for divergence). The white dashed box in panel (c) is the domain of figure 11. The vertical axes of the main panels (height above sea level) are non-linear, to emphasize the lower troposphere.

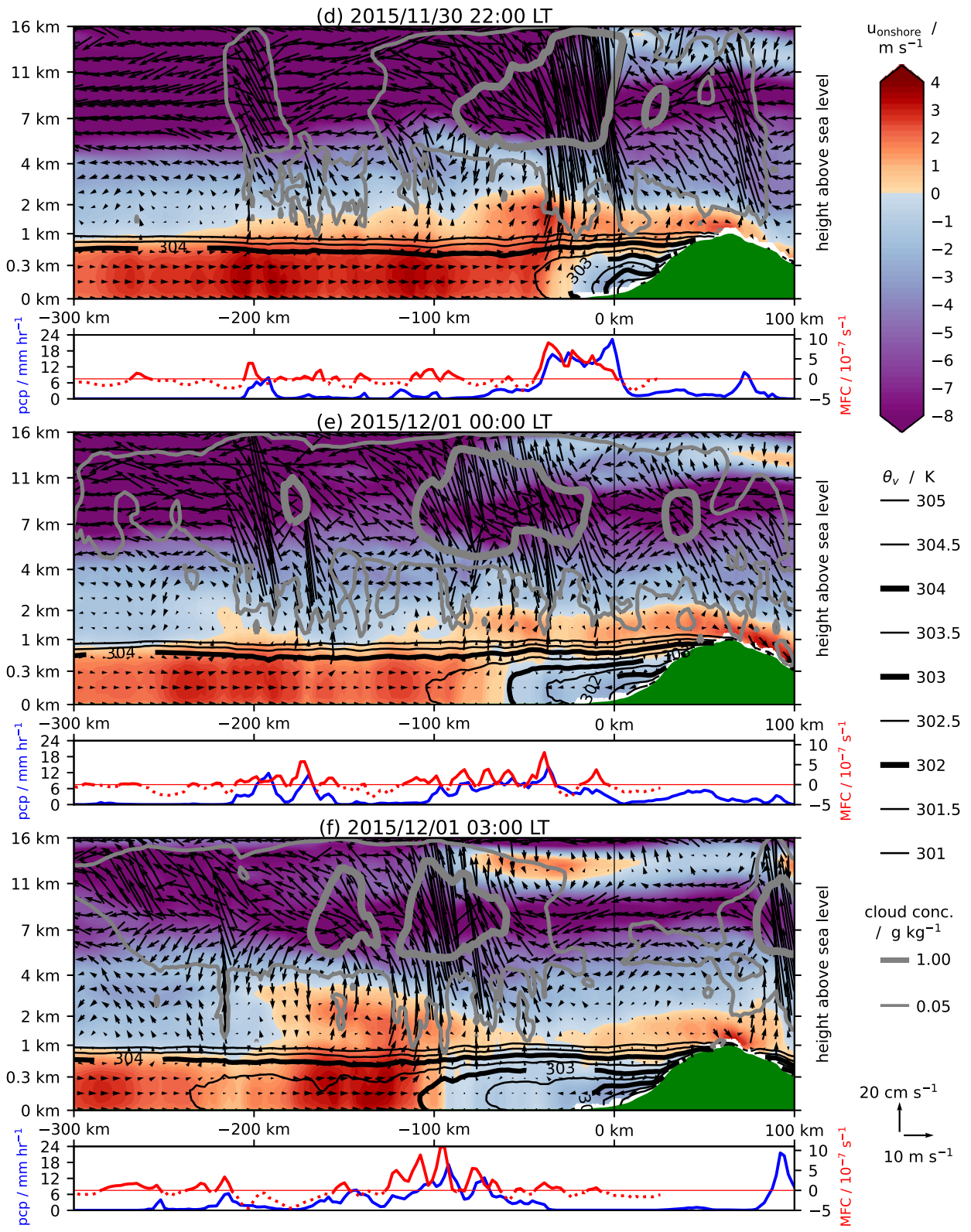


FIG. 7: Continued.

348 far more localized than that which propagates away from the coast with the land breeze, the latter
349 being a large, organized squall line. The waves and offshore convection will be discussed in more
350 detail later in this section.

351 Having examined the vertical cross-sections of one case study and seen that the propagating
352 squall line follows the low-level convergence due to the land breeze front, we now demonstrate
353 that this is common to all three of our case studies. Figures 8a–c show Hovmöller diagrams of the
354 onshore component of 10 m wind. The red shading during the day is the sea breeze, with the land
355 breeze in blue being initiated slightly inland over the mountains at around 16–18 LT. It propagates
356 offshore at approximately $2.5\text{--}3.5\text{ m s}^{-1}$. As seen in the vertical cross-sections of the 2015/11/30
357 case, in all three cases the precipitation propagates along with the land breeze front, sometimes
358 with a lag of 1–2 hours between the change in wind direction and the heaviest rainfall. This may
359 suggest the convection can take time to respond to the changing low-level wind; or it may be an
360 indication of the convection forcing a cold pool in front of it as it advances.

361 Figures 7 and 8a–c have considered the onshore component of wind and convergence only.
362 To confirm that the precipitation is synchronized with the total low-level horizontal convergence,
363 we also show Hovmöller diagrams of the 10 m divergence in figures 8d–f. Again, we see the
364 precipitation is synchronized with the convergence (blue shading).

365 The evidence, therefore, points towards a similar mechanism as that of Houze et al. (1981), with
366 the convergence between the land breeze and the environmental winds being the chief driver of
367 the nocturnal offshore propagation. However, given the numerous references to the role of gravity
368 waves in the literature (see section 1), we now turn our attention to the gravity waves in the model
369 to examine any role for them, using the 2015/11/30 case as an example.

370 Figure 9 shows wavenumber-frequency spectra, computed as described in section 2c, at (a) 2 km
371 and (b) 13 km above sea level. At 13 km, the gravity wave activity is significant (i.e., distinguishable
372 from the background spectrum, which has been removed) in the offshore direction across all
373 frequencies up to the Nyquist frequency of $\omega_N = 1/(10\text{ minutes})$. At 2 km the picture is very
374 similar, but with less spectral power at the very highest frequencies. The absence of power in the
375 onshore direction is likely due to the prevailing wind direction. The theoretical gravity wave phase
376 speeds were computed using $N = 0.01014\text{ s}^{-1}$, a mean value diagnosed from the model. At both
377 altitudes, there is power in the $n = 2$ (25.0 m s^{-1}), $n = 3$ (16.7 m s^{-1}) and $n = 4$ (12.5 m s^{-1}) modes.

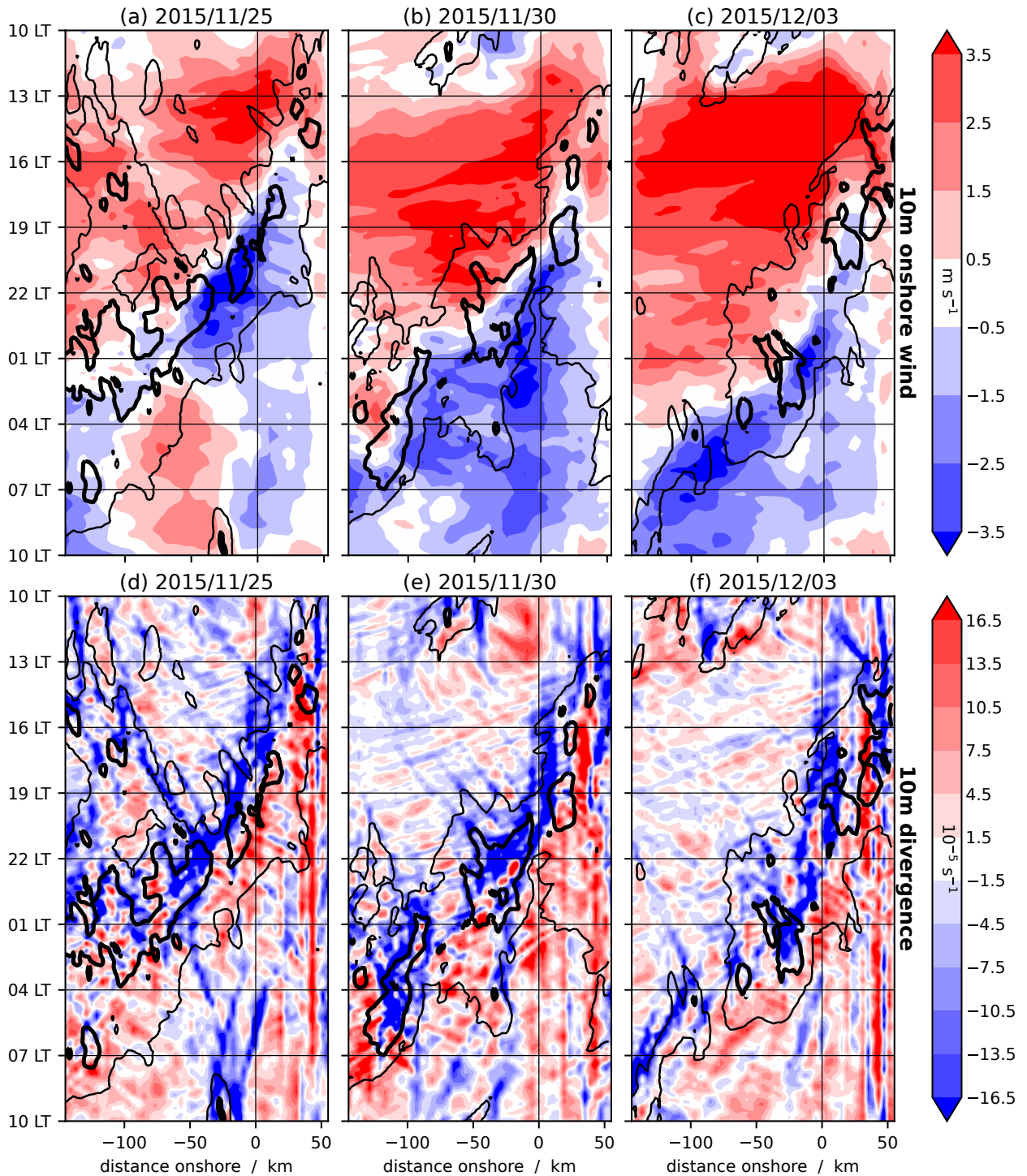


FIG. 8: Hovmöllers of 10 m (a–c) onshore wind and (d–f) horizontal divergence, with precipitation rate overlaid (contour levels: 1 and 10 mm hr⁻¹), for case studies on (a,d) 2015/11/25, (b,e) 2015/11/30 and (c,f) 2015/12/03.

378 It is perhaps surprising that the $n = 1$ (50.0 m s^{-1}) mode appears to be absent. However, we note
 379 that observed gravity waves propagate slightly more slowly than in the dry theory used to estimate
 380 theoretical speeds here; and the waves do not necessarily propagate in the same direction as the
 381 transect chosen in figures 6e–h, so it is possible they are diagnosed as propagating more slowly
 382 than their true speed, due to their projection onto the chosen onshore-offshore coordinate. The box
 383 is at an angle of about 45° to the equator, so in a fairly extreme case of the waves propagating due
 384 west or due south, the diagnosed speed would be a factor of $\cos 45^\circ$ times the true speed. Hence,
 385 we can consider that the wave speeds may be as diagnosed or up to around 40% faster.

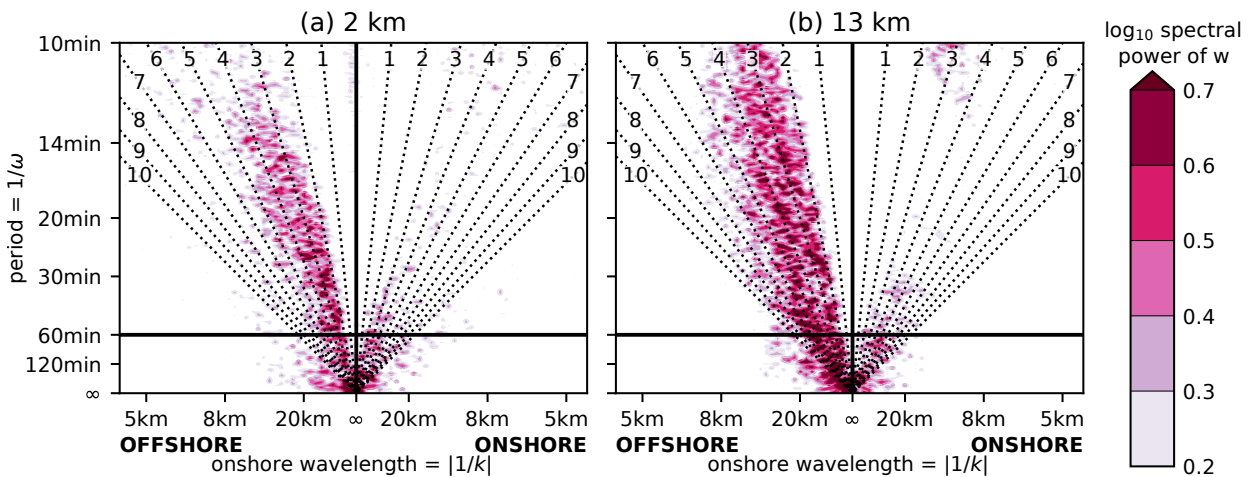


FIG. 9: Wavenumber-frequency spectra of w at (a) 2 km and (b) 13 km altitude, from a Hovmöller diagram of the 2015/11/30 case study forecast, with the background spectrum removed. Dashed lines are theoretical dispersion relations for gravity waves of modes $n \in \mathbb{N}$; see equation (2). A horizontal line is drawn at a period of 60 minutes, which is used as the cut-off frequency for the high-pass filter of w used in figures 10a, c and d.

386 Having demonstrated that spectra of w have power along the dispersion lines of gravity waves,
 387 we consider how their propagation relates to the occurrence of convection in the model. Figure 10b
 388 is a Hovmöller diagram of the column-mean (over the troposphere) of w for the 2015/11/30 case
 389 study. This, of course, most prominently shows the deep convection, in the brighter colours. We
 390 also see, in purple, small-scale features propagating at faster speeds. These are the gravity waves
 391 which were found in figure 9. To isolate these waves from the convection, we use a high-pass filter
 392 as described in section 2c (figure 10c). Precipitation rate and theoretical gravity wave speeds are
 393 overlaid. This is repeated for the other two case studies in figures 10a and d.

394 As expected from the wave spectra, the speeds of propagation are consistent with modes $n = 2-4$.
 395 The propagation of deep convection is much slower, at the speed of a theoretical wave of order
 396 $n = 10$ or even higher, which is much higher than typically observed, so it is not reasonable to
 397 suppose that this convection propagates as a direct result of gravity waves.

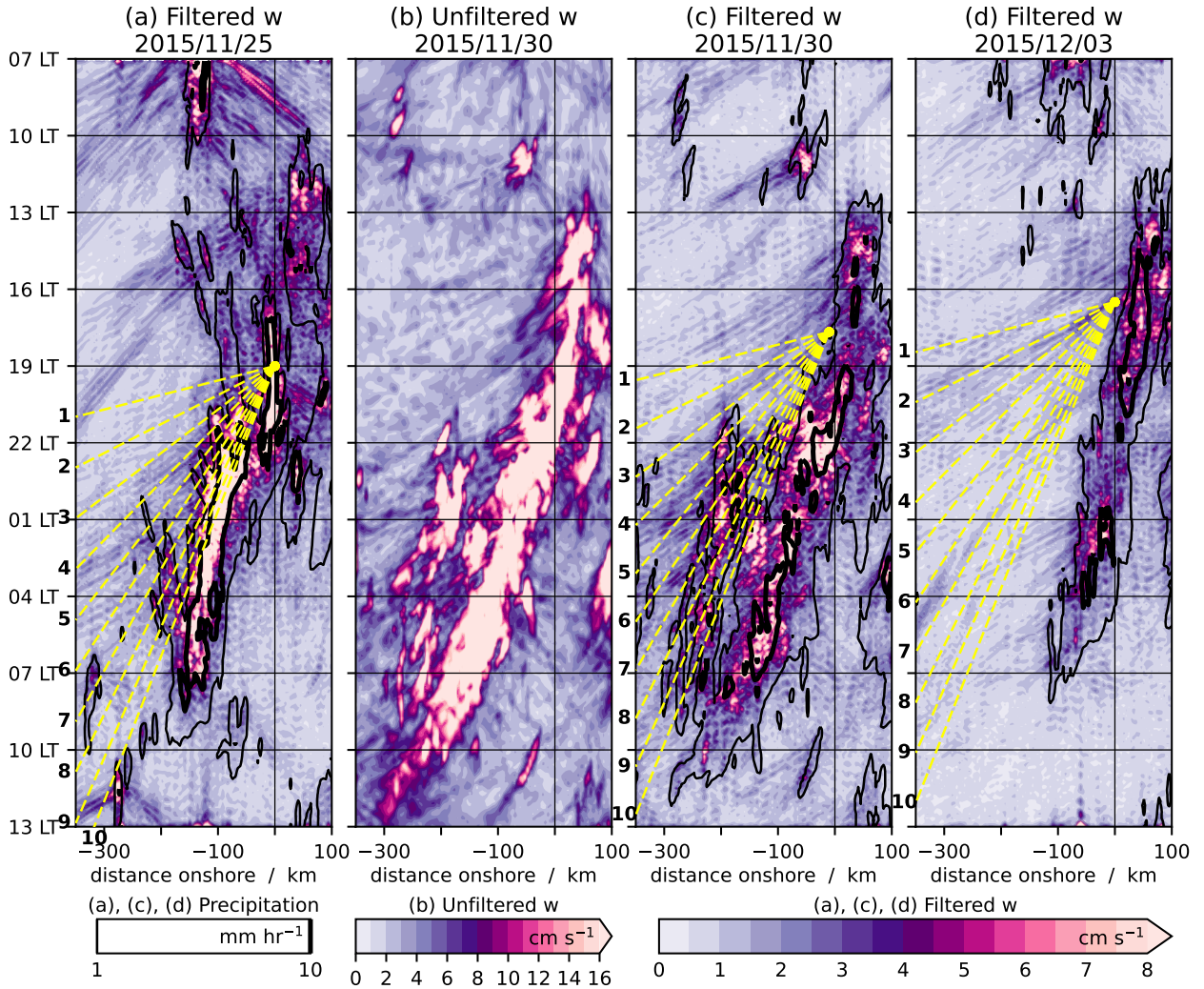


FIG. 10: (a,c,d) Hovmöller diagrams of column-mean absolute value of w_{hp} (frequency cut-off at 60 minutes, shown by the horizontal line in figure 9), with precipitation rate overlaid, for the 2015/11/25, 2015/11/30 and 2015/12/03 case studies, respectively. Also overlaid are theoretical gravity wave phase speeds, starting at a subjectively-chosen time and position (marked with a yellow dot), and labelled with their mode number n . (b) As (c) but for w , not w_{hp} .

398 However, there are some instances of precipitation occurring ahead of the main branch of
 399 propagating convection. Examples for 2015/11/25 (figure 10a) include -130 km at 22:15 LT,
 400 -145 km at 23:00 LT and -235 km at 00:00 LT; and for 2015/11/30 (figure 10c), -105 km at

401 20:25 LT and -175 km at 20:35 LT. Since these regions of precipitation occur well ahead of the
402 land breeze, it is conceivable that they are associated with the gravity waves.

403 The latter example (-175 km in figure 10c) develops into the intense convection which was
404 highlighted earlier in this section, occurring at around -190 km in figures 7d and e. In this instance,
405 we can see the process of a gravity wave triggering the convection in vertical cross-sections.
406 Figure 11 is a repetition of figure 7 but for the white dashed region in figure 7c only, between
407 19 LT and 21 LT. The ascending air highlighted with the green ellipse is a gravity wave (see the
408 discussion of figure 12, below), propagating at around 16 m s^{-1} , which is close to the theoretical
409 speed of an $n = 3$ wave. In this instance it happens that there is a small perturbation at around
410 -190 km, in the form of a small region of non-precipitating cloud. This was caused by a cold pool
411 (not shown) which propagated away from some other convection found offshore earlier in the day
412 (which can be seen around $101\text{--}102^\circ\text{E}$, $5.5\text{--}6^\circ\text{S}$ in figure 6e, outside the transect area). By 21 LT,
413 figure 11d shows that the wave has reached this perturbation, strengthening the ascending air there
414 and causing the cloud to precipitate. An hour later, the convection has reached the tropopause
415 (figure 7d) and by midnight it has intensified into a major convective feature (figure 7e). The
416 feature can be seen centred around -190 km in figure 10c.

417 We can satisfy ourselves that the propagating region of ascending air in figure 11 really is a
418 gravity wave by studying figure 12. Here, we take Hovmöller diagrams of w (no longer filtered
419 or vertically-averaged), at 2 km above sea level since this is where the feature was seen in the
420 vertical cross-sections. In figure 12b we overlay contours of potential temperature θ . The yellow
421 dashed-dotted line is the feature in question. We see that w and θ are in quadrature, as expected for
422 a gravity wave. For example, at -125 km there is a maximum in θ at 19:25 LT, after the descending
423 (blue) phase of the wave. This is consistent, as the descent has brought down higher- θ air from
424 above. The maximum in w is at 19:45 LT, at which point θ is decreasing, which is again consistent
425 as the ascent is bringing up lower- θ air from below.

426 The yellow stars in figure 12a are locations where ascent is triggered along the trajectory of the
427 gravity wave propagation, followed by precipitation within around 15–20 minutes. In figure 12b we
428 see that these ascent features do not propagate or have the same quarter-phase relationship with θ .
429 Instead, the maximum in w coincides with a maximum in θ , consistent with diabatic heating due to
430 convection. Since these instances of precipitation are mostly low-intensity, small-scale and isolated

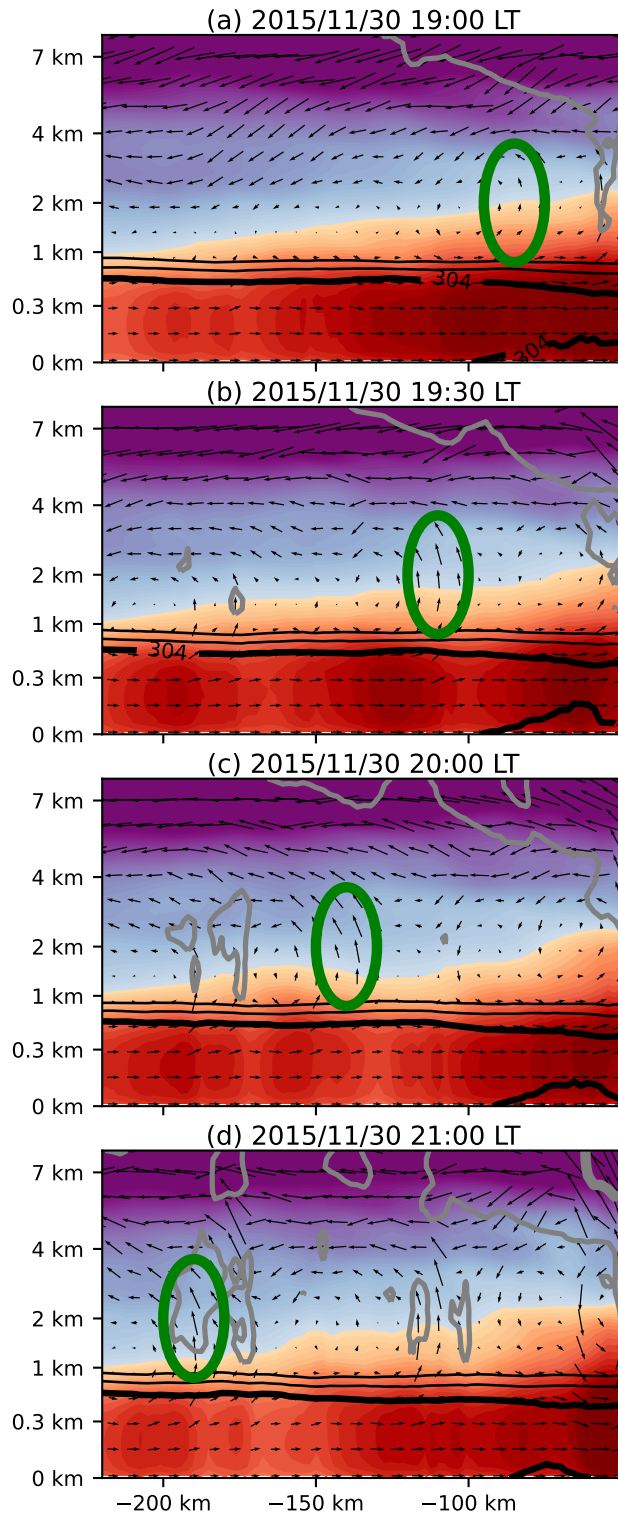


FIG. 11: As figure 7 but for a limited horizontal and vertical extent, as indicated by the white dashed box in figure 7c, and for different selected times. Green ellipses indicate the ascending phase of a particular gravity wave (see figure 12) propagating offshore. For a description of the plotted quantities, see the colour bars and caption of figure 7.

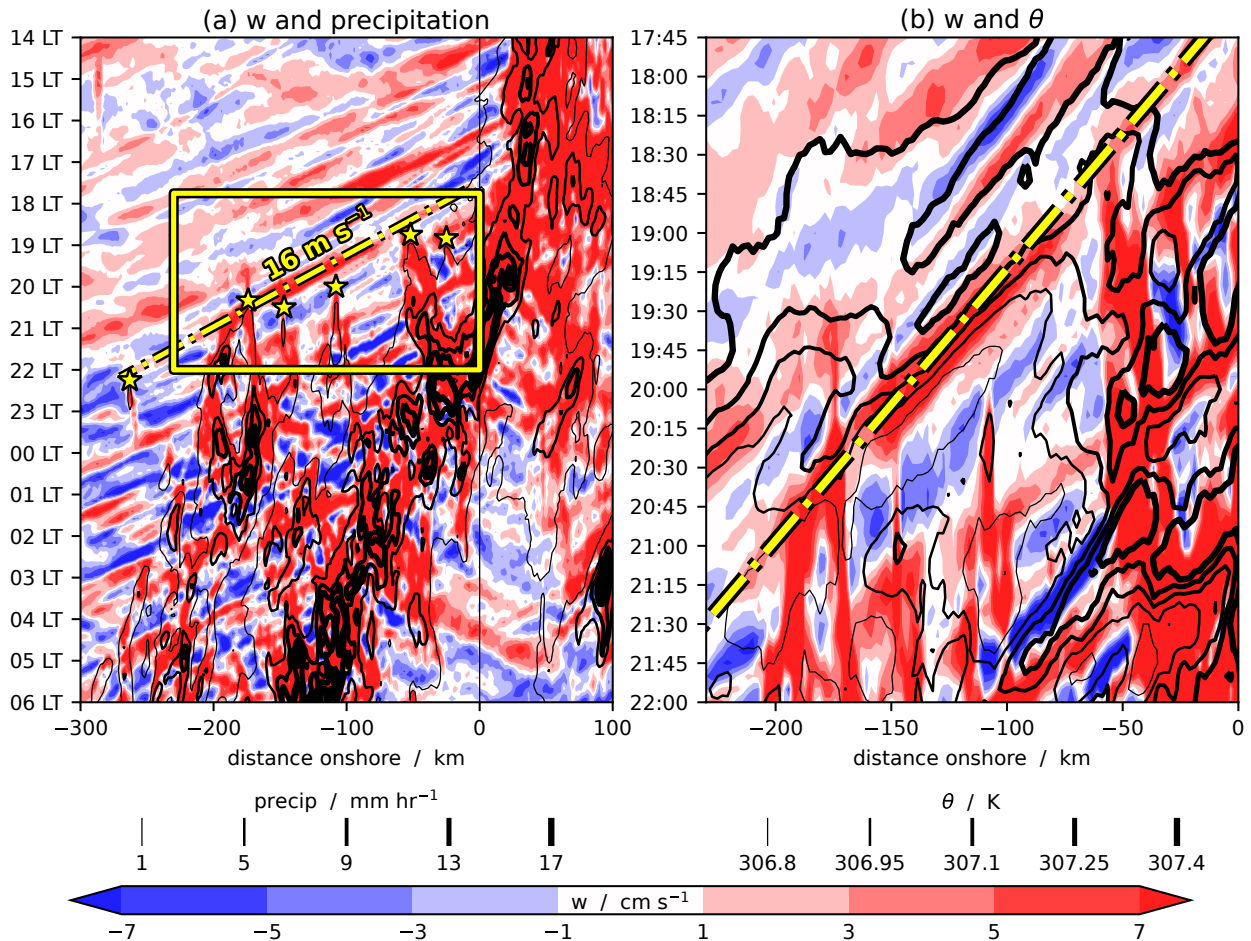


FIG. 12: Hovmöller diagrams of unfiltered vertical velocity w at 2 km altitude from the 2015/11/30 forecast case study (coloured shading). The yellow dashed-dotted line is the gravity wave indicated in figure 11. Black contours are (a) precipitation rate and (b) potential temperature θ at 2 km altitude. In panel (a), the six yellow stars indicate initiation of precipitation, likely triggered by the gravity wave; and the yellow box is the domain of panel (b).

431 (not organized), they cannot truly be considered propagation. However, when compositing over
 432 many days, such precipitation would be smoothed out in the averaging and appear as propagation.
 433 This explains the faster part of the widening propagating envelope that is seen in figure 2a.

434 *c. Verification of physical mechanisms*

435 The previous section proposed physical mechanisms of the offshore propagation, derived from
 436 an in-depth analysis of high-resolution model case studies. There are not many observations that
 437 can be used to verify these mechanisms, but the 2015 field campaign (see section 2b) provides us
 438 with point measurements from the *R/V Mirai*. In figures 13a–g we present time series of forecast

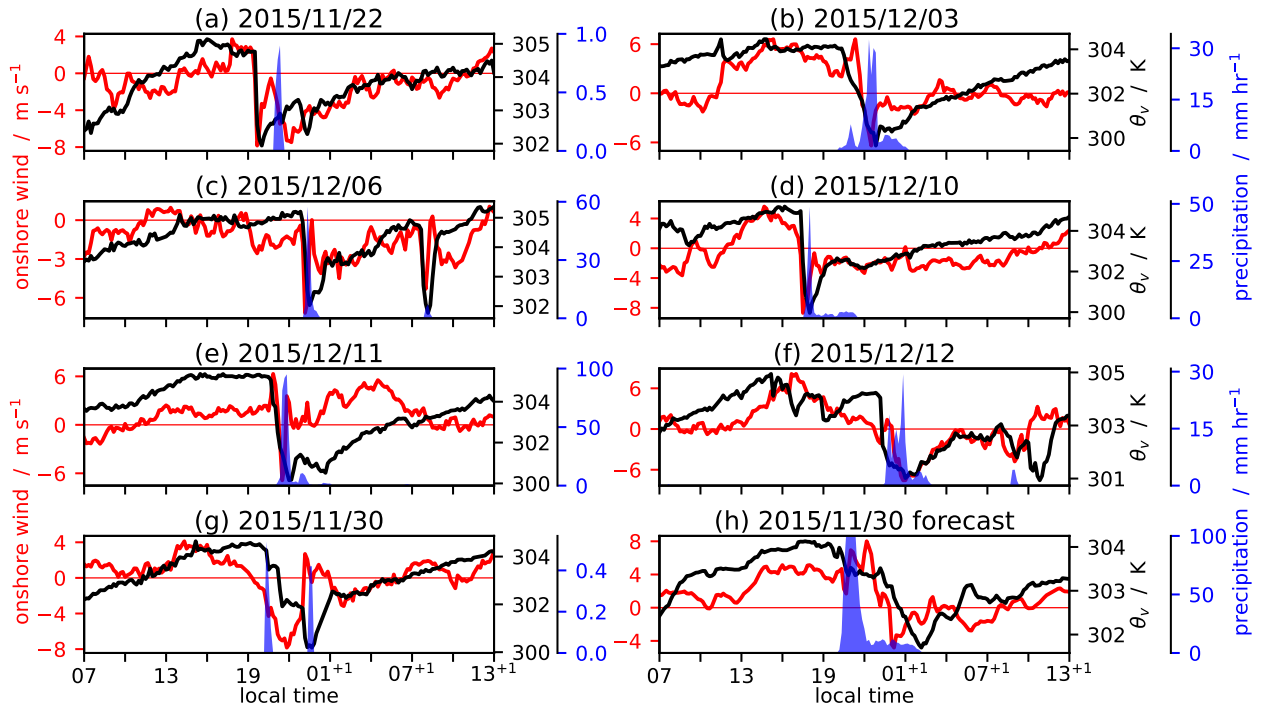


FIG. 13: (a–g) Time series from the *R/V Mirai* AWS on seven chosen dates. Curves are onshore wind (red) and θ_v (black), and filled blue region is precipitation rate. Local time is shown on the horizontal axis, with $^+$ indicating the day after the date given in the panel title. Note that (g) is out of chronological order, for ease of comparison with (h) the 2015/11/30 case study from the forecasts.

439 onshore wind, θ_v and precipitation rate, for seven chosen dates. Figure 13h is the equivalent for
 440 the 2015/11/30 forecast, interpolated to the same location as the *R/V Mirai*, and with wind and θ_v
 441 interpolated to the same levels as recorded by the AWS.

442 The observed cases in figures 13a–f all show a density current passing over the *R/V Mirai* in
 443 the evening or overnight. The onshore component of the wind switches from positive (onshore)
 444 to negative (offshore), accompanied by a sharp decrease in θ_v of 2–3 K, typically in around
 445 30–40 minutes. There are two candidates for these density currents – the land breeze and cold
 446 pools. It is very difficult to differentiate between the two, especially in observations, as they are in
 447 essence the same phenomenon, albeit with a different cause.

448 However, the land-sea temperature contrast driving an offshore-directed land breeze is well
 449 documented, so we expect the land breeze to be seen in these observations. At around 08 LT on
 450 2015/12/07 (figure 13c; i.e., 08 $^+$ LT for the 2015/12/06 case) there is a brief reduction in θ_v , along

451 with a shift to offshore winds and fairly light rain. This occurs around 9 hours after the probable
452 land breeze feature, and may be a cold pool front passing over.

453 These frontal features are always either accompanied by or shortly followed by rainfall, which
454 may be intense but does not fall over the *R/V Mirai* for long. This is consistent with the land breeze
455 being responsible for rainfall propagating away from the land.

456 The variability in rainfall intensity is considerable. For the six cases chosen, the peak rainfall
457 rate varies from around 0.9 mm hr^{-1} to around 95 mm hr^{-1} . However, precipitation rate can be
458 highly spatially variable, so the large range of values may arise from the fact that we are considering
459 precipitation at a single point only, so we are sampling from this highly variable field. Although
460 the land breeze-driven propagating convection tends to take the form of a squall line, it does not
461 necessarily follow that all precipitating parts of the system are contiguous.

462 For the 2015/11/30 case (figures 13g,h) we again see a reversal of the sign of the onshore
463 wind component and a reduction of θ_v , as expected from the vertical cross-sections in the model
464 (figure 7). The forecast (figure 13h) has a fairly gradual decline in θ_v (around 2.3 K in 6 hours),
465 which may be explained by the 2.2 km grid spacing not being fine enough to resolve the structure of
466 the front. However, the observations (figure 13g) also show a more gradual transition to lower- θ_v air
467 than is observed in other cases. The observed rainfall in this case is even lighter than in the lightest
468 of the other six case studies, peaking at around 0.5 mm hr^{-1} , while the forecast peaks at around
469 12 mm hr^{-1} during the time of the front. (The earlier peak, at around 21 LT, is due to an unrelated,
470 small but intense convective cell which appears before the land breeze propagation arrives.) As
471 explained above, the fact that we are dealing with point data may explain this discrepancy, so this
472 does not necessarily point to a bias in the forecast rainfall.

473 Also notable is the variability in the timing of the land breeze front. In the cases shown, this
474 varies from around 17 LT to around 01 LT. This is likely to be related to the environmental wind,
475 but an investigation of the causes of this variability is beyond the scope of this paper.

476 **4. Discussion and conclusions**

477 The offshore propagation of convection overnight from islands in the Maritime Continent is a
478 key feature of the diurnal cycle in the region. However, the existing literature (see section 1)
479 offers no strong consensus on the physical mechanisms behind this propagation. Understanding

480 the dynamics of intense precipitating storms is crucial for accurate forecasting and issuing early
481 warnings of high-impact weather.

482 The most commonly proposed mechanisms are the triggering of convection progressively offshore
483 due to low-level convergence between environmental onshore winds and offshore-propagating
484 density currents (either due to the land breeze or cold pools); or the triggering of precipitation by
485 gravity waves. Some studies suggest that gravity waves trigger convection directly; while others
486 suggest they destabilize the atmosphere ahead of the convection but a further trigger is still required
487 for the offshore convection to occur. A further contribution to the offshore propagation may arise
488 from signals propagating from the other side of the island (seen propagating from right to left in
489 the top-right corner of figure 2b), although these are beyond the scope of the present study, which
490 focuses on orographic and coastal processes. The diurnal cycle is often studied in composites
491 over many days, which may hide day-to-day variability in the timing, speed or direction of the
492 propagation; and may appear to show propagation through averaging over what are in fact stationary
493 or near-stationary precipitation features occurring on different days.

494 In this study we use a convection-permitting model and choose three case studies of offshore
495 propagation observed during the JAMSTEC pre-YMC field campaign in November and December
496 2015, to examine the physical processes involved. These cases all had a prevailing wind in the
497 offshore direction, and distinct offshore-propagating rainfall during the evening and overnight. The
498 model grid spacing was 2 km and diagnostics were output every 5 minutes, which is sufficient to
499 identify gravity waves and the gust front of a density current.

500 The results are summarized by the schematic diagram in figure 14. This shows a slice through
501 Sumatra, as in figure 7, with the horizontal axis being approximately south-west (on the left) to
502 north-east (on the right), the green peak being the Barisan mountains. The large grey arrow on the
503 right indicates the prevailing wind direction. Black arrows are density currents within the planetary
504 boundary layer; pink arrows represent gravity waves; and blue pointed lines are cold fronts, the
505 points indicating the direction of propagation. The colour of the sky indicates day (light blue),
506 evening (dark blue) or night (black). Note that the vertical heights, which are non-linear, and the
507 timings shown in this diagram are representative only, as they may vary from day to day.

508 Insolation warms the land surface during the day, causing an onshore sea breeze density current.
509 In reality the land surface warms and cools faster than the sea surface, meaning the land-sea

510 temperature contrast is warmer over the land during the day and warmer over the sea during the
511 night. In our model the SST is fixed during the day, but the warming and cooling of the land still
512 mean that the temperature contrast has the correct sign by day and by night. In a model with a
513 diurnal cycle of SST, there would likely be a second-order impact on the speed and intensity of the
514 land-sea breeze circulation and associated convection.

515 The sea breeze, whose front is indicated as a cold front on the diagram, propagates inland and
516 up the slopes of the Barisan mountains (figure 14a). There is a corresponding upslope wind on
517 the onshore side of the ridge causing convergence and ascent over the peaks; and some ascent due
518 to low-level heating from the ground. Clouds form over the mountains and begin to precipitate
519 (figure 14b).

520 The mountain-top convection deepens to form cumulonimbus storm clouds, typically by
521 early-to-mid evening (figure 14c). Since this occurs along the mountain range, the deep convection
522 is typically organized into a squall line, oriented along the direction of the mountains (which is
523 also the direction of the coast). The convection triggers gravity waves which may propagate in all
524 directions and their signals may be seen in the model at almost any altitude in the troposphere.
525 In the schematic diagram we summarize this by drawing a single wave, propagating offshore at
526 around 2 km altitude and with a phase speed of around 16 m s^{-1} , consistent with the $n = 3$ mode.

527 Once the land surface has cooled sufficiently to reverse the sign of the land-sea temperature
528 contrast (figure 14d), a downslope density current appears. At first, katabatic flow occurs down
529 the mountain, reinforced by the temperature contrast to form a land breeze density current which
530 propagates offshore. The squall line also causes cold pools – regions of cold air due to evaporation,
531 melting and sublimation of falling hydrometeors, which sink to due their high density and spread
532 out upon hitting the surface, again creating a density current. Although in general it is possible
533 to identify cold pools in the model (e.g., by plotting the gradient of low-level θ_v to identify their
534 associated fronts, as in figure 6), close to the coastline it is extremely difficult to disentangle these
535 cold pools from the sea breeze. Therefore, the cold front shown in figures 14d,e is assumed to
536 be the net result of the katabatic wind, land breeze and cold pools. In future work, we hope to
537 distinguish between these through the use of tracers in further convection-permitting model runs.

538 The motion of this offshore density current coincides with the squall line propagating offshore.
539 The convection and its precipitation are always collocated with the low-level convergence between

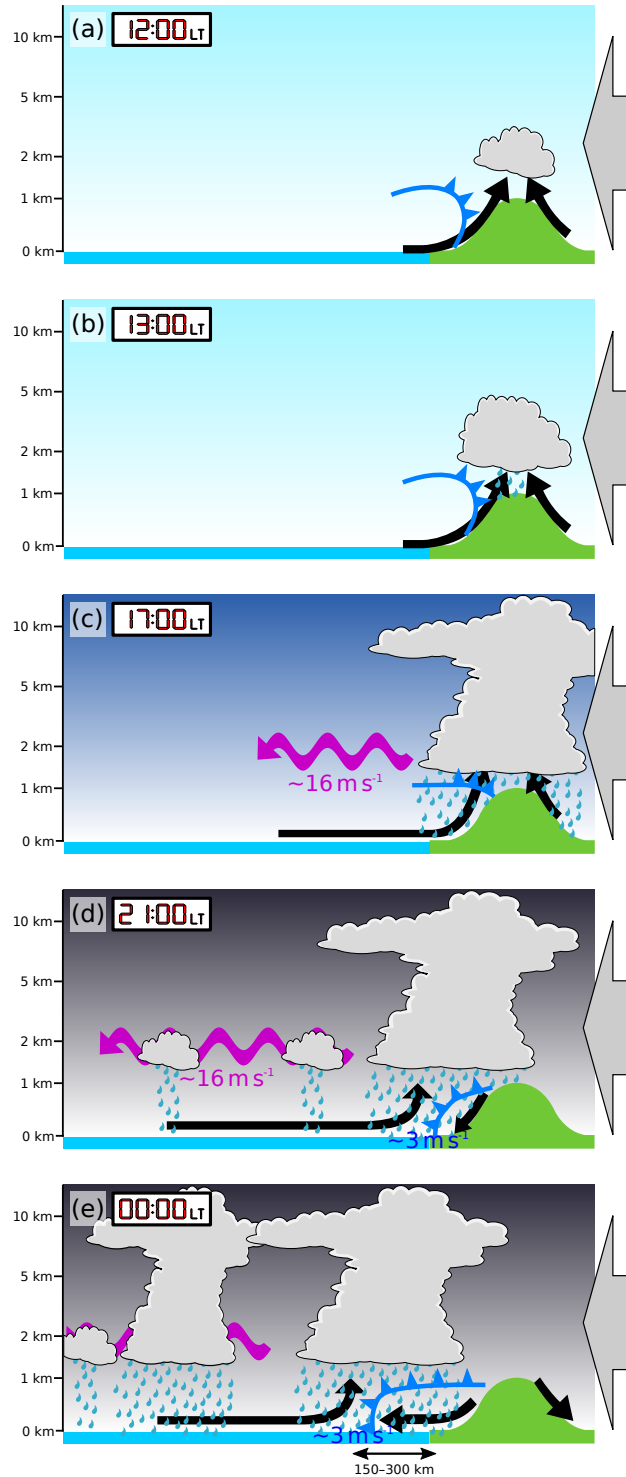


FIG. 14: Schematic diagram showing a day with an offshore prevailing wind and nocturnal offshore propagation of convection. The large grey arrow shows the prevailing wind direction and black arrows indicate density currents. Pink sinusoidal arrows represent gravity waves. Blue pointed lines are cold fronts, caused here by boundary layer density currents. The heights on the vertical axis, the horizontal length scale and the local times on the clock are representative only. The timing of the convection and its propagation may vary by several hours from day to day.

540 the density current and the remnant of the sea breeze from earlier in the day, out to around
541 150–300 km offshore. In the model runs, the propagation of the squall line is at a speed on the
542 order 3 m s^{-1} . This provides further evidence that the propagation cannot be directly caused by
543 gravity waves as the speed is far too slow. Moreover, observations from the AWS on the *R/V Mirai*
544 show a density current propagating through at the time of day indicated by the model, accompanied
545 by precipitation, verifying the mechanism outlined here. Hence, the physical processes involved
546 are more akin to the mechanism of Houze et al. (1981) than the gravity wave mechanism of Mapes
547 et al. (2003b,a; see section 1 for details), even though gravity wave mechanisms are often thought of
548 as responsible for the nocturnal offshore propagation in the Maritime Continent region (e.g., Love
549 et al. 2011; Yokoi et al. 2017; Coppin and Bellon 2019a). However, our results are consistent with
550 those studies which draw a distinction between slower propagation near to the coast, associated
551 with propagation of low-level convergence; and faster propagation further offshore, associated with
552 gravity waves (e.g., Vincent and Lane 2016; Bai et al. 2021; Fang and Du 2022).

553 Our results also show evidence of gravity waves playing a role in offshore convection. It
554 was not possible to establish any evidence for the waves destabilizing the atmosphere ahead of
555 the propagation, as hypothesized by studies such as Love et al. (2011) and Hassim et al. (2016).
556 Hovmöller diagrams of convective available potential energy, convective inhibition and moist static
557 energy (not shown) did not show any signal specifically associated with the waves. However, a
558 gravity wave may trigger convection which precipitates (e.g., Tulich and Mapes 2008), as indicated
559 in the left-hand side of figure 14d, but it is generally weak and localized. However, if there is already
560 an appreciable perturbation present, which the gravity wave interacts with, such as cloud caused
561 by an earlier cold pool, it is possible for the gravity wave to trigger deep and intense convection
562 there. This is indicated by the smaller cumulonimbus cloud in the left-hand side of figure 14e.
563 Such events have previously been studied in observations and models. Marsham and Parker (2006)
564 described a case study of a convective storm over southern England with three storms subsequently
565 triggered nearby, with gravity waves being the likely cause. Birch et al. (2013) showed that an
566 earlier version of the MetUM was able to reproduce observed secondary initiation of convection
567 over continental west Africa, due to gravity waves interacting with pre-existing dry convection and
568 an elevated boundary layer top.

569 Because the diurnal cycle over the Maritime Continent varies considerably from day to day, and
570 the offshore propagation is not always as distinct as in the case studies chosen here, it is common to
571 investigate the diurnal cycle in a composite sense (e.g., Peatman et al. 2014). As was demonstrated
572 in figure 2, this can give the impression of a coherent envelope of precipitation propagating offshore
573 over a range of speeds. The evidence of this study shows that the propagation at the slower speed
574 can indeed be coherent, due to the organization of the convection that propagates along with the
575 offshore density current. However, the gravity waves, which propagate at approximately the speed
576 of the faster edge of the composite envelope, mainly trigger isolated rainfall in our case studies.
577 Computing a composite may average over these isolated regions of rainfall, giving the appearance
578 of a propagating signal at the same phase speed as the wave, even though the individual precipitation
579 features causing it do not necessarily propagate at all, or may propagate slowly. This precipitation
580 happens much further offshore than the propagating organized squall line.

581 The distinction between the day-to-day propagation and the apparent propagation in a composite
582 is not necessarily important for climate studies, for example where we are interested in mean cloud
583 cover and albedo forcing. In this case, the average gravity wave interactions with pre-existing
584 convection are important in a climatological sense. However, in numerical weather prediction, and
585 specifically in quantitative precipitation forecasting, where it is necessary to forecast the location
586 and timing of convection on a particular day, the distinction is crucial.

587 The present study focuses on three case studies in which the prevailing wind is offshore, and the
588 propagation is both strong and coherent. This study furthers the existing literature by analyzing
589 individual cases at the process level, but the limited number of cases is insufficient to resolve
590 completely the lack of consensus in the literature regarding the underlying physical mechanisms of
591 the propagation. Further research will examine these mechanisms statistically over a much larger
592 number of cases. It should also be noted that there are comparable previous studies investigating
593 convection triggered by density currents and gravity waves over land (e.g., Birch et al. 2013) but,
594 in the case of the Maritime Continent, the convection's continued evolution is over ocean. The
595 presence of the ocean boundary layer likely means that the dynamics of the storm later in its lifetime
596 are quite different from those seen over land, and this is again a topic worthy of future investigation.

597 However, the present study demonstrates, in order to forecast the timing and location of
598 rainfall skilfully, the necessity of forecasting the propagation of the nocturnal land breeze and

599 the low-level convergence it generates; and understanding the more limited role of gravity waves
600 in the distribution of rainfall.

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616 Information Services Center (GES DISC; [https://pmm.nasa.gov/data-access/downloads/
617 gpm](https://pmm.nasa.gov/data-access/downloads/gpm)). GLOBE data are provided by the National Oceanic and Atmospheric Administration ([https://
618 www.ngdc.noaa.gov/mgg/topo/globe.html](https://www.ngdc.noaa.gov/mgg/topo/globe.html)). Radar and AWS data from the *R/V Mirai* are
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620 www.jamstec.go.jp/ymc/obs/pre-YMC/Mirai/index.html](https://www.jamstec.go.jp/ymc/obs/pre-YMC/Mirai/index.html)).

621 APPENDIX

622 **Identification of the direction of offshore propagation**

623 Here we detail the algorithm for diagnosing the direction of offshore propagation on an individual
624 day (if propagation can be identified). This algorithm was used to create figure 2b.

625 The gridded precipitation data over 36 hours, from 07 LT one day to 19 LT the next, are
626 interpolated to a rotated transect such as the one in figure 1a, trying every integer number of

627 degrees of orientation from 0° (propagation due west) to 90° (propagation due south). The mean is
628 computed over the width of the transect to produce a rotated Hovmöller diagram with dimensions
629 of onshore distance (x) and time (t). An example is shown in figure A1a, for 2015/11/21 and an
630 angle of 69° .

631 For each direction, the Hovmöller diagram is converted to a binary field, indicating whether
632 the precipitation rate exceeds a subjectively-chosen threshold of 1 mm hr^{-1} (figure A1b), and
633 contiguous features are identified. For propagation to exist, a precipitation feature must exist in
634 the centre of the x -direction of this field (the equivalent of 102.05°N , 3.55°S – on the coast near
635 Bengkulu) between 12 LT and 00 LT. If there are multiple such features, the only largest is retained
636 (figure A1c).

637 The speed of propagation is then found by linear regression through points where $x < 0$ (i.e., over
638 the sea) only. For each time from local noon until the feature disappears, all well-separated local
639 maxima in precipitation rate are found and the maximum closest to the coast is retained. Linear
640 regression is performed through the maxima that were chosen for each time, using the Theil-Sen
641 method (Sen 1968; Conover 1980), which chooses the median line through all possible pairs of
642 points. The gradient of this regression line gives us a speed associated with this transect angle
643 (figure A1c).

644 The angle which maximizes the speed is chosen as the direction of propagation. In general, not
645 all angles will give rise to a precipitation feature in the correct part of the Hovmöller diagram. If
646 there is no such feature at any angle, that day is deemed to have no propagation. By this measure,
647 propagation occurred on 1,254 days (used in the composite in figure 2b), or 33%, of the 3,806 days
648 in figure 2a.

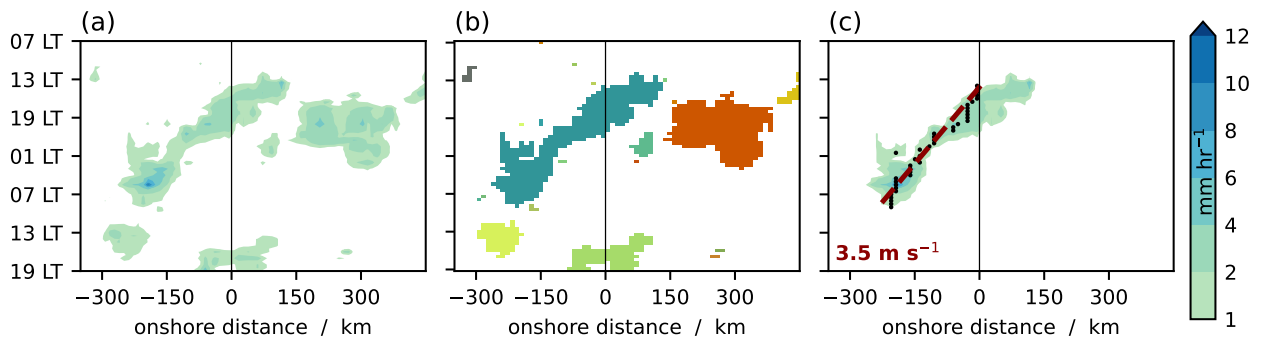


FIG. A1: Example of finding the offshore propagation speed for a Hovmöller diagram of precipitation rate at a given angle. (a) Hovmöller diagram of precipitation rate from IMERG for 2015/11/21 at an angle of 69° . (b) Contiguous features identified in a binary field of grid points where precipitation rate exceeds 1 mm hr^{-1} , coloured randomly. (c) Picking out only the contiguous feature which exists at the coast during local afternoon or evening (12 LT to 00 LT), a linear regression line is drawn (red dashed) through the local maximum at each time from 12 LT onwards. See appendix for full details.

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