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# Effects of unsteady mountain-gap winds on eddies in the Red Sea

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## Abstract

Atmosphere and ocean simulations capture the Tokar mountain-gap wind-jet and associated eddy dipole in the Red Sea. Results suggest that, once generated by a wind event, the anticyclonic eddy is further intensified through the summer by subsequent events. An asymmetry in magnitude of the two eddies is observed with the anticyclonic eddy attaining greater magnitude. A series of simulations with idealised wind forcing show that this asymmetry is not present under steady wind forcing. This effect depends on the intermittency of the wind forcing and the presence of a background northerly wind field.

## 1 Introduction

Mountain gap winds have been seen to induce oceanic eddy dipoles in the Gulfs of Tehuantepec, Papagayo and Panama (e.g. McCreary *et al.*, 1989; Liang *et al.*, 2009) and the Philippine Archipelago (e.g. Pullen *et al.*, 2008). The gap wind jets in the Gulf of Tehuantepec cause an eddy dipole which then propagates westward, away from the coast (Liang *et al.*, 2009). It was found (Chang *et al.*, 2012) that a subsequent wind jet can lead to an intensification of the eddy generated by the previous event. This will happen providing the interval between wind events is sufficiently small compared to the eddy propagation speed, so that the previous eddy is still in a region of enhanced wind stress curl.

A similar mountain gap wind jet and associated eddy dipole has been observed in the Red Sea (Zhai and Bower, 2013) in the region of the Tokar Gap (Figure 1a), which is located on the west side of the Red Sea, and is a 100km wide depression at around 300m elevation in the 1000m high mountains. They also noted that in the Red Sea, the eddy dipole generated by the Tokar Gap

26 jet will not be able to freely propagate zonally due to the eastern boundary of  
27 the basin. It therefore seems possible that these eddies might be subject to the  
28 same kind of intensification from successive wind jet events as seen in the Gulf  
29 of Tehuantepec (Chang *et al.*, 2012).

30 The atmospheric conditions in the region are strongly influenced by the  
31 local topography. Mountain chains run along much of the length of the Red  
32 Sea on both coastlines. This means that winds are generally constrained to  
33 flow along the axis of the Red Sea (Patzert, 1974). North-westerlies in the  
34 northern half of the basin are seen throughout the year, while in the south  
35 there is a distinct seasonality, with north-westerlies during summer and south-  
36 easterlies in winter (e.g. Pedgley, 1974; Patzert, 1974). It was noted that in  
37 winter the converging winds from the north and south funnel out of the Tokar  
38 Gap (Pedgley, 1974) while in summer winds blow through this gap onto the  
39 Red Sea (Hickey and Goudie, 2007) (see Figure 2 of Zhai and Bower, 2013 for  
40 characteristic wind fields in summer and winter). This summer wind jet through  
41 the Tokar Gap adds a cross-basin component to the wind field which would  
42 likely lead to enhanced eddy formation (Clifford *et al.*, 1997). This relationship  
43 was confirmed by (Zhai and Bower, 2013) who, using satellite altimetry and  
44 scatterometer data, found a correlation between the Tokar wind jet strength  
45 and that of an eddy dipole centered around 19°N. The anticyclonic eddy within  
46 this dipole was also observed by ship borne ADCP measurements (Sofianos and  
47 Johns, 2007).

48 The formation of the eddy dipole has not previously been investigated. Satel-  
49 lite sea-level anomaly measurements are based on 7 day resolution data, whereas  
50 the spin-up time of the eddies has been estimated at around a day. For this  
51 reason, and given the sparsity of in-situ measurements, we take a modelling  
52 approach to investigating the impact of the Tokar Gap Jet on eddy formation.  
53 In particular we focus on how the unsteady wind forcing can affect the eddies.

## 54 2 Data and Models

### 55 2.1 Data sets

56 By measuring sea surface height (SSH), altimetry data can detect mesoscale  
57 eddies from their sea level anomaly. Here we use merged sea level anomaly data  
58 from AVISO ([www.aviso.oceanobs.com](http://www.aviso.oceanobs.com)), which combines data from the Jason-  
59 1/2, Envisat, TOPEX/Poseidon and ERS satellites. These data are provided on  
60 a 0.33° Mercator spatial grid, and a 7-day temporal grid, and have previously  
61 been used (Zhai and Bower, 2013) to observe the eddy dipole near the Tokar  
62 Gap in 2001.

63 Our modelled wind fields are compared with observations from the Blended  
64 Sea Winds data set which provides daily wind speeds on a global 0.25° grid  
65 (Zhang *et al.*, 2006). This product combines data from up to six satellites,  
66 whose measurements are taken at different points throughout the day which  
67 reduces possible temporal sub-sampling errors. Combining these different data

68 sources also serves to reduce the amount of missing data, and there is >75%  
69 temporal coverage of every grid cell globally.

## 70 **2.2 Ocean Model**

71 In this study we use the Regional Ocean Modeling System (ROMS) (Haidvogel  
72 *et al.*, 2000; Shchepetkin and McWilliams, 2005). The model domain is rotated  
73 clockwise at  $35^\circ$  to reduce the number of land points and increase efficiency,  
74 while it encompasses the entire the Red Sea, and a portion of the Gulf of Aden  
75 which connects the Red Sea to the Indian Ocean. The model has 2 km horizontal  
76 resolution, so is able to resolve the mesoscale variability, and there are 20 vertical  
77 s-levels with increased resolution near the surface.

78 Initial conditions and lateral boundary forcing are taken from the Hybrid  
79 Coordinate Ocean Model (HYCOM) which is available daily at  $1/12^\circ$  resolution  
80 (Chassignet *et al.*, 2007). The only open boundary is in the Gulf of Aden, where  
81 the Flather boundary condition is imposed for 2d momentum and radiative  
82 nudging for 3d fields. While the Red Sea is subject only to weak tides, tidal  
83 forcing is also applied at this open boundary.

84 ROMS has previously been used by (Liang *et al.*, 2012) to study eddy forcing  
85 mechanisms in the northeastern tropical Pacific. In that study it was found  
86 through investigating different atmospheric forcing products that the model  
87 agreed best with observations when the temporal and spatial resolution of the  
88 surface boundary conditions was increased. In order to best resolve the Tokar  
89 Gap itself, we use a regional climate model to provide surface forcing to our  
90 ocean model. Atmospheric forcing of surface energy and water fluxes and wind  
91 stress is taken from simulations of the WRF model as described in Section 2.3.  
92 Three separate, realistic, ROMS simulations are performed, from 1st June to  
93 1st September of 2009, 2010 and 2011, the first month of each of these is used  
94 to spin-up the model before results are analysed for July and August.

95 A series of further simulations of ROMS is performed with idealised atmo-  
96 spheric forcing. Each of these has the same integrated wind forcing, but with  
97 varying degrees of intermittency. The three-day event around 3rd August 2010  
98 is selected as a time when a strong Tokar Jet forcing is present, but winds  
99 elsewhere are comparatively weak. A steady forcing experiment is then forced  
100 with the wind-stress fields from these 3 days scaled to half their strength and  
101 repeated throughout July and August. In one intermittancy experiment this  
102 3-day event is repeated every 6 days with 3 days of calm between events, and in  
103 two other experiments the event is scaled by 1.5 and repeated every 9 days and  
104 increased by a factor of 3 and repeated every 18 days. Each of these experiments  
105 is also forced with the shortwave radiation from 2010 and is spun-up for  
106 June with zero winds. To isolate the effect of wind stress forcing, spatially and  
107 temporally homogeneous net surface heat and water fluxes are applied.

## 108 2.3 Atmosphere Model

109 We use the Advanced Research core of the Weather Research and Forecasting  
110 (ARW-WRF) model (Skamarock *et al.*, 2008) to provide atmospheric forcing to  
111 our ocean model. WRF is a fully compressible, non-hydrostatic model which  
112 uses a terrain-following coordinate system. Our model is run on two one-way  
113 nested grids of 30 and 10 km resolution, with the inner grid centered on the  
114 Red Sea region and completely encompassing the ocean model grid. This 10  
115 km resolution is sufficient to accurately represent the Tokar Gap and the wind  
116 jet through it (e.g. Jiang *et al.*, 2009). Initial and boundary conditions are  
117 taken from the 6 hourly reanalysis fields of ERA-Interim (Dee *et al.*, 2011).  
118 The boundary conditions are applied over a buffer zone of five grid cells and  
119 there is no nudging within the interior of the grid. Three WRF simulations are  
120 performed for the summer seasons of 2009-2011, each running from 1st June to  
121 1st September. The output, including net surface heat and water fluxes, short  
122 wave radiation and wind-stress components is used to force the ocean model at  
123 3 hour intervals.

## 124 3 Results

125 The July and August mean wind speed field from the Sea Winds observations  
126 over the three years of simulation is shown in Figure 1b. The effect of the  
127 Tokar Gap is clearly seen in a band of strong winds emerging from the gap  
128 and then showing signs of being advected southward. This is compared to  
129 the WRF simulation results (Figure 1a) which also demonstrate strong winds  
130 in the Tokar Gap region curving southward with the prevailing along-basin  
131 winds. There is good qualitative agreement between simulated and observed  
132 wind speeds, with the areas of intense and calm winds showing good consistency.  
133 As expected, our mean wind fields are also in close agreement with those seen  
134 in the WRF simulations of (Jiang *et al.*, 2009) whose model compared well with  
135 in-situ measurements.

136 The model consistently shows stronger anticyclonic than cyclonic eddies.  
137 Figure 1d shows the mean SSH anomaly from July and August 2009-2011 in  
138 the AVISO satellite data. A number of eddies are apparent: there is a string of  
139 positive and negative SSH anomalies at around 24, 23, 22, 20 and 18°N, with  
140 the positive features clearer than the negative and the largest of these located at  
141 around 24°N. The eddies at 18 and 20°N were previously seen in other years of  
142 the AVISO data by (Zhai and Bower, 2013) and are associated with the Tokar  
143 Jet. The negative feature near 20°N curves southward around the positive  
144 anomaly just south of it, and is caused by the southward curving of the wind  
145 jet. These observations can be qualitatively compared with ROMS results in  
146 Figure 1c. In Figure 1c there is a very strong negative feature northward of  
147 25°N, which is not seen in the AVISO data. This cyclonic eddy was observed in  
148 (Sofianos and Johns, 2007) as a year-round feature, and so would not show up  
149 in Figure 1d which is an anomaly field. In ROMS the same string of positive

150 and negative SSH features between 21-24°N as AVISO is noted, although shifted  
151 slightly southwards in the simulations. The positive anomaly at 18-19°N caused  
152 by the Tokar Jet is very strong in ROMS, however the negative anomaly is not  
153 evident here. The cyclonic eddy can be seen in the individual years of data  
154 (not shown), however as it is weaker than the anticyclonic eddy and the eddy  
155 location varies temporally (seen in the next section) this is washed out in the  
156 3 year average SSH shown in Figure 1c. The eddies throughout the central  
157 and southern Red Sea compare well between AVISO observations and ROMS  
158 simulations, in particular the presence of an eddy caused by Tokar Jet forcing  
159 is noted in each.

160 We now examine the Tokar Jet strength and its relation with the ocean  
161 eddies. We define a box near the jet exit, as shown in Figure 1a, where we  
162 calculate the absolute magnitude of the wind stress curl at each point, and take  
163 a mean across the box. This metric is used as a measure of the strength of the  
164 Tokar Jet. In each year there are no jet events in June and they are concentrated  
165 in July and August (Figures 2a,c,e). Our model results also demonstrate the  
166 strong diurnal cycle, caused by the land-sea breeze, seen in the model results  
167 of (Jiang *et al.*, 2009) and the observations of (Zhai and Bower, 2013) although  
168 we remove this from the results shown in Figure 2.

169 By the end of August in each year there is an eddy dipole present, with  
170 a positive sea-surface height around 19°N and a negative height north of this  
171 (Figures 2b,d,f). There is however some variation in the location of the eddies,  
172 with those in 2010 further south than 2009, and a northward progression during  
173 August 2011. This shift in location is largely responsible for the missing negative  
174 SSH in the 3-year mean in Figure 1d as at 20°N there is a strong negative  
175 feature in 2009, but positive throughout 2010, and then positive in July 2011  
176 and negative in August 2011. In each year it is also clear that the eddy structure  
177 is not present in the initial conditions and throughout June. In 2009 (Figure  
178 2b) the eddies are initiated after the Tokar Jet event of 3rd July and then the  
179 magnitude of the anticyclone is increased by the event starting around July 16th  
180 while the cyclone intensity decreases, after which the magnitude of the dipole  
181 remains largely unchanged until the end of the simulation. During 2010 the  
182 eddies do not appear until later as there are no wind events until July 13th.  
183 After that point the anticyclonic eddy is intensified by a series of wind events  
184 on July 18th and 23rd, then another series of events starting on August 3rd  
185 regenerate the eddy slightly further south. In 2011 the eddy dipole emerges in a  
186 weak state around mid-June, although there is no obvious wind jet forcing event  
187 for this. In July however the anticyclonic eddy becomes much stronger, again  
188 driven by a series of Tokar Jet events particularly the ones on July 17th, 30th  
189 July and August 3rd. A clear tendency emerges throughout these simulations  
190 that the anticyclonic eddy tends to be stronger than the cyclonic eddy.

191 Figure 3 shows the cross-basin currents against depth. The anticyclonic eddy  
192 can be identified here as the positive across-basin velocities just north of 19°N  
193 and negative velocities south of this, while the cyclonic eddy involves the same  
194 positive velocities and the negative velocities around 20°N. The anticyclonic  
195 eddy is stronger than the cyclonic eddy as was apparent in Figure 2. The

196 core of both eddies is confined to the top 40 m of the water column, but the  
197 anticyclonic eddy in particular extends to between 200-250 m, which is deeper  
198 than the previous observation. At most other points during the simulations  
199 the depth of the anticyclonic eddy was between 100-150 m which agrees with  
200 the observation (Sofianos and Johns, 2007), but suggests that the eddy can get  
201 significantly deeper than this.

202 From the realistic model simulations, the overall picture is that there is an  
203 asymmetry in the eddy dipole, which is initially generated by Tokar Jet forcing  
204 and that once the anticyclonic eddy forms it is intensified by subsequent wind  
205 jet events. This is further investigated in a series of experiments with idealised  
206 atmospheric forcing (described in Section 2.2), where each simulation has the  
207 same integrated wind forcing, but with varying degrees of intermittency.

208 Along axis slices of the SSH for each of these idealised experiments are shown  
209 in Figure 4. In Figure 4a where the steady forcing is applied, the eddy dipole  
210 appears during early July, as expected, and continues to strengthen throughout  
211 the two months of forcing. In this case both the cyclonic and anticyclonic eddies  
212 seem to act similarly. When the forcing is doubled but applied only half of the  
213 time (Figure 4b), the eddies also begin to appear in early July and then increase  
214 in magnitude intermittently with each subsequent wind event. The magnitude  
215 of the positive SSH anomaly is however greater both than the negative anomaly  
216 and also than when the steady forcing is applied. The opposite is true of the  
217 cyclonic eddy, which seems not to further strengthen after the first month and  
218 which is therefore weaker than under steady forcing. There is a similar picture  
219 when the intermittency of the forcing is increased further (Figures 4c,d) of  
220 stronger positive SSH, and weaker negative SSH than with steady forcing. From  
221 Figure 4d it can clearly be seen that the cyclonic eddy is at its strongest at the  
222 time when the wind forcing is applied and weakens thereafter, whereas, the  
223 anticyclonic eddy weakens at the onset of forcing before strengthening when the  
224 forcing is removed. It is therefore clear that the intermittency of the wind jet  
225 forcing is the cause of the asymmetry in the eddy dipole response, and that this  
226 asymmetry comes both in timing and intensity of the eddies.

## 227 4 Discussion

228 By performing realistic atmospheric and oceanic simulations of the Red Sea  
229 during summer months we were able to replicate the observed mountain gap  
230 wind jet, and associated eddy dipole near the Tokar Gap. The mechanism for  
231 the formation of this eddy dipole is believed to be the Ekman pumping caused  
232 by a positive wind stress curl to the left of the wind jet, and negative curl to  
233 the right. This results in a cyclonic eddy with a negative sea level anomaly to  
234 the left of the jet, and an anticyclonic eddy and positive sea level anomaly to  
235 its right. Given the 3-dimensional nature of the simulations, we were able to  
236 find that while the anticyclonic eddy caused by the Tokar Jet generally extends  
237 to 100-150 m depth as in the only existing measurement (Sofianos and Johns,  
238 2007), the eddy can reach deeper than this, to around 250 m.

239 The eddies can be intensified by subsequent Tokar Jet events as seen by  
240 (Chang *et al.*, 2012) in the north-eastern tropical Pacific. It was also noted that  
241 there was an asymmetry in the eddy dipole, with a stronger anticyclone than  
242 cyclone. It was found that under steady wind jet forcing there is no asymmetry  
243 in the eddy behaviour. However under oscillatory forcing, the anticyclonic eddy  
244 is stronger than with steady forcing and the cyclonic eddy is weaker. There  
245 is also observed to be an asymmetry in the timing of the onset of the eddies,  
246 with the cyclone appearing immediately with the forcing, but the anticyclone  
247 appearing behind the wind jet.

248 What causes this asymmetric response between the cyclonic and anticyclonic  
249 eddies is less clear. For a wind jet without background flow, there will be no  
250 turning of the wind stress at a fixed ocean grid point (Figure 5a). However given  
251 that the climatology of the surface wind field in the Red Sea is for northerly  
252 winds, the picture is different. Before a jet event the winds are northerly, during  
253 the event they veer westerly, and then when the event dies out they return to  
254 northerlies. This means that at a fixed point in the path of the jet, winds veer  
255 cyclonically as the jet approaches and then anticyclonically as it passes (Figure  
256 5b). This leads to a directly wind-driven enhancement of the, Ekman forced,  
257 cyclonic eddy as the jet approaches, and an enhancement of the anticyclonic  
258 eddy behind it. This asymmetry in timing can explain the idealised simulations  
259 with unsteady forcing. If this mechanism is present then it would imply that  
260 the asymmetry is caused by the intermittency of the wind forcing, which would  
261 explain why it was not seen in the idealised simulations of (Zhai and Bower,  
262 2013) who applied steady Tokar Gap wind forcing.

## 263 5 Conclusions

264 Our realistic atmosphere and ocean simulations of the Red Sea captured the  
265 observed summer-time Tokar Gap wind-jet and associated eddy dipole. It was  
266 found that throughout the summer the magnitude of the anticyclonic eddy was  
267 increased by each wind event. It was also noted that the anticyclonic eddy  
268 was generally stronger than the cyclonic eddy. Experiments with idealised wind  
269 forcing confirmed this result when the forcing was intermittent. Under steady  
270 forcing the two eddies were symmetric. Another asymmetry between the two  
271 eddies was in timing: the cyclonic eddy appears at the onset of wind forcing  
272 and the anticyclonic eddy after it. It is reasoned that this effect is caused by a  
273 cyclonic veering of the wind stress from its background northerly direction to  
274 westerlies as the event begins, and then an anticyclonic veering back to norther-  
275 lies as the jet passes. This veering could enhance the respective eddies, and  
276 result in this asymmetry in timing.



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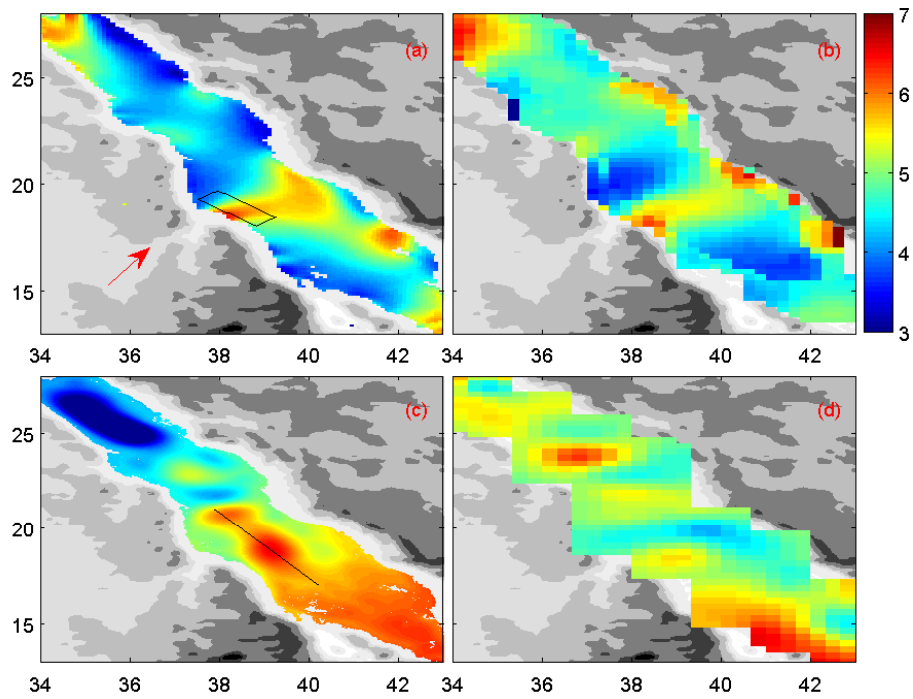


Figure 1: (top) Mean July and August wind speeds (m/s) for 2009-2011 from (a) WRF simulations, (b) Sea Winds satellite measurements. (bottom) July and August mean from 2009-2011 for (c) sea-surface height from ROMS simulations, (d) mean sea-level anomaly from AVISO data. Topography is shown in the gray scale, with shading intervals of <0m, 0-250, 250-500, 500-1000, 1000-2000, 2000-3000m.

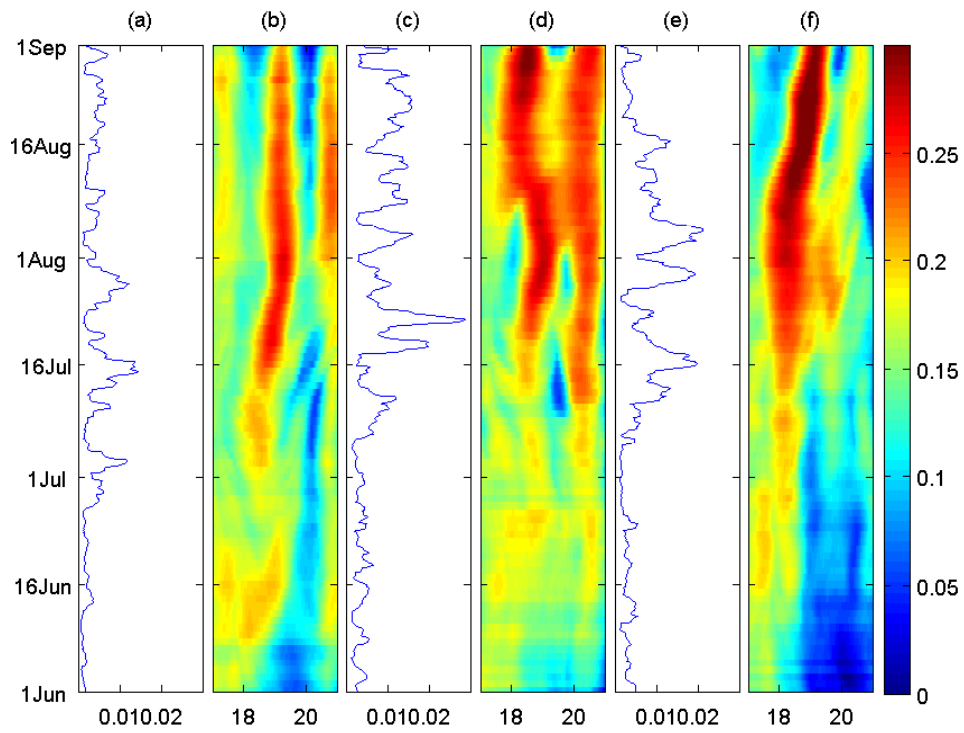


Figure 2: Mean absolute wind stress curl ( $N/m^3$ ) in the box indicated in Figure 1b for (a) 2009, (c) 2010, (e) 2011 from WRF simulations. Temporal evolution of sea-surface height (m), on an along axis cross-section, from ROMS simulations for (b) 2009, (d) 2010, (f) 2011 where the x-axis displays the latitude along the basin-section.

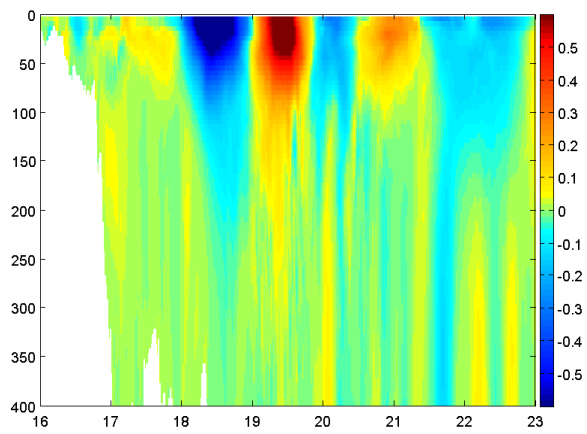


Figure 3: Along-axis cross-section of across-basin currents ( $m/s$ ) from 24th August 2011 plotted against depth (m), with latitude on the x-axis.

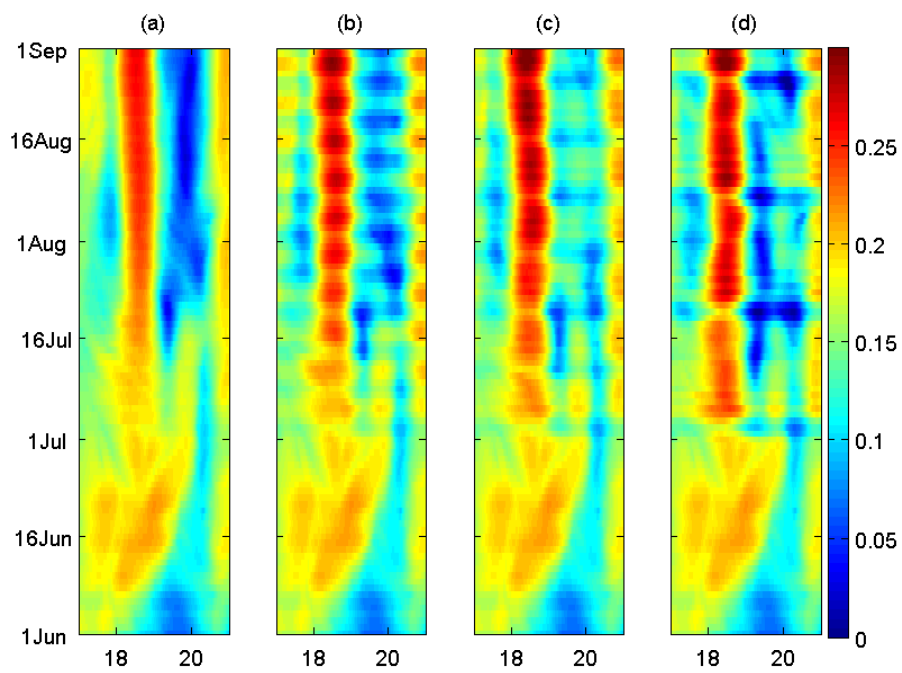


Figure 4: Temporal evolution of sea-surface height (m), on an along axis cross-section, from idealised ROMS simulations. (a) Steady Tokar Gap forcing, (b) double strength forcing applied 3 days on 3 days off, (c) 3 times the forcing applied 3 days on 6 days off, (d) 6 times forcing applied 3 days on 15 days off.

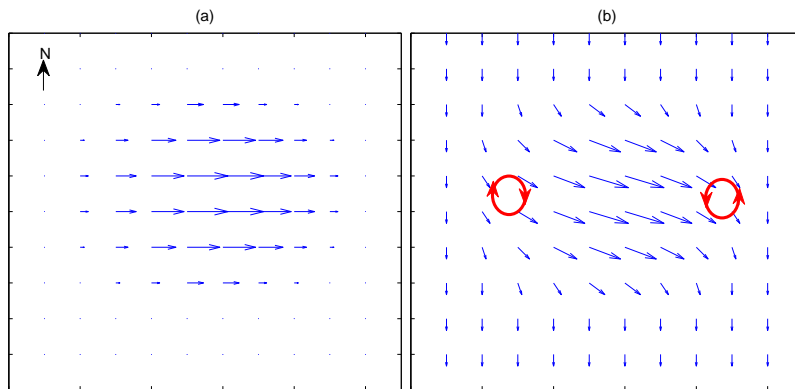


Figure 5: Schematic of (a) westerly wind-jet, (b) westerly wind-jet with background northerly wind field. Viewed from a fixed point on the sea, the wind stress veers cyclonically at the onset of the jet and anticyclonically behind it when there is this northerly background wind field, as indicated by the red arrows.