J.Natn.Sci.Foundation Sri Lanka 2022 50 (4): 851 - 861
DOI: http://dx.doi.org/10.4038/jnsfsr.v50i4.10636

RESEARCH ARTICLE

Physical oceanography

Anomalous propagation pathway of the Sri Lanka Dome in summer 2014

G Pathirana[1,2*], G Chen[2], D Wang[2], MK Abeyratne[3] and T Priyadarshana[4]

[1] *Department of Oceanography and Marine Geology, Faculty of Fisheries and Marine Sciences & Technology, University of Ruhuna, Wellamadama, Matara, Sri Lanka.*
[2] *South China Sea Institute of Oceanology, Guangzhou, China.*
[3] *Department of Mathematics, Faculty of Science, University of Ruhuna, Wellamadama, Matara, Sri Lanka.*
[4] *Department of Limnology and Water Technology, Faculty of Fisheries and Marine Sciences & Technology, University of Ruhuna, Wellamadama, Matara, Sri Lanka.*

Submitted: 28 July 2021; Revised: 17 July 2022; Accepted: 26 August 2022

Abstract: During the southwest monsoon, a cold dome referred to as the Sri Lanka Dome (SLD) appears east of Sri Lanka in response to the cyclonic curl in the local wind field. In general, SLD forms around May/June and decays during September. However, satellite altimetry data revealed an anomalous propagation pattern of the SLD with a lifetime of more than six months in the southwestern region of the Bay of Bengal (BoB) during 2014. This study is carried out to explain the plausible forcing factors which produced this anomalous event using satellite altimetry, surface currents, and wind data. Previous studies have shown that in general, the SLD propagates westward and then northward from its initial centre. In 2014, the SLD first appeared east of Sri Lanka around 6.5–7.5° N, 83–84° E and propagated eastward/northeastward from its initial centre (7.1018° N, 83.7111° E) with forcing from wind-stress and the Summer Monsoon Current (SMC). During early August it started to propagate westward until mid-October with the support of westward geostrophic flow and upwelling Rossby waves, noted between 10-12° N latitudes. Then the western boundary currents (WBC) in the BoB further facilitated the trajectory of the SLD and guided it towards the Arabian Sea (AS), where it disappeared in late December 2014, south of India (4.5474° N, 78.0223° E). Thus, we show that the stronger winds and SMC existed during 2014 in the BoB, guided the SLD eastward/northeastward compared to its general trajectory, and that allowed the SLD to undergo the forcing from westward geostrophic flow, upwelling Rossby waves and WBC.

Keywords: Rossby waves, Sri Lanka Dome, summer monsoon current, western boundary currents, wind-stress.

INTRODUCTION

The Bay of Bengal (BoB) is a unique oceanic basin due to seasonally reversing currents associated with the Asian Monsoon (Schott & McCreary, 2001; Jinadasa *et al*., 2020) and forced remotely by seasonal winds in the equatorial Indian ocean (Han & Webster, 2002), propagating waves (Poterma *et al*., 1991; Yu *et al*., 1991) and freshwater discharges (Harenduprakash & Mitra, 1988; Thadathil *et al*., 2002). The upper circulation in the BoB is characterized by a cyclonic flow during November and an anti-cyclonic gyre during most of the months (Somayajulu *et al*., 2003). Previous studies have revealed the presence of the eastward flowing Southwest Monsoon Current (SMC) during summer (June – September) and westward flowing Northeast Monsoon Current (NMC) during winter (December – February) in the BoB (Schott *et al*., 1994; Vinayachandran & Yamagata, 1998; Jensen, 2001). Vinayachandran *et al*. (1999) pointed out the intrusion of the SMC into the BoB with a mean seasonal transport during May – September. The East Indian Coastal Current (EICC), which flows along the western boundary of the BoB reverses its direction twice a year as a response to the local and remote alongshore winds (McCreary *et al*., 1993; Shankar *et al*., 1996). EICC flows poleward during February – September with a strong peak during March – April and starts to reverse its direction with the onset of

* Corresponding author (upgpathirana@fish.ruh.ac.lk; https://orcid.org/0000-0002-4670-579X)

post-summer monsoon (October) and flows equatorward until February with a strong peak in November (Shetye *et al*., 1993; Eigenheer & Quadfasel, 2000).

Mesoscale eddies, which could transport heat, salt, and nutrient-rich waters are an important phenomenon in the world's oceans (Dong *et al*., 2014). Energetic deep ocean eddies are capable of transporting water that wraps around it (Cherion & Brink, 2016). Previous studies were carried out on the existence of mesoscale eddies and their importance in the BoB based on observations (Cheng *et al*., 2013; Patnaik *et al*., 2014) and model simulations (Chen *et al*., 2012). Cui *et al*. (2016) showed that the mesoscale eddies are prominent in the western region of the BoB, and Chen *et al*. (2012) concluded that baroclinic instability at the interface of two opposing boundary currents supports eddy generation in the western BoB. The presence of the SMC, strong positive wind stress curl, and the Ekman pumping associated with the curl are considered as the dominant processes that contribute to the formation of cyclonic eddies (CEs) to the east of Sri Lanka (the southwestern region of the BoB) (Dandapat & Chakraborty, 2016). Based on theoretical and observational analyses, Dong *et al*. (2014) suggested that CEs and anticyclonic eddies (ACEs) propagate westward and move pole-ward and equatorward, owing to the β of the Earth's rotation (i.e. β represents the meridional gradient of the Coriolis parameter at a fixed latitude). In addition, Chen *et al*. (2012), showed that the eddy propagation south of 10° N in the BoB is westward at first and then northwestward with velocities of 2.0 – 11.0 cm/s.

The Sri Lanka Dome (SLD) is an upwelling dome that forms in a region of strong wind stress curl within the SMC system (Vinayachandran & Yamagata, 1998; Cullen & Shroyer, 2019). In general, the SLD appears in May, matures in July, and decays around September (Burns *et al*., 2017). The presence of the SMC, strong positive wind stress curl, and Ekman pumping associated with the curl are considered to be the dominant processes which contribute to the formation of the SLD east of Sri Lanka (Burns *et al*., 2017). In general, the cyclonic and anti-cyclonic eddies formed in the southwestern region of the BoB propagate westward from the initial centre. However, in 2014, satellite altimetry data revealed an anomalous propagation pattern of the SLD, with a lifetime of more than six months in the southwestern region of the BoB, east of Sri Lanka. Therefore, this study is carried out to explain the plausible forcing factors which produced this anomalous event using satellite altimetry, surface currents, and wind data.

MATERIALS AND METHODS

The anomalous trajectory of the SLD during 2014 and the forcing mechanisms are examined utilizing gridded sea level anomaly (SLA) data from Physical Oceanography Active Archive Data Centre (PODAAC), wind-stress data from the Advanced Scatterometer (ASCAT), ocean surface currents data from Ocean Surface Current Analysis Real-time (OSCAR) and drifter observations. The drifter observation was conducted under the ASIRI program by National Aquatic Resources Research and Development Agency (NARA) of Sri Lanka collaboratively with Scripps Institution of Oceanography (SIO), USA (Wijesekera *et al*., 2016). Details of the utilized datasets are given in Table 1.

Table 1: Summary of the datasets used in the study

Parameter	Source	Resolution	Time period
SLA (m)	*https://podaac.jpl.nasa.gov/*	1/6° × 1/6° (5-day)	1993 – 2018
ASCAT wind-stress (pa)	*http://apdrc.soest.hawaii.edu/*	0.25° × 0.25° (monthly)	2008 – 2017
OSCAR currents (m/s)	*http://apdrc.soest.hawaii.edu/*	1/3° × 1/3° (5-day)	2005 – 2015

Further, we estimate the geostrophic currents based on the equations given in Yu *et al*. (2014).

$$u = -\frac{g}{f}\frac{\partial h}{\partial y} \quad ...(1)$$

$$v = \frac{g}{f}\frac{\partial h}{\partial x} \quad ...(2)$$

In which, u and v are the zonal and meridional components of the geostrophic velocities, g is the gravitational constant, f is the Coriolis parameter ($f = 2\Omega sin\varphi$, where φ is the latitude) and h is the sea surface height.

Eddy detection and tracking procedure

Aside from the observed trajectory of the SLD through SSHA, we have estimated the trajectory of the SLD using the winding-angle method (Chen *et al*., 2012). Considering the SLD as a cyclonic eddy with a cold-water upwelling region, the winding-angle method is used as the eddy identification procedure in this study (Sadarjoen & Post, 2000; Chaigneau *et al*., 2008; Chen *et al*., 2012). OSCAR currents and SSHA data are interpolated to a resolution of $1/4° \times 1/4°$ to facilitate the analysis. Then an algorithm specifically developed for the BoB region by Chen *et al*. (2012) is used to track eddies in the region.

Initially, possible anti-cyclonic eddy (cyclonic eddy) centres were identified through searching for local SSHA maxima (minima) in a moving window of $4° \times 4°$ grid points (Chen *et al*., 2012; Dong *et al*., 2014). Then streamlines were computed for each possible eddy centre, following the trajectories of virtual particles release in the geostrophic current field. The closed streamlines are then selected by computing the winding angle, and clustered into different eddies. The detailed description of the eddy tracking procedure is explained by Chen *et al*. (2012). Then using the detected eddy information, eddy centre location, time, eddy polarity, eddy size, and eddy movement velocity are calculated. The eddy kinetic energy (EKE) was computed following Chen *et al*. (2012) (equation 3).

$$EKE = \frac{1}{2}\left(u'^2 + v'^2\right) \qquad ...(3)$$

RESULTS AND DISCUSSION

SLD climatology

The monthly climatology of sea level anomalies (SLA) is presented in Figure 1. In agreement with previous studies (Vinayachandran & Yamagata, 1998), a negative SLA appears east of Sri Lanka in June, which has been referred as the SLD. The mechanism of formation and propagation of the SLD has been verified by Burns *et al*. (2017), and their study confirms the importance of strong cyclonic curl in wind stress which starts in May and continues up to October, facilitating the SLD. The SLD decays during September in response to the downwelling Rossby waves that reach the SLD region (Vinayachandran & Yamagata, 1998). The westward propagating Rossby waves push the thermocline down and prevents cold water upwelling to the surface, and act as a negative feedback for sustaining SLD during September (Vinayachandran & Yamagata, 1998). Even though the mechanism for the formation and decay of the SLD is well-known (Burns *et al*., 2017), the anomalous propagation pathway of the SLD identified in 2014 raised the importance of understanding the responsible forcing factors.

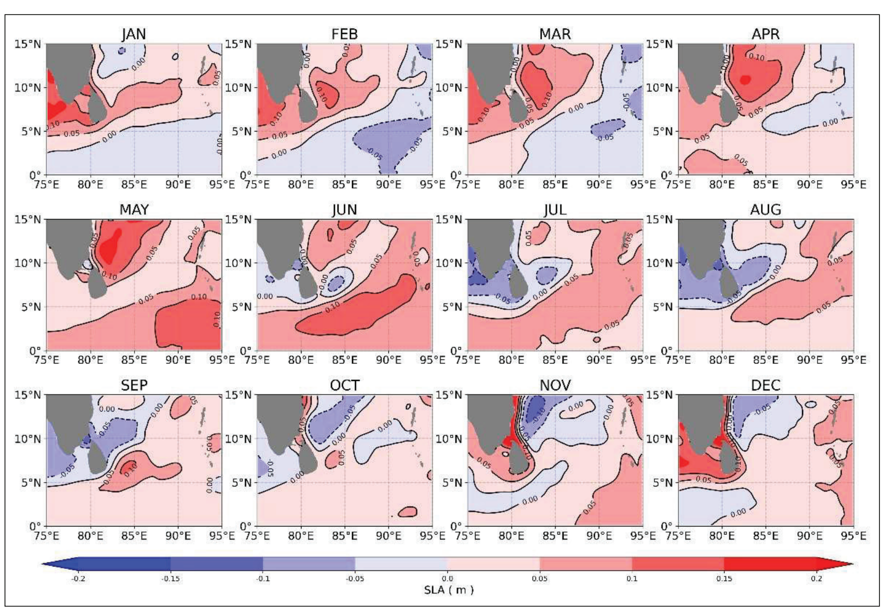

Figure 1: Monthly climatology of sea level anomalies in the southwestern Bay of Bengal

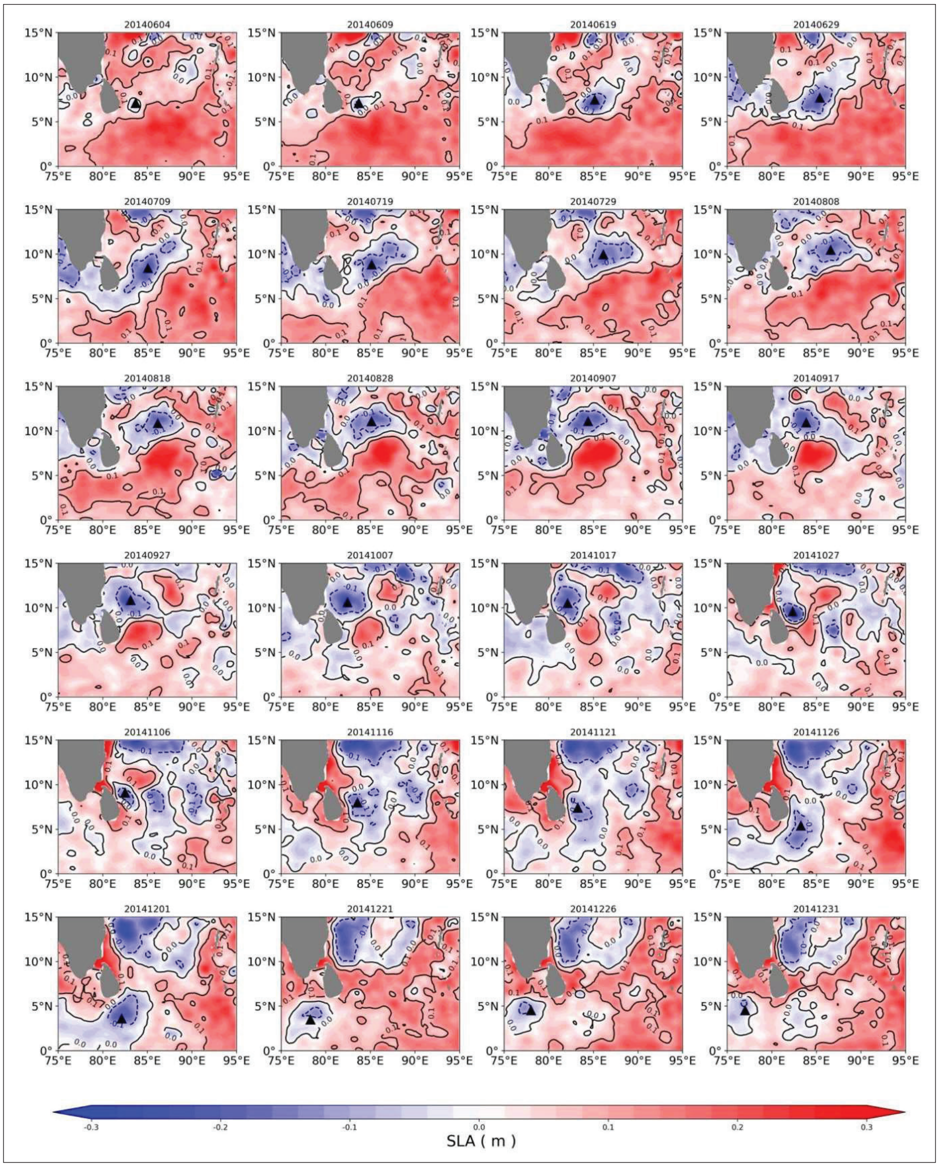

Figure 2: Snapshots of sea level anomalies from June 2014 to December 2014 in the southwestern Bay of Bengal. The centre of the anomalous Sri Lanka Dome (SLD) (tracked using winding-angle method) is marked with a black triangle in each time-step.

Trajectory of SLD during 2014

Time series (1995 - 2015) of SLAs have been examined to identify the propagation pathways of CEs and ACEs in the BoB region. The presence of a negative sea level anomaly east of Sri Lanka, referred to as the SLD, was observed during 2014 with an exceptional lifetime of nearly six months. This observed anomalous feature during 2014 follows a different propagation trajectory compared to the general pattern described in earlier studies (Burns *et al*., 2017). The snapshots of SLAs from June to December in 2014 is presented in Figure 2 to illustrates the propagation pathway of the SLD. Using the winding-angle method with SLA and OSCAR currents, the anomalous propagation of the SLD has been tracked and the results are presented in Figure 3.

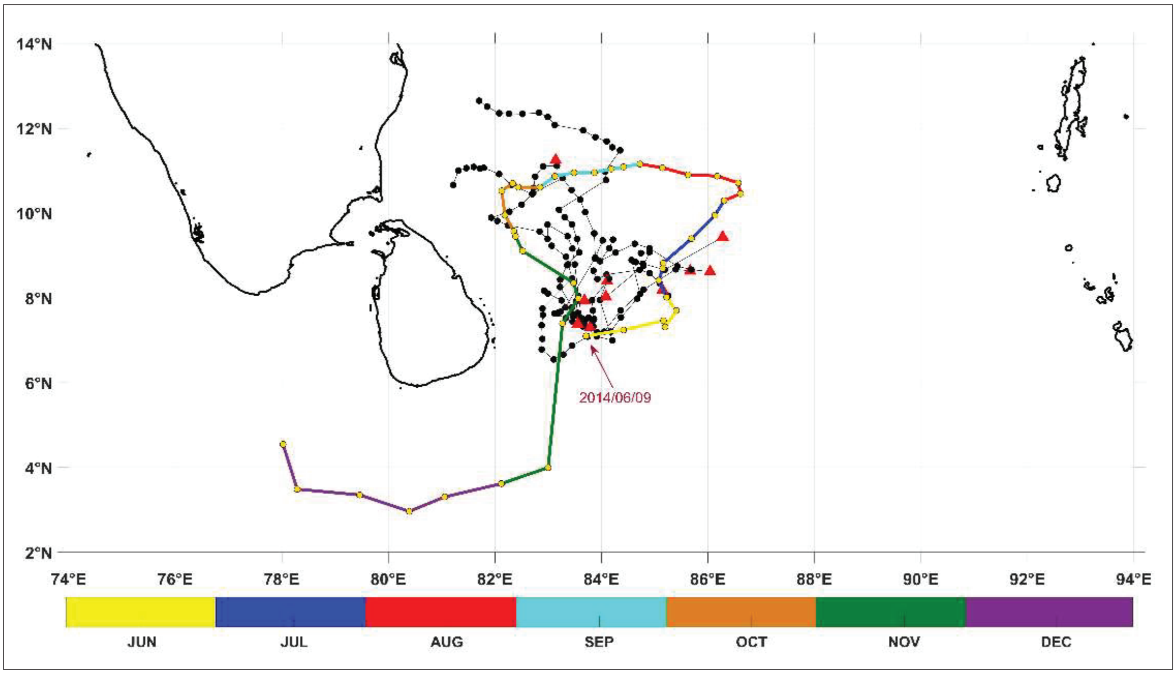

Figure 3: Anomalous propagation pathway of the Sri Lanka Dome (SLD) in 2014. Yellow circles illustrate the anomalous trajectory of the SLD observed in 2014, and the associated colour lines indicate the calendar months. The arrow denotes the initial centre of the anomalous SLD. Black circles indicate the general trajectories of the SLD in selected years. The red triangles denote the initial centre of the SLD in each year.

Based on the SLA observations, the SLD first appeared east of Sri Lanka around 6.5-7.5° N, 83-84° E in early June and propagated eastward from its initial centre (7.1018° N, 83.7111° E) until the end of June. Then it turned north/northeastward and continued until the end of July. During early August it started to propagate westward until mid-October. From mid-October, it propagated equatorward along the eastern and southern coast of Sri Lanka and disappeared in the Arabian Sea in late December 2014, south of India (4.5474° N, 78.0223° E). Thus, the data revealed the existence of the SLD with a lifetime of more than six months during 2014, and its anomalous propagation pathway. As the eddies which form in the southwestern region of the BoB generally propagate westward, the observed anomalous feature highlighted the importance of influence from other forcing factors in the BoB on eddy formation and propagation. Therefore, to decipher the anomalous propagation of the SLD and the forcing mechanisms, monthly wind stress and curl, surface currents, geostrophic currents, Rossby wave forcing, and kinetic energy has been examined from June – December 2014 in the region.

Forcing mechanism

Wind-stress and SMC

With the onset of the summer monsoon (June), wind stress over the BoB is directed northeastward and it is sustained until the end of the summer monsoon (August). Relaxation of winds over the bay is identical during the post-summer monsoon (October – November). A strong positive wind stress curl is a prominent feature which appears east of Sri Lanka and points out the Ekman divergence in the region. The presence of positive wind stress curl associated with Ekman divergence supports the formation of cyclonic eddies in the region (Burns *et al*., 2017). Relatively strong surface wind stress directed northeastward is observed during May-June in 2014, which weakens with the progress of the season (Figure 4). The SMC triggered by the seasonal winds starts to flow into the BoB from the Arabian Sea south of Sri Lanka with the onset of the summer monsoon (Figure 4). The SMC strengthens with the progression of the summer monsoon and weakens during the post-summer period. Results revealed that the SMC was stronger during July 2014 and occupied a larger area in the southwestern BoB.

As the strong wind-stress curl and SMC are well-known initial forcing factors for the SLD (Cullen & Shroyer, 2019), we show the monthly anomalies of wind-stress (Figure 5a) and surface currents (Figure 5b) to explain the initial forcing that facilitated the SLD in 2014. The results indicate the presence of an eastward wind stress and surface currents in the southern BoB during June 2014, which has further strengthened towards the east/northeast during July 2014. Based on the quasi-geostrophic theory, Chen *et al*. (2012) reported that eddies on a β-plane should propagate westward. However, the anomalous trajectory of the SLD, observed in 2014, indicated an eastward/northeastward propagation. Therefore, here we suggest that the existence of weaker

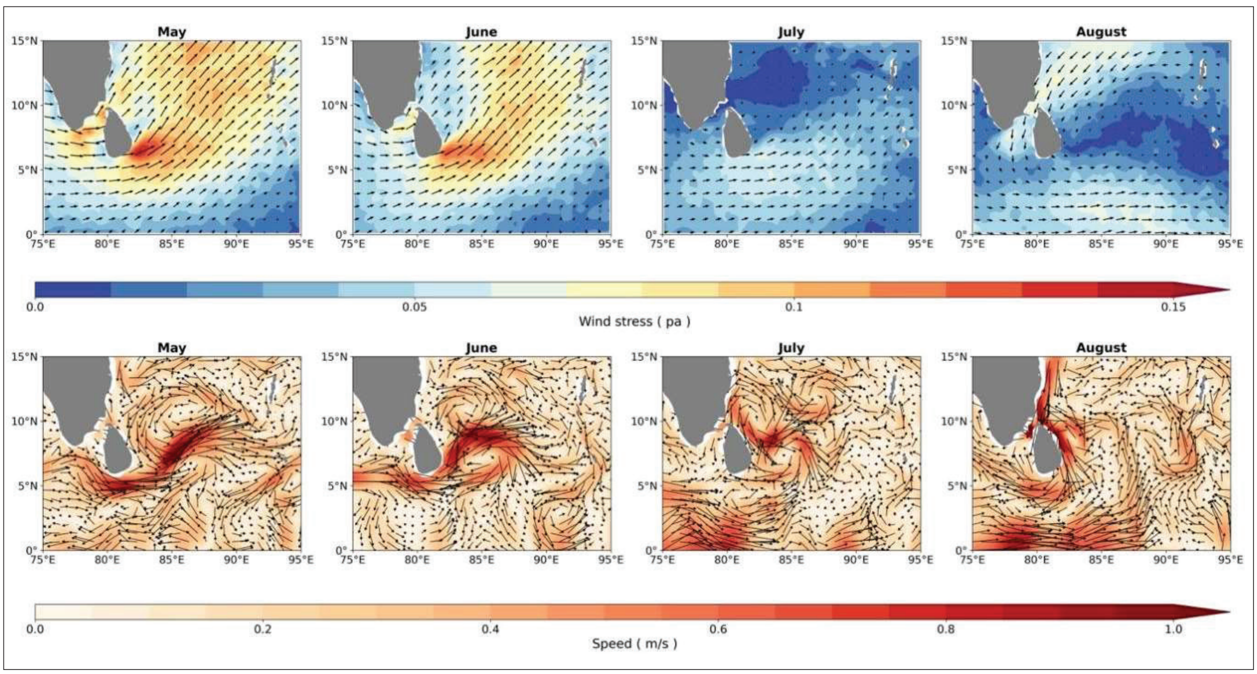

Figure 4: Monthly average wind-stress (upper panel) and surface currents (lower panel) in the Bay of Bengal during May – August 2014.

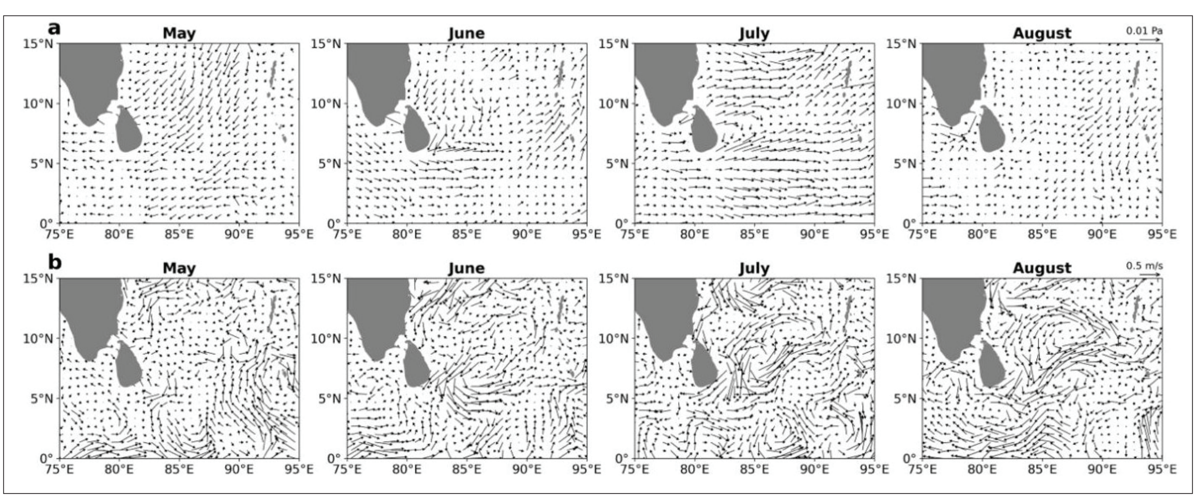

Figure 5: Anomalies of wind-stress (a), and surface currents (b) in the Bay of Bengal during May–August 2014.

upwelling Rossby waves in the 6 – 8° N latitudinal band (Figure 6a) and the presence of strong eastward winds and SMC during June – July 2014 may have facilitated the eastward/northeastward propagation of the SLD. The eastward/northeastward propagation may have allowed the SLD to undergo the forcing of upwelling Rossby waves and the western boundary currents during August to December in 2014.

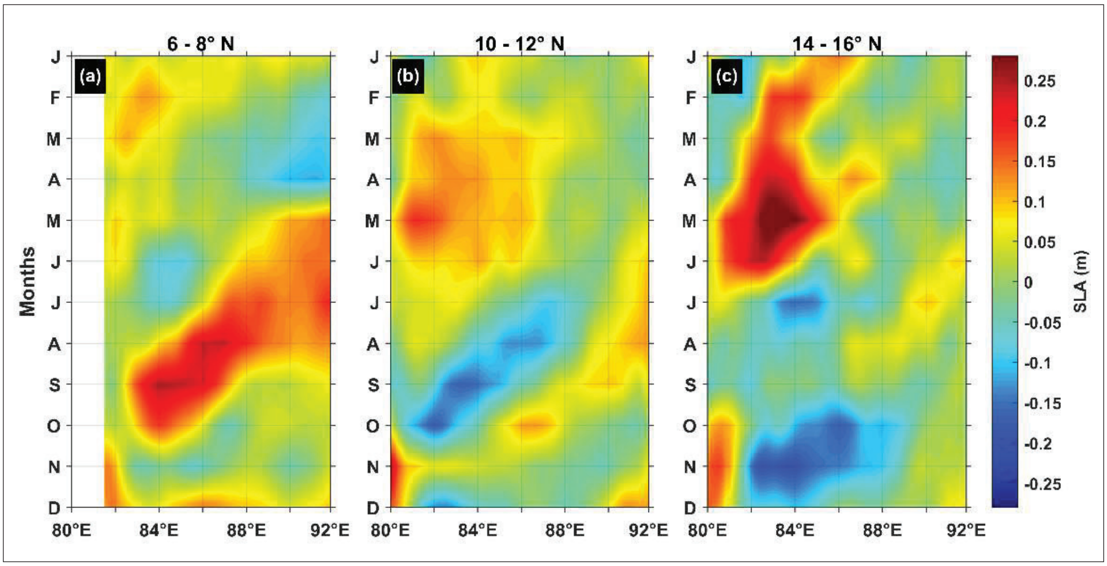

Figure 6: Hövmoller diagram (time versus longitude) of monthly average sea level anomalies (a) 6-8° N, 80-92° E; (b) 10-12° N, 80-92° E, and (c) 14-16° N, 80-92° E, during 2014.

Rossby waves

The Kelvin and Rossby waves are important dynamic features of the ocean, as they are capable of transferring wind energy in the basin due to their non-dispersive character (Kumar *et al*., 2000; Rao *et al*., 2013). Chelton *et al*. (2007) reported that in the oceans, westward propagating eddies and Rossby waves are intrinsically related to each other and sometimes it is difficult to differentiate these two phenomena. A Hövmoller diagram for the BoB at three different latitudinal bands (6-8° N, 10-12° N, 14-16° N) shows the westward propagation of positive and negative of SLAs during the study period implying the existence of westward propagating Rossby waves (Figure 6). The existence of downwelling Rossby waves is thought to inhibit the CEs, while upwelling Rossby waves facilitate the CEs in the ocean. However, the observed Rossby waves were weaker in the 6–8° N latitudinal band and hence had less influence on the propagation of the SLD in 2014 (Figure 6a). Further, a negative SLA signal was observed between 10–12° N latitudes during June – October 2014 which propagated westward in the BoB (Figure 6b). This is in agreement with earlier findings (Kumar *et al*., 2000), that reported the movement of low SLAs from the eastern boundary in March and ended up in the western boundary during October, along the 12° N latitude in the BoB. Therefore, the upper ocean variability of the BoB, to a larger extent is controlled by the westward propagating Rossby waves. Also, Chen *et al*. (2012) have pointed out that, in the central BoB (10° N–15° N) eddies tend to move quasiwestward with velocities of 2.0-11.0 cm/s. Thus, in agreement with the earlier findings, the SLD after crossing the 10° N latitude turned and propagated westward until mid-October with the forcing of upwelling Rossby waves. Further, it has been identified that the northeastward propagating SLD merged with another westward propagating CE around 10° N latitude in mid-July 2014, and the combination propagated westward until mid-October (Figure 2). Generally, the SLD does not undergo forcing from westward propagating waves during the late summer monsoon, as it decays at the western boundary of the BoB. However, the initial eastward propagation has allowed the SLD to undergo the forcing of upwelling Rossby waves in the region.

Western boundary currents (WBC)

With the onset of the post-summer monsoon, winds over the BoB weaken and reverse direction towards southwest. The seasonally reversing East Indian Coastal Current (EICC) starts to flow equatorward along the western boundary of the BoB as a response to the seasonally reversing winds during October. The seasonal changes in the western boundary currents of the BoB are evident in the SVP drifter observations (Jinadasa *et al*., 2020) and they are presented in Figure 7. During mid-October, the westward propagating SLD interacts with the equatorward flowing EICC at the northeast of Sri Lanka (Figure 2) and starts to flow equatorward (Figure 3). The SLD is located about 100 km away from the east coast of Sri Lanka during late October, it interacts with the WBC of the BoB, and it is evident in the surface current observations (Figure 7). In general, the SLD does not prevail till post-summer monsoon to undergo the forcing of the WBC. Thus, with the forcing of WBC the SLD propagated along the east coast of Sri Lanka towards AS during late October – December, in 2014.

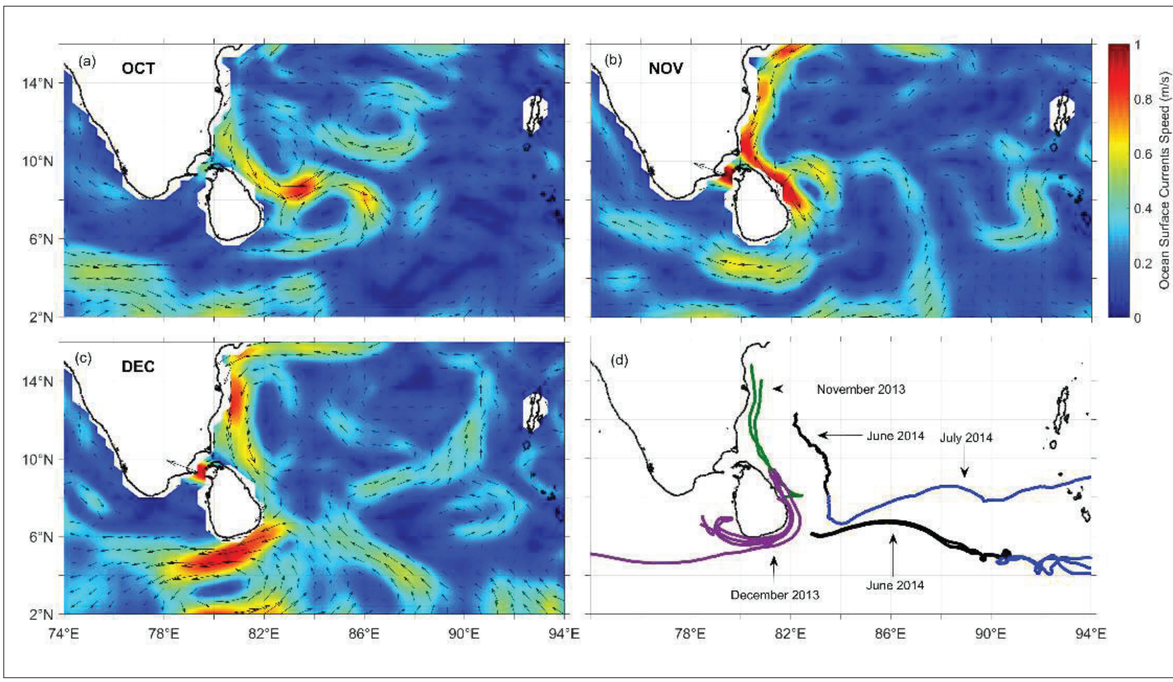

Figure 7: Surface currents in the southwestern Bay of Bengal, (a) October, (b) November, and (c) December 2014. And (d) surface currents estimated from SVP drifter trajectories in the southwestern region of the Bay of Bengal during 2013-2014.

Variability of eddy properties

The variability of eddy kinetic energy (EKE), eddy radius, and eddy velocity of the anomalous SLD have been examined and the results are presented in Figure 8. From mid-June to mid-October 2014 EKE remained in between 0 – 0.05 cm^2s^{-2} and peaked to a maximum of 0.25 cm^2s^{-2} in early November. From late-October to mid-December the EKE was comparatively high and varied between 0.05 and 0.25 cm^2s^{-2}. Further, the calculated eddy velocity along the trajectory clearly indicates that it was increasing during November – December and peaked to a maximum of 8.76 cm/s in end of November 2014 (Figure 8). The identified peaks in EKE and eddy velocity during late-October to mid-December clearly emphasize the influence of the WBC on the anomalous SLD. Rao *et al*. (2013), pointed out that the kinetic energy is higher along the coastally trapped waves during December-January, and in January the energy is much higher (10,000 cm^2 s^{-2}) in the western BoB. Using 17 years (Oct. 1992 to Oct. 2009) of altimetry data, Chen *et al*. (2012) have explained the mean eddy properties and their temporal and spatial variations in the BoB. Further, they pointed out that the total eddy energy observed at the western boundary is higher than in the east, while the baroclinic and barotropic conversions are both active at the western boundary but the former is stronger. The growth of the perturbation through baroclinic instability is more important to the formation and existence of eddies in the west, hence strong EICC should be the key factor. This could be the reason that the detected SLD has a higher EKE and velocity after reaching the western boundary of the BoB. Further it highlights that the forcing from the boundary current on the SLD is higher compared to that of forcing from westward propagating Rossby waves and seasonal winds. The eddy radius varies in between about 75 and 400 km and the maximum radius, ~ 375 km, is observed during mid-August. Merging with another eddy during this period and the weakening of external forcing may be one of the reasons for the noted increase in the radius (Figure 2). Theoretically when an eddy is more energetic, the radius remains small. It is evident in Figure 8, and it clearly indicates that, during November, the radius of the anomalous CE is reducing when the EKE reaches a maximum.

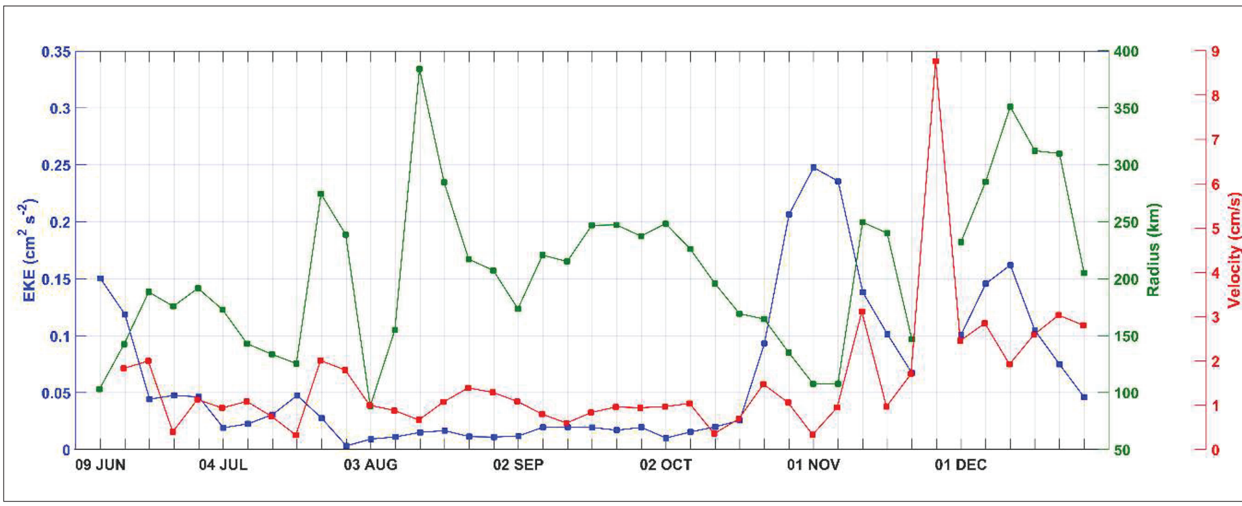

Figure 8: Variability of EKE, eddy radius and velocity along the trajectory of the Sri Lanka Dome (SLD) during 2014.

CONCLUSION

The SLD is a feature that appears off the east of Sri Lanka during May-September. The formation of the SLD seems to be linked with the strong cyclonic curl in wind stress and SMC, while the decay is linked with the westward propagating downwelling Rossby waves in the region. However, an anomalous propagation pathway of the SLD has been identified during 2014, and this study is carried out to understand the influencing factors behind this anomalous propagation. The trajectory, radius, EKE, and velocity of the SLD are identified using the winding-angle method with SLA and OSCAR currents. The SLD propagated eastward/northeastward from its initial centre with the forcing of strong wind and SMC which dominates over the westward propagating Rossby waves during June – July 2014. After crossing the 10° N latitude, the SLD tended to propagate westward until mid-October, with the forcing of Rossby waves. Finally, the SLD was carried towards the AS along the east coast of Sri Lanka by the WBC that exists in the BoB. The detected SLD existed for a longer time in the ocean while propagating over longer distances compared to its behaviour in the previous years. The initial eastward shift facilitated by the SMC may have supported the observed anomalous trajectory of the SLD. As the processes involved in SLD are conceived to influence the formation and frequency of monsoon depressions in the northern BoB (Burns *et al*., 2017), further studies are recommended to understand inter-annual variability of the dome dynamics.

Acknowledgements

This work is supported by the National Key R&D program of China (2017YFC1405103), the Major

State Research Development Program of China (2016YFC1402603), XDA 20060502, NSFC (grants 41822602 and 41521005), XDA (grant 11010102), Guangzhou Science and Technology Foundation (201804010133), State Key Laboratory of Tropical Oceanography SCSIO (LTOZZ1702), and the Youth Innovation Promotion Association of CAS (2017397). The encouragement and the facilities provided by the China-Sri Lanka Joint Centre for Education and Research (CSL-CER) are gratefully acknowledged.

Data availability

All the information related to the data sources used in this study can be accessed using the links provided in the text.

REFERENCES

Burns J.M., Subrahmanyam B. & Murty V.S.N. (2017). On the dynamics of the Sri Lanka Dome in the Bay of Bengal. *Journal of Geophysical Research: Oceans* **122**: 7737–7750. DOI: https://doi.org/10.1002/2017JC012986

Chaigneau A., Gizolme A. & Grados C. (2008). Mesoscale eddies off Peru in altimeter records: identification algorithms and eddy spatio-temporal patterns. *Progress in Oceanography* **79**(2–4): 106–119. DOI: https://doi.org/10.1016/j.pocean.2008.10.013

Chelton D.B., Schlax M.G., Samelson R.M. & De Szoeke R.A. (2007). Global observations of large oceanic eddies. *Geophysical Research Letters* **34**. DOI: https://doi.org/10.1029/2007GL030812

Chen G., Wang D. & Hou Y. (2012). The features and interannual variability mechanism of mesoscale eddies in the Bay of Bengal. *Continental Shelf Research* **47**: 178–185. DOI: https://doi.org/10.1016/j.csr.2012.07.011

Cheng X., Xie S.-P., McCreary J.P., Qi Y. & Du Y. (2013). Intraseasonal variability of sea surface height in the Bay of Bengal. *Journal of Geophysical Research* **118**: 816–830. DOI: https://doi.org/10.1002/jgrc.20075

Cherion D.A. & Brink K.H. (2016). Offshore transport of shelf water by deep-ocean eddies. *Journal of Physical Oceanography* **46**: 3599–3621. DOI: https://doi.org/10.1175/JPO-D-16-0085.1

Cui W., Yang J. & Ma Y. (2016). A statistical analysis of mesoscale eddies in the Bay of Bengal from 22–year altimetry data. *Acta Oceanologica Sinica* **35**(11): 16–27. DOI: https://doi.org/10.1007/s13131-016-0945-3

Cullen K.E. & Shroyer E.L. (2019). Seasonality and interannual variability of the Sri Lanka dome. *Deep Sea Research Part II* **168**: 104642. DOI: https://doi.org/10.1016/j.dsr2.2019.104642

Dandapat S. & Chakraborty A. (2016). Mesoscale Eddies in the Western Bay of Bengal as observed from satellite altimetry in 1993–2014: statistical characteristics, variability and three-dimensional properties. *IEEE Journal of Selected Topics in Applied Earth Observations and Remote Sensing* **9**(11): 5044–5054. DOI: https://doi.org/10.1109/JSTARS.2016.2585179

Dong C., McWilliams J.C., Liu Y. & Chen D. (2014). Global heat and salt transports by eddy movement. *Nature Communications* **5**. DOI: https://doi.org/10.1038/ncomms4294

Eigenheer A. & Quadfasel D. (2000). Seasonal variability of the Bay of Bengal circulation inferred from TOPEX/ Poseidon altimetry. *Journal of Geophysical Research* **105**(C2): 3243–3252. DOI: https://doi.org/10.1029/1999JC900921

Han W. & Webster P.J. (2002). Forcing mechanisms of sea level interannual variability in the Bay of Bengal. *Journal of Physical Oceanography* **32**: 216–239. DOI: https://doi.org/10.1175/1520-0485(2002)032<0216: FMOSLI>2.0.CO;2

Harenduprakash L.L. & Mitra A.K. (1988). Vertical turbulent mass flux below the sea surface and air-sea interaction: monsoon region of the Indian ocean. *Deep Sea Research I* **43**(8): 423–1451. DOI: https://doi.org/10.1016/0198-0149(88)90014-3.

Jensen T.G. (2001). Arabian Sea and Bay of Bengal exchange of salt and tracers in an ocean model. *Geophysical Research Letters* **28**(20): 3967–3970. DOI: https://doi.org/10.1029/2001GL013422

Jinadasa S.U.P., Pathirana G., Ranasinghe P.N., Centurioni L. & Hormann V. (2020). Monsoonal impact on circulation pathways in the Indian Ocean. *Acta Oceanologica Sinica* **39**: 103–112. DOI: https://doi.org/10.1007/s13131-020-1557-5

Kumar S.P., Unnikrishnan A.S. & Muraleedharan P.M. (2000). Signatures of Kelvin and Rossby wave propagation in the northern Indian Ocean from TOPEX/POSEIDON Altimeter. *PORSEC Proceeding*, volume 1, Goa, India, pp. 276–279.

McCreary J.P., Kundu P.K. & Molinari R.L. (1993). A numerical investigation of dynamics, thermodynamics, and mixed layer processes in the Indian Ocean. *Progress in Oceanography* **31**(3): 181–224. DOI: https://doi.org/10.1016/0079-6611(93)90002-U

Patnaik K.V.K.R.K., Maneesha K., Sadhuram Y., Prasad K.V.S.R., Ramana Murty T.V. & Brahmananda Rao V. (2014). East India coastal current induced eddies and their interaction with tropical storms over Bay of Bengal. *Journal of Operational Oceanography* **7**(1): 58–68. DOI: https://doi.org/10.1080/1755876X.2014.11020153

Poterma J.T., Luther M.E. & O'Brien J.J. (1991). The seasonal circulation of the upper ocean in the Bay of Bengal. *Journal of Geophysical Research* **96**(C7): 12667–12683. DOI: https://doi.org/10.1029/91JC01045.

Rao A.R., Rao G.N. & Kumar P.V.H. (2013). Propagating waves and their influence on eddy field variability in the northeastern Arabian Sea and northern Bay of Bengal. *International Journal of Remote Sensing* **35**(14): 5467 – 5478. DOI: https://doi.org/10.1080/01431161.2014.926418.

Sadarjoen A. & Post F. (2000). Detection, quantification, and tracking of vortices using streamline geometry. *Computers and Graphics* **24**(3): 333–341.
DOI: https://doi.org/10.1016/S0097-8493(00)00029-7.

Schott F.A. & McCreary J.P. (2001). The monsoon circulation of the Indian Ocean. *Progress in Oceanography* **51**: 1 – 123.
DOI: https://doi.org/10.1016/S0079-6611(01)00083-0.

Schott F., Reppin J., Fischer J. & Quadfasel D. (1994). Currents and transports of the monsoon current south of Sri Lanka. *Journal of Geophysical Research: Oceans* **99**(C12): 25127–25141.
DOI: https://doi.org/10.1029/94JC02216.

Shankar D., McCreary J.P., Han W. & Shetye S.R. (1996). Dynamics of the east India coastal current: analytic solutions forced by interior Ekman pumping and local alongshore winds. *Journal of Geophysical Research* **101**(C6): 13975–13991.
DOI: https://doi.org/10.1029/96JC00559.

Shetye S.R., Gouveia A.D., Shenoi S.S.C., Sundar D., Michael G.S. & Nampoothiri G. (1993). The western boundary current of the seasonal subtropical gyre in the Bay of Bengal. *Journal of Geophysical Research* **98**(C1): 945 – 954.
DOI: https://doi.org/10.1029/92JC02070.

Somayajulu Y.K., Murty V.S.N. & Sarma Y.V.B. (2003). Seasonal and inter-annual variability of surface circulation in the Bay of Bengal from TOPEX/Poseidon altimetry. *Deep Sea Research II* **50**(5): 867–880.
DOI: https://doi.org/10.1016/S0967-0645(02)00610-0.

Thadathil P., Gopalakrishna V.V, Muraleedharan P.M., Reddy G.V, Araligidad N. & Shenoy S. (2002). Surface layer temperature inversion in the Bay of Bengal. *Deep Sea Research I* **49**: 1801–1818.
DOI: https://doi.org/10.1016/S0967-0637(02)00044-4.

Vinayachandran P.N., Masumoto Y., Mikawa T. & Yamagata T. (1999). Intrusion of the southwest monsoon current into the Bay of Bengal. *Journal of Geophysical Research: Oceans* **104**(C5): 11077–11085.
DOI: https://doi.org/10.1029/1999JC900035.

Vinayachandran P.N. & Yamagata T. (1998). Monsoon response of the sea around Sri Lanka: Generation of thermal domes and anticyclonic vortices. *Journal of Physical Oceanography* **28**: 1946–1960.
DOI: https://doi.org/10.1175/1520-0485(1998)028<1946:MROTSA>2.0.CO;2.

Wijesekera H.W. *et al.* (2016). ASIRI: An Ocean-Atmosphere Initiative for Bay of Bengal. *American Meteorological Society* **97**: 1859–1884.
DOI: https://doi.org/10.1175/BAMS-D-14-00197.1.

Yu L., O'Brien J.J. & Yang J. (1991). On the remote forcing of the circulation in the Bay of Bengal. *Journal of Geophysical Research* **96**(C11): 20449–20454.
DOI: https://doi.org/10.1029/91JC02424.

Yu Y., Wang L., Zheng Q. & Li Z. (2014). Geostrophic current estimation using altimetric cross-track method in northwest Pacific. *Earth and Environmental Science* **17**: 012105.
DOI: https://doi.org/10.1088/1755-1315/17/1/012105.