Rapid response to coastal upwelling in a semienclosed bay

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*Key Points*

Bidirectional estuarine exchange fluxes can be driven by coastal upwelling

Ria de Vigo response to coastal upwelling is faster than the inertial period

Remote wind with local wind and stratification modulate the rapid response

# Abstract

Bays/estuaries forced by local wind show bi-directional exchange flow. When forced by remote wind, they exhibit uni-directional flow adjustment to coastal sealevel. ADCP observations over one year show the Ria de Vigo (Iberian Upwelling) responds to coastal wind events with bi-directional exchange flow. The duration of the upwelling and downwelling events, estimated from the current variability, was ~3.3 days and ~2.6 days, respectively. Vectorial correlations reveal a rapid response to upwelling/downwelling, in which currents lag local wind by <6h and remote wind by <14h, less than the Ekman spin up (17.8h). This rapidity arises from the ria’s narrowness (non-rotational local response), equatorward orientation (additive remote and local wind responses), depth greater than the Ekman depth (penetration of shelf circulation into the interior), and vertical stratification (shear reinforcing shelf circulation). Similar rapid responses are expected in other narrow bays where local and remote winds act together and stratification enhances bi-directional flow.

# Introduction

The four major Eastern Boundary Upwelling Ecosystems (EBUE) of our oceans [[*Fréon et al.*, 2009](#_ENREF_13)] are among the most productive fisheries areas in the world oceans [[*Pauly and Christensen*, 1995](#_ENREF_39)]. Coastal upwelling occurs when alongshore wind forces an offshore Ekman transport of surface water – a persistent phenomenon due to equatorward winds along mid-latitude eastern boundaries of the major oceans. The divergence in Ekman transport near-surface results in the upwelling of deep waters near the coast, delivering nutrients to the near-surface euphotic zone. At higher latitudes, upwelling alternates with downwelling, which occurs during poleward wind events along these eastern boundaries. The large-scale circulation in EBUE [[*Hill et al.*, 1998](#_ENREF_21); [*Mackas et al.*, 2006](#_ENREF_30); [*Strub et al.*, 2013](#_ENREF_50)] and the mesoscale shelf circulation [[*Barth*, 2005](#_ENREF_2); [*Largier et al.*, 2006](#_ENREF_28)] have been well studied, but the landward extension of upwelling into bays is less well understood.

Four semienclosed and elongated bays (the Rias Baixas) are found at the northern limit of the Iberian-Canary Current EBUE [[*Evans and Prego*, 2003](#_ENREF_10)] (Figure 1). Coastal upwelling delivers nutrients to these rias, supporting high productivity and a major mariculture industry [[*Blanton et al.*, 1987](#_ENREF_6)]. Prior studies have shown that a residual vertical circulation in the Rías Baixas, in which upwelled water penetrates far into each ria [[*Álvarez-Salgado et al.*, 2000](#_ENREF_1); [*Barton et al.*, 2016](#_ENREF_3); [*Fraga*, 1981](#_ENREF_12); [*Souto et al.*, 2003](#_ENREF_48)], fundamentally characterizes these bay ecosystems.

Bays interact with the adjacent shelf as a result of buoyancy, wind and tidal forcing. Putting aside diffusive tidal exchanges (e.g., Tomales Bay: *Largier et al*. [[1997](#_ENREF_27)]) and tidal induced residual circulation [[*Winant*, 2008](#_ENREF_59)], the most effective shelf-bay exchange arises from vertical gradients in the momentum balance due to baroclinic and barotropic pressure gradients or boundary stresses. In estuaries, vertically bi-directional exchange flow with the ocean is envisaged as being purely baroclinic (induced by along-channel density gradients [[*Hickey et al.*, 2002](#_ENREF_20); [*Rattray and Hansen*, 1962](#_ENREF_44)]), barotropic (generated by local wind and volume conservation [[*Officer*, 1976](#_ENREF_36)]) or a combination of both [[*Blanton*, 1996](#_ENREF_5)]. When the remote (offshore) wind is the main barotropic forcing of the residual circulation then the exchange flow is observed [[*Noble et al.*, 1996](#_ENREF_34); [*Walters and Gartner*, 1985](#_ENREF_56); [*Wang*, 1979a](#_ENREF_57); [b](#_ENREF_58); [*Wong and Garvine*, 1984](#_ENREF_61)] and modeled, numerically and analytically, as uni-directional in plain estuaries [[*Garvine*, 1985](#_ENREF_14); [*Janzen and Wong*, 2002](#_ENREF_22); [*Masse*, 1990](#_ENREF_31)] and as uni-directional with transverse variations in triangular, parabolic or v-shaped cross-section estuaries [[*Kasai et al.*, 2000](#_ENREF_23); [*Valle-Levinson*, 2008](#_ENREF_52); [*Wong*, 1994](#_ENREF_60)].

Despite the difference in vertical structure (uni- vs. bi-directional) among remote and local wind induced circulations, the issue of identifying which forcing is more important in a particular estuary is still a difficult task [[*Wong and Mosses-Hall*, 1998](#_ENREF_62); [*Wong and Valle-Levinson*, 2002](#_ENREF_63)] . For example, equatorward facing bays seem to maintain uni-directional flows under remote upwelling winds [[*Valle-Levinson et al.*, 2003](#_ENREF_53)] and bi-directional flows with local winds [[*Valle-Levinson et al.*, 2004](#_ENREF_54)]. Interestingly, prior works in the Rias Baixas have shown that the remote wind in this coastal upwelling region interacts with the Ría de Vigo through a bi-directional exchange flow and not through a uni-directional one [[*Gilcoto et al.*, 2007](#_ENREF_16); [*Souto et al.*, 2003](#_ENREF_48)].

In this paper, we use a year-long record of velocity profiles in the central reaches of the Ría de Vigo to explore the bi-directional flow response to wind forcing. We find that this bi-directional exchange flow in the ria responds rapidly to the onset of winds, with lag times shorter than the inertial period which characterizes the Ekman transport response over the shelf. Our understanding is that local and remote wind forcing combine to account for this strong and rapid upwelling response, which is further enhanced by stratification associated with ria-ocean density differences. This general result can be expected in elongated bays where the two wind effects are reinforcing, but not in other elongated bays where the two effects are counteracting (e.g., Tomales Bay, California). This distinction implies that upwelling circulation (and related ecosystem productivity) will extend quickly and far into some bays in EBUE regions, but not others.

# Data and methods

Two meteorological stations and one Acoustic Doppler Current Profiler (ADCP) provide the data for this analysis. The meteorological stations provide 2-year time series of wind, from an offshore location (remote winds) and another inside the ria (local winds), from January 2013 to December 2014. The ADCP recorded, in the center of the ria, current velocity at multiple levels for one year from 20 June 2013 to 13 August 2014 (Figure 1).

The offshore wind station was the Cabo Silleiro oceanographic/meteorological buoy deployed and operated by Puertos del Estado (dataset available at <http://www.puertos.es/en-us/oceanografia/Pages/portus.aspx>). The buoy position was 9.43ºW 42.12ºN, on the 600m isobath and ~70km from the mouth of the Ría de Vigo. The sampling interval was 1 hour with data available for 82.20% of the 2-year period. Time series of wind velocity components were screened with a 6-hour filter to remove any values outside a range defined by twice the standard deviation from the window mean. Gaps shorter than 6 hours were filled by linear interpolation.

In the ria a Vantage Pro2 anemometer from Davis Instruments was installed on the breakwater of the Vigo harbor (8.757ºW 42.233ºN). Wind speed and direction were recorded every 1 minute and data are available for 96.77% of the 2-year measurement period. A 2-hour filter (as above) was applied to remove outliers and gaps shorter than 2 hours were filled by linear interpolation. The time series was sub-sampled at 1-hour intervals to allow direct comparison with the remote wind data.

The ADCP was deployed in the center of the channel in the middle of the ria (8.7615ºW 42.2398ºN), on the 45-meter isobath. A 1-km submarine cable provided power and communications for real-time access to the data at a high sampling rate (2Hz). The ADCP was a 600kHz RDI Workhorse Sentinel with 60 levels (bins) of cell-size 0.75m . It was deployed on a ballasted pyramid with a gimbal system to level the ADCP. The currents were recorded in radial coordinates and transformed to Earth coordinates following *Gilcoto et al.* [[2009](#_ENREF_15)]. This transformation also estimates errors for each velocity component: single-ping observations with errors greater than 20 cm/s were discarded. The 2Hz velocities were then averaged every minute, with an estimated error of ~1.5 cm·s-1, and subsequently sub-sampled hourly, yielding time series synchronous with both local and remote wind records. The ADCP also recorded water temperature at 0.6 m above the bottom (mab). ADCP observations are available for 93.76% of the 14-month deployment.

All time series were filtered with a Godin A24/24/25 filter [[*Godin*, 1972](#_ENREF_17)] to yield subtidal/sub-diurnal time series that are used in the following sections. Relationships between time series were investigated with the vector cross-correlation technique of *Kundu* [[1976](#_ENREF_25)], which gives a determination coefficient (R2) and the angle () at which maximum correlation is found. The angle of the principal axis was computed for each vector time-series following *Kundu and Allen* [[1976](#_ENREF_26)]. The significance levels of the correlations were calculated using the effective degrees of freedom, obtained by dividing the length of each series by its integral time scale estimated from the lagged auto-correlations following *Davis* [[1976](#_ENREF_8)]. All correlation reported here are significant at the 99.0% level.

# Results

Vertical profiles of R2 for cross-correlations between both local and remote winds and residual currents (Figure 2a) show an enhanced influence of winds near-surface (R2~60-70% of explained variance above 40 mab). This influence diminishes to a minimum (R2~5%) at ~33 mab, but increases again with depth until ~25 mab where it remains more or less constant (R2~50-60%) until the bottom. This correlation profile clearly demarcates a surface layer (>36 mab) and a bottom layer (<30 mab) where the wind explains more than 50% of the variance of the residual current, separated by a thin transition layer.

Similarly, the two-layer structure is evident in the profiles of  for both winds (Figure 2b). Near-surface currents are best correlated in the direction of the local wind forcing (~2.3º), while near-bottom currents are best correlated in a direction almost opposed to the wind (~171.3º). For remote winds, the angle of maximum correlation for surface currents in the ria is ~326.4º (i.e., when winds over the shelf are southerly, near-surface currents in the ria flow in a direction 35.6º east of north – into the ria), while near-bottom currents have maximum correlation at ~137.2º (i.e., flow out of the ria during southerly winds offshore).

These directional relationships between winds and currents are illustrated in Figure 3, with the wind direction shown as the principal axes for the 2-year time series. The principal axis for remote winds is orientated approximately North-South (13º east of North), while the local wind is oriented more East-West, parallel to the axis of the ria (72º east of North). Importantly, in both cases (local and remote) the  values are such that the currents are also oriented along the axis of the ria. Near-surface waters move out of the ria during northerly winds over the shelf and easterly winds over the ria, while near-bottom waters move landward (i.e., upwelling vertical circulation; Figures 3c and 3d). In contrast, during southerly winds over the shelf and westerly winds over the ria near-surface waters move into the ria, while near-bottom waters move seaward (i.e., downwelling vertical circulation; Figures 3e and 3f).

Winds both offshore and over the ria are related to large-scale atmospheric patterns and can be expected to be correlated. Vector correlation between wind records gives R2~80% with maximum correlation found at an angle of ~26º, so that Northerly winds over the shelf are strongly correlated with Northeasterly winds over the ria (and Southwesterly winds over the ria correlate with Southerly winds over the shelf). However, winds over the ria tend to lag winds over the shelf by a few hours: maximum correlation is at 3 hours (Figure 4a), but the lagged correlation curve has a broad peak because it includes variation over a whole year. In individual events, the ria winds can lag shelf winds by half a day or even lead shelf winds by few hours, depending on the speed and propagation direction of synoptic meteorological features.

Also important is the lag in the response of currents to winds (Figures 4c and 4d). Currents in the ria respond immediately and uniformly to local winds (lag less than 1 hour), except in the interface layer where R2 is lower. For forcing by remote winds, the response time increases to 6-8 hours, with more variable lags through the water column. However, while these lags represent maximum correlations, the correlation peaks are not sharp as the fluctuations in low-pass filtered winds and currents have time scales longer than a day (Figure 4d).

To examine the more event-driven nature of the ria sub-tidal dynamics, a running window technique was applied to the time series. The lagged Kundu vectorial cross-correlations between wind and residual currents in the surface and bottom layers were estimated in windows of 9 and 6 days for remote and local winds, respectively, long enough to frame inside them a complete upwelling or downwelling long-event while short enough to include only one or two short-events. The lag of maximum R2 was obtained for each window across the complete 1-year record (Figures 4e and 4f). For both winds and in both layers, the most frequent lag-class is that shorter than 1 hour related to events that persist longer than the window length. For the local wind, the most probable lag is 4 hours and then longer lags are less common. In contrast, the remote wind shows high relative frequency values at lags around 12±1 and 10±1 hours for the bottom and surface layers, respectively.

# Discussion

Prior studies in the Rias Baixas have identified remote wind as the primary forcing of residual circulation in these rias [[*Álvarez-Salgado et al.*, 2000](#_ENREF_1); [*Fraga*, 1981](#_ENREF_12); [*Pardo et al.*, 2001](#_ENREF_38); [*Rosón et al.*, 1997](#_ENREF_46)]. Although in the external part of the Ria de Vigo a three-dimensional circulation may take place [[*Barton et al.*, 2015](#_ENREF_4); [*Gilcoto et al.*, 2007](#_ENREF_16); [*Piedracoba et al.*, 2016](#_ENREF_42); [*Souto et al.*, 2003](#_ENREF_48)], in the middle ria a bi-directional exchange flow is seen clearly in vector cross-correlations of residual currents with filtered winds (this paper). Further, based on these correlations between long data series, our results reaffirm the importance of remote winds – they explain a little more of the variance in residual currents than the local wind – but we also note that the two wind signals are well correlated with each other (Figure 4a), which makes it difficult to use correlations to attribute the response in currents to either form of wind forcing. The high and low pressure systems in this region usually travel from West to East, from the Atlantic Ocean towards Europe, and their spatial scales of hundreds of kilometers (meso to synoptic meteorological scales) are larger than the length of the ria (~30km). The distance between our local (Vigo) and remote (Silleiro Buoy) wind stations is 82 km while the wind signal propagates between them in ~3 hours (Figure 4a), a sensible travelling velocity (~27km/h) for a storm (high pressure system) generating intense downwelling (upwelling).

Based on wind data, both *Nogueira et al.* [[1997](#_ENREF_35)] and *Torres et al.* [[*Torres et al.*, 2003](#_ENREF_51)] estimated that the average duration of a complete upwelling/downwelling cycle is 14-15 days (considering the wind as a stationary signal). However, *Nogueira et al.* [[1997](#_ENREF_35)] also showed that the seasonal cycle of the Ekman transport over the shelf accounts for only ~20% of the total variance, with additional variance due to higher frequencies. During one year there are many short and transient events of favorable upwelling/downwelling winds that are not close to the definition of a stationary signal cycle. Instead of using wind observations to estimate them, we can now directly observe the number and duration of upwelling and downwelling events with the filtered ADCP time series (Figure 4b). Defining an upwelling event by inflow in the layer 16-26 mab simultaneous with outflow in the layer 37-42 mab during a period longer than 8 hours, and a downwelling event by reversed flows in these layers, it is found that the most common upwelling events are only 1-day long and the average length of the 53 upwelling events detected in the ria was ~3.3 days; the most common downwelling events are 2-days long and the average length of the 49 downwelling events was ~2.6 days. Therefore, despite the tail of 10 days or longer in the histogram of event durations, the ADCP indicates event lengths notably shorter than the 7-8 day length implied by earlier analyses of wind data.

The shelf-ria system is very dynamic, and not only are upwelling events shorter than thought, but the year-round average response, estimated as the lag for maximum correlation, is only 8 hours for remote wind and less than one hour for local wind. Several authors have already noted that the residual circulation in the ria can respond rapidly to wind events, sometimes within a few hours [[*Piedracoba et al.*, 2005](#_ENREF_43); [*Souto et al.*, 2003](#_ENREF_48); [*Villacieros-Robineau et al.*, 2013](#_ENREF_55)]. However, the Ekman response in cross-shelf currents is expected to occur on a time scale given by the inertial period (17.8 hours) and even slower upwelling response is typically observed in bays that are sheltered topographically from winds and currents over the shelf [[*Rosenfeld et al.*, 1994](#_ENREF_45)]. Given that the remote wind is the first order forcing of the ria residual circulation, one could expect year-round lags closer to the inertial period. However, there are other forcing effects in addition to the remote wind, for instance, the local sea breezes in summer or increased river runoff during winter. These transient events and the events longer than the window length (>9 days), which may show zero lag, are distributed more or less randomly through the year. The combined result of this will be: i) to reduce both the maximum wind-current R2 and its corresponding lag and ii) to broaden the lag-R2 curve, thus rendering the calculated lag of maximum R2 less precise.

To account for the transient and event driven nature of the residual circulation of the ria, lagged vectorial cross-correlations were calculated between wind and current in running windows of 6 (local wind) or 9 (remote wind) days. In this way, it was found that the event response time of the surface and bottom layer to remote wind increases to 10 and 12 hours, respectively (Figure 4e). It is reasonable to have a slower response in the bottom layer than in the surface one. These lag times are closer to the inertial period but still 22% shorter. Therefore, we suggest that the initial forcing of bi-directional circulation in the ria is local wind stress.

The response of bays to winds over the shelf is typically thought of as being due to shelf-bay density differences (e.g.,[[*Blanton*, 1996](#_ENREF_5); [*Hickey et al.*, 2002](#_ENREF_20); [*Monteiro and Largier*, 1999](#_ENREF_32)]) or the interaction of the alongshore coastal jet with topography [[*Graham and Largier*, 1997](#_ENREF_18); [*Ryan et al.*, 2009](#_ENREF_47)]. These bay responses exhibit lag times of a day or longer, consistent with scales for an inviscid buoyancy-driven intrusion (time scale tbc~L/c’ bay length L~30km and c’~0.1m/s in the Ria de Vigo) or the spin-up time for a headland eddy (e.g., several days for St Helena Bay [[*Penven et al.*, 2000](#_ENREF_41)]). While [[*Blanton*, 1996](#_ENREF_5)] suggests that gravitational circulation over the southeastern U.S. continental shelf is reinforced by upwelling-favorable winds and stalled by downwelling winds, *Souto et al*. [[2003](#_ENREF_48)] and *Gilcoto et al*. [[2007](#_ENREF_16)] argue that since downwelling events can invert the circulation then the wind stress should be the primary forcing. Gravitational circulation may still reinforce upwelling circulation in an estuary or bay where denser ocean waters are found at the mouth – but this density difference will counter downwelling circulation. However, both in Galician rias and over the southeastern U.S continental shelf downwelling responses develop in a day or less, too fast to be explained by baroclinic adjustment.

To understand the way wind stress forces ria circulation we review the anticipated effects of local wind and remote wind independently. For time or space scales where coriolis can be ignored, the purely barotropic effect of local wind stress on residual currents in estuaries [[*Officer*, 1976](#_ENREF_36)], in lakes [[*Csanady*, 1973](#_ENREF_7)] and over the inner shelf [[*Lentz and Fewings*, 2012](#_ENREF_29)] has been shown to be bi-directional, with a surface layer driven directly by wind stress and a bottom layer driven by a pressure gradient due to the slope in the sea surface setup. In contrast, remote winds are expected to induce a uni-directional flow [[*Garvine*, 1985](#_ENREF_14); [*Wong*, 1994](#_ENREF_60)]. As the remote wind blows on the shelf, it produces a large-scale (compared with the estuary length) sea level setup or setdown at the mouth of the estuary and a net uni-directional inflow or outflow. Nevertheless, if the depth of the estuary mouth is deeper than the Ekman depth then, instead of the uni-directional flow due to sea level variations, the standard two-layer upwelling/downwelling circulation (surface flow and compensatory bottom flow) can also penetrate into the estuary [[*Gilcoto et al.*, 2007](#_ENREF_16); [*Souto et al.*, 2003](#_ENREF_48)][.](#_ENREF_46)

The response to wind stress over the shelf is expected to develop slowly over an inertial period when starting smoothly from a calm situation. However, the bi-directional exchange flow in the ria develops quickly as the weather fronts, large compared with the ria dimensions, approach the coast. This rapid response may be partially due to the propagation of offshore wind-driven sea level perturbations to the ria. While this effect is expected to drive a uni-directional flow in the ria, the local wind stress on the ria associated with arrival of the weather front can be expected to drive a rapid response in the surface layer of the ria and establish a bi-directional flow within a few hours (Figures 4d and 4f). This bi-directional flow in the ria due to local easterlies (westerlies) is accompanied by upwelling (downwelling) favorable winds over the shelf, which spin up Ekman transport and coastal upwelling within an inertial period. Thus, the transport due to non-rotational wind-driven surface flow in the ria merges with the developing rotational (Ekman) wind-driven surface flow over the shelf. In this way, the Ekman response over the shelf extends rapidly far into the ria in a way previously not observed in upwelling bays [[*Valle-Levinson et al.*, 2003](#_ENREF_53); [*Valle-Levinson et al.*, 2004](#_ENREF_54)] but which may resemble the important effects of coastal upwelling on fjords [[*Erga et al.*, 2012](#_ENREF_9); [*Straneo et al.*, 2010](#_ENREF_49)].

In spite of the secondary role of buoyancy forcing in longitudinal exchange flow in the ria, stratification plays an important role in its response to upwelling winds. In the first few days, easterly winds over the ria advect warm low-salinity water from the inner ria over denser outer-ria waters and later the upwelling of cold dense water at the mouth of the ria and intrusion in the bottom layer maintains stratification [[*Barton et al.*, 2015](#_ENREF_4)]. The strong vertical shear observed at the base of the surface layer (Figure 5) is enabled by this stratification: observed tidal and sub-tidal shears of ~0.25m/s or less over Δz~10m remain stable (Ri>1) if the density difference is Δρ~0.64kg/m3 or stronger, which is typically true (see *Barton et al*.[[2015](#_ENREF_4)]). Stratification retains the momentum gain from surface wind stress in a well-defined, shallow surface layer, allowing rapid acceleration and strong shear across the pycnocline. This stratification also allows the deeper waters to simultaneously move in the opposite direction with minimal fluid drag. A buoyant surface layer may also form during downwelling phases, when low-salinity waters in the Western Iberian Buoyancy Current [[*Peliz et al.*, 2005](#_ENREF_40)] are found at the mouth of the ria [[*Mouriño and Fraga*, 1982](#_ENREF_33)]. But, while an ongoing buoyancy influx may maintain stratification during the upwelling phase (small freshwater inflow at the head of the ria, plus surface warming along ria), stratification may not be maintained during downwelling and the shear in the bi-directional flow may weaken, although it will persist as long as wind forcing persists [[*Barton et al.*, 2016](#_ENREF_3)]. This difference between upwelling and downwelling processes is reflected in the average velocity profiles for each situation (Figure 5). The zero residual velocity in upwelling events is located ~2 meters above of that corresponding to downwelling circulations, and the maximum velocity in the bottom layer is also shallower in the upwelling profile. Both observations indicate a shallower surface boundary layer and suggest more stratification under upwelling conditions.

This study of a long, narrow bay in the Galician upwelling region provides insight useful for understanding bays more generally in upwelling regions. Three factors account for the rapid extension of upwelling far into this ria bay – in other bays upwelling may not extend as far or not as quickly. Firstly, the bay is narrower than the Rossby radius so that the primary response to local winds is non-rotational and rapid; broader bays would exhibit a slower, rotational response. Secondly its orientation ensures that local winds inside the bay drive transport in the surface boundary layer in the same direction as the Ekman layer transport over the shelf; the opposite occurs in bays that face into upwelling winds, e.g., Tomales Bay, California. In the absence of local wind forcing, upwelling effects penetrate slowly into longer bays only through density-driven intrusions and/or headland eddy flows. Thirdly, vertical stratification supports strong and shallow vertical shear that allows a more rapid response and extension of the vertical circulation further into the bay. Thus the aspect ratio of the ria (length >> width) has allowed us to address two associated but distinct roles of surface wind stress in upwelling bays – both the remote wind stress over the shelf and local wind stress over the ria. In many smaller bays, the size/shape of the bay [[*Monteiro and Largier*, 1999](#_ENREF_32)] or the complexity of the wind field [[*Paduan and Rosenfeld*, 1996](#_ENREF_37)] precludes a clear exposition of these two forcing mechanisms. Also, we differentiate the two roles of density structure: (i) longitudinal density gradients that may drive an estuarine circulation, and (ii) vertical density gradients that enable high vertical shear. While these two roles are woven together in classical estuarine literature (e.g., *Hansen and Rattray* [[1965](#_ENREF_19)]), here we show the importance of the latter even when the bi-directional flow is mainly driven by surface wind stress.

This description of vertical circulation in the Ria de Vigo shows the importance of specific bay characteristics that allow extension of the shelf upwelling system far into the sheltered waters of the bay. This helps to explain the extent and reliability of the productive ecosystem and high levels of bivalve culture in the ria [[*Figueiras et al.*, 2002](#_ENREF_11)], which is not found in all upwelling bays. Given the high frequency of upwelling/downwelling wind variability in this region [[*Nogueira et al.*, 1997](#_ENREF_35); [*Torres et al.*, 2003](#_ENREF_51)], the rapid response of the ria means that the basin is well-flushed (ventilated), ensuring a continuous influx of high-nutrient waters from upwelling over the adjacent shelf and continuous removal of any oxygen depletion effects in the ria. In comparable but shallower bays, or bays with local winds opposed to wind effects over the shelf, like Tomales Bay [[*Kimbro et al.*, 2009](#_ENREF_24); [*Largier et al.*, 1997](#_ENREF_27)], this rapid exchange does not happen and the mid/inner bay exhibits severe nutrient depletion and reduced productivity [[*Kimbro et al.*, 2009](#_ENREF_24)].

# Conclusions

Vectorial cross-correlations, between winds measured at two different locations (remote and local) and water velocity profiles recorded in the interior of the Ria de Vigo allowed us to examine the forcing of the bi-directional flow in the residual circulation of the ria. Both determination coefficients and angles of maximum correlation portrayed this vertical structure. Residual currents are more correlated with remote than with local winds. Both winds are also well correlated with each other since they are driven by the incoming eastward traveling atmospheric lows and highs. From 2 years of observations, the remote wind over the continental shelf leads the local wind in the ria interior by 3 hours.

The response of the residual currents in the ria to the winds is fast. Local wind produces responses within 6 hours while the lag associated with remote wind is twice as long. Even the latter is faster than the spin up time (~17.8h) of the coastal upwelling. Further, the average duration of upwelling events is ~3.3 days and of downwelling events is ~2.6 days. Therefore, ria subtidal dynamics, revealed by direct current observations, are more rapid and variable in time than suggested by previous estimations from winds.

The rapid response of the ria to the wind encourages us to consider the wind-driven barotropic bi-directional flow as the first order mechanism of residual circulation in the ria. In such mechanism, the upwelling/downwelling cross-shelf circulation structure penetrates into the ria aided by an internal bi-directional flow initiated by local wind and the reduction of friction between layers generated by pre-existing vertical stratification. More studies must be undertaken to increase our knowledge on secondary forcings that may modulate the residual circulation of the Rias Baixas, or primary forcings that predominate in other types of events.

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# References

Álvarez-Salgado, X. A., J. Gago, B. M. Míguez, M. Gilcoto, and F. F. Pérez (2000), Surface waters of the NW Iberian margin: upwelling on the shelf versus outwelling of upwelled waters from the Rías Baixas, *Estuarine, Coastal and Shelf Science*, *51*(6), 821-837.

Barth, J. A. (2005), Introduction to special section: Coastal Advances in Shelf Transport, *Journal of Geophysical Research*, *110*(C10).

Barton, E. D., R. Torres, F. G. Figueiras, M. Gilcoto, and J. Largier (2016), Surface water subduction during a downwelling event in a semienclosed bay, *Journal of Geophysical Research: Oceans*, *121*(9), 7088-7107.

Barton, E. D., J. L. Largier, R. Torres, M. Sheridan, A. Trasviña, A. Souza, Y. Pazos, and A. Valle-Levinson (2015), Coastal upwelling and downwelling forcing of circulation in a semi-enclosed bay: Ria de Vigo, *Prog. Oceanogr.*, *134*, 173-189.

Blanton, J. O. (1996), Reinforcement of gravitational circulation by wind, in *Buoyancy effects on coastal and estuarine dynamics*, edited by D. G. Aubrey and C. T. Friedrichs, pp. 47-58, American Geophysical Union, Washington, DC.

Blanton, J. O., K. R. Tenore, F. Castillejo, L. P. Atkinson, F. B. Schwing, and A. Lavín (1987), The relationship of upwelling to mussel production in the rias on the western coast of Spain, *Journal of Marine Research*, *45*(2), 497-511.

Csanady, G. T. (1973), Wind-Induced Barotropic Motions in Long Lakes, *Journal of Physical Oceanography*, *3*(4), 429-438.

Davis, R. E. (1976), Predictability of sea surface temperature and sea level pressure anomalies over the North Pacific Ocean, *Journal of Physical Oceanography*, *6*(3), 249-266.

Erga, S. R., N. Ssebiyonga, Ø. Frette, B. Hamre, J. Aure, Ø. Strand, and T. Strohmeier (2012), Dynamics of phytoplankton distribution and photosynthetic capacity in a western Norwegian fjord during coastal upwelling: Effects on optical properties, *Estuarine, Coastal and Shelf Science*, *97*, 91-103.

Evans, G., and R. Prego (2003), Rias, estuaries and incised valleys: is a ria an estuary?, *Marine Geology*, *196*(3-4), 171-175.

Figueiras, F. G., U. Labarta, and M. J. Reiriz (2002), Coastal upwelling, primary production and mussel growth in the Rías Baixas of Galica, *Hydrobilogia*, *484*, 121-131.

Fraga, F. (1981), Upwelling off the Galician coast, northwest Spain, in *Coastal upwelling*, edited by F. A. Richards, pp. 176-182, Coastal and esturarine science 1, Washington, DC.

Fréon, P., M. Barange, and J. Arístegui (2009), Eastern Boundary Upwelling Ecosystems: Integrative and comparative approaches, *Prog. Oceanogr.*, *83*(1-4), 1-14.

Garvine, R. W. (1985), A simple model of estuarine subtidal fluctations forced by local and remote wind stress, *Journal of Geophysical Research*, *90*(C6), 11945-11948.

Gilcoto, M., E. Jones, and L. Fariña-Busto (2009), Robust estimations of current velocities with four-beam broadband ADCPs, *Journal of Atmospheric and Oceanic Technology*, *26*(12), 2642-2654.

Gilcoto, M., P. C. Pardo, X. A. Álvarez-Salgado, and F. F. Pérez (2007), Exchange fluxes between the Ría de Vigo and the shelf: A bidirectional flow forced by remote wind, *Journal of Geophysical Research*, *112*(C06), 21.

Godin, G. (1972), *The analysis of tides*, 264 pp., Liverpool University Press, Liverpool.

Graham, W. M., and J. L. Largier (1997), Upwelling shadows as nerashore retention sites: the example of northern Monterey Bay, *Cont. Shelf Res.*, *17*(5), 509-532.

Hansen, D. V., and M. Rattray (1965), Gravitional Circulation in Straits and Estuaries, *Journal of Marine Research*, *23*(2), 104-122.

Hickey, B. M., X. Zhang, and N. Banas (2002), Coupling between the California Current System and a coastal plain estuary in low riverflow conditions, *Journal of Geophysical Research*, *107*(C10), 30\_31-30\_20.

Hill, A. E., B. M. Hickey, F. A. Shillington, P. T. Strub, K. H. Brink, E. D. Barton, and A. C. Thomas (1998), Eastern Ocean Boundaries, in *The Global Coastal Ocean: Regional Studies and Syntheses*, edited by A. R. Robinson and K. H. Brink, pp. 29-67, Harvard University Press, Cambridge.

Janzen, C. D., and K.-C. Wong (2002), Wind-forced dynamics at the estuary-shelf interface of a large coastal plain estuary, *Journal of Geophysical Research*, *107*(C10), 2\_1-2\_12.

Kasai, A., A. E. Hill, T. Fujiwaka, and J. H. Simpson (2000), Effect of the Earth's rotation on the circulation in regions of freshwater influence, *Journal of Geophysical Research*, *105*(C07), 16961-16969.

Kimbro, D. L., J. L. Largier, and E. D. Grosholz (2009), Coastal oceanographic processes influence the growth and size of a key estuarine species, the Olympia oyster, *Limnology and Oceanography*, *54*(5), 1425-1437.

Kundu, P. K. (1976), Ekman veering observed near the ocean bottom, *Journal of Physical Oceanography*, *6*(2), 238-242.

Kundu, P. K., and J. S. Allen (1976), Some three-dimensional characteristics of low-frequency current fluctuations near the Oregon coast, *Journal of Physical Oceanography*, *6*(2), 181-199.

Largier, J. L., J. T. Hollibaugh, and S. V. Smith (1997), Seasonally Hypersaline Estuaries in Mediterranean-climate Regions, *Estuarine, Coastal and Shelf Science*, *45*(6), 789-797.

Largier, J. L., et al. (2006), WEST: A northern California study of the role of wind-driven transport in the productivity of coastal plankton communities, *Deep Sea Research Part II: Topical Studies in Oceanography*, *53*(25-26), 2833-2849.

Lentz, S. J., and M. R. Fewings (2012), The wind- and wave-driven inner-shelf circulation, *Annual Review of Marine Sciences*, *4*, 317-343.

Mackas, D. L., P. T. Strub, A. Thomas, and V. Montecino (2006), Eastern regional ocean boundaries pan-regional overview, in *The Global Coastal Ocean, Interdisciplinary Regional Studies and Syntheses. Panregional Syntheses and the Coast of North and South America and Asia*, edited by A. R. Robinson and K. H. Brink, pp. 21-59, Harvard University Press, Cambridge.

Masse, A. K. (1990), Withdrawal of shelf water into an estuary: a barotropic model, *Journal of Geophysical Research*, *95*(C9), 16085-16096.

Monteiro, P. M. S., and J. L. Largier (1999), Thermal stratification in Saldanha Bay (South Africa) and subtidal, density-driven exchange with the coastal waters of the benguela upwelling system, *Estuarine, Coastal and Shelf Science*, *49*(6), 877-890.

Mouriño, C., and F. Fraga (1982), Hidrografía de la Ría de Vigo. 1976-1977. Influencia anormal del Río Miño, *Invest. Pesq. (Barc.).* *46*(3), 459-468.

Noble, M. A., W. W. Schoeder, W. J. Wiseman, Jr., H. F. Ryan, and G. Gelfenbaum (1996), Subtidal circulation patterns in a shallow, highly stratified estuary: Mobile Bay, Alabama, *Journal of Geophysical Research*, *101*(C11), 25689-25703.

Nogueira, E., F. F. Pérez, and A. F. Ríos (1997), Seasonal patterns and long-term trends in an estuarine upwelling ecosystem (Ría de Vigo, NW Spain), *Estuarine, Coastal and Shelf Science*, *44*(3), 285-300.

Officer, C. B. (1976), *Physical oceanography of estuaries (and associted coastal waters)*, 465 pp., John Wiley & Sons, New York.

Paduan, J. D., and L. K. Rosenfeld (1996), Remotely sensed surface currents in Monterey Bay from shore-based HF radar (Coastal Ocean Dynamics Application Radar), *Journal of Geophysical Research*, *101*(C9), 20669.

Pardo, P. C., M. Gilcoto, and F. F. Pérez (2001), Short-time scale coupling between termohaline and meteorological forcing in the Ría de Pontevedra, *Scientia Marina*, *65*(Suppl. 1), 229-240.

Pauly, D., and V. Christensen (1995), Primary production required to sustain global fisheries, *Nature*, *374*, 255-257.

Peliz, Á., J. Dubert, A. M. P. Santos, P. B. Oliveira, and B. Le Cann (2005), Winter upper ocean circulation in the Western Iberian Basin—Fronts, Eddies and Poleward Flows: an overview, *Deep Sea Research Part I: Oceanographic Research Papers*, *52*(4), 621-646.

Penven, P., C. Roy, A. Colin de Verdière, and J. Largier (2000), Simulation of a coastal jet retention process using a barotropic model, *Oceanologica Acta*, *23*(5), 615-634.

Piedracoba, S., G. Rosón, and R. A. Varela (2016), Origin and development of recurrent dipolar vorticity structures in the outer Ría de Vigo (NW Spain), *Cont. Shelf Res.*, *118*, 143-153.

Piedracoba, S., X. A. Álvarez-Salgado, G. Rosón, and J. L. Herrera (2005), Short-timescale thermohaline variability and residual circulation in the central segment of the coastal upwelling system of the Ría de Vigo (northwest Spain) during four contrasting periods, *Journal of Geophysical Research*, *110*(C03), 15.

Rattray, M., and D. V. Hansen (1962), A similarity solution for circulation in an estuary, *Journal of Marine Research*, *20*(2), 121-133.

Rosenfeld, L. K., F. B. Schwing, N. Garfield, and D. E. Tracy (1994), Bifurcated flow from an upwelling center: a cold water source for Monterey Bay, *Cont. Shelf Res.*, *14*(9), 931-964.

Rosón, G., X. A. Álvarez-Salgado, and F. F. Pérez (1997), A non-stationary box model to determine residual fluxes in a partially mixed estuary, based on both thermohaline properties: Application to the Ria de Arousa (NW Spain), *Estuarine, Coastal and Shelf Science*, *44*(3), 249-262.

Ryan, J. P., A. M. Fischer, R. M. Kudela, J. F. R. Gower, S. A. King, R. Marin, and F. P. Chavez (2009), Influences of upwelling and downwelling winds on red tide bloom dynamics in Monterey Bay, California, *Cont. Shelf Res.*, *29*(5-6), 785-795.

Souto, C., M. Gilcoto, L. Fariña-Busto, and F. F. Pérez (2003), Modelling the residual circulation of a coastal embayment affected by wind driven upwelling: circulation of the Ria de Vigo (NW Spain), *Journal of Geophysical Research*, *108*(C11), 3340-3357.

Straneo, F., G. S. Hamilton, D. A. Sutherland, L. A. Stearns, F. Davidson, M. O. Hammill, G. B. Stenson, and A. Rosing-Asvid (2010), Rapid circulation of warm subtropical waters in a major fjord in East Greenland, *Nature Geosciences*, *3*(3), 182-186.

Strub, P. T., V. Combes, F. A. Shillington, and O. Pizarro (2013), Currents and Processes along the Eastern Boundaries, in *Ocean Circulation & Climate. A 21st Century Perspective*, edited by G. Siedler, S. M. Griffies, J. Gould and J. A. Church, pp. 339-384, Elsevier-Academic Press, Amsterdam.

Torres, R., E. D. Barton, P. Miller, and E. Fanjul (2003), Spatial patterns of wind and sea surface temperature in the Galician upwelling region, *Journal of Geophysical Research*, *108*(C4), 3130.

Valle-Levinson, A. (2008), Density-driven exchange flow in terms of the Kelvin and Ekman numbers, *Journal of Geophysical Research*, *113*(C04), 10.

Valle-Levinson, A., L. P. Atkinson, D. Figueroa, and L. Castro (2003), Flow induced by upwelling winds in an equatorward facing bay: Gulf of Arauco, Chile, *Journal of Geophysical Research*, *108*(C2), 3054-3067.

Valle-Levinson, A., et al. (2004), Wind-induced exchange at the entrance to Concepción Bay, an equatorward facing embayment in central Chile, *Deep Sea Research Part II: Topical Studies in Oceanography*, *51*(20-21), 2371-2388.

Villacieros-Robineau, N., J. L. Herrera, C. G. Castro, S. Piedracoba, and G. Roson (2013), Hydrodynamic characterization of the bottom boundary layer in a coastal upwelling system (Ría de Vigo, NW Spain), *Cont. Shelf Res.*, *68*, 67-79.

Walters, R. A., and J. W. Gartner (1985), Subtidal sea level and current variations in the Northern Reach of San Francisco Bay, *Estuarine, Coastal and Shelf Science*, *21*, 17-32.

Wang, D.-P. (1979a), Wind-driven circulation in the Chesapeak Bay, winter 1975, *Journal of Physical Oceanography*, *9*, 564-572.

Wang, D.-P. (1979b), Subtidal sea level variations in the Chesapeake Bay and relations to atmospheric forcing, *Journal of Physical Oceanography*, *9*, 413-421.

Winant, C. D. (2008), Three-dimensional residual tidal circulation in an elongated, rotating basin, *Journal of Physical Oceanography*, *38*(6), 1278-1295.

Wong, K.-C. (1994), On the nature of transverse variability in a coastal plain estuary, *Journal of Geophysical Research*, *99*(C07), 14209-14222.

Wong, K.-C., and R. W. Garvine (1984), Observations of wind-induced, subtidal variability in the Delaware Estuary, *Journal of Geophysical Research*, *89*(C6), 10589-10597.

Wong, K.-C., and J. E. Mosses-Hall (1998), On the relative importance of the remote and local wind effects to the subtidal variability in a coastal plain estuary, *Journal of Geophysical Research*, *103*(C9), 18393-18404.

Wong, K.-C., and A. Valle-Levinson (2002), On the relative importance of the remote and local wind effects on the subtidal exchange at the entrance to the Chesapeake Bay, *Journal of Marine Research*, *60*(3), 477-498.

Figure 1. Map showing the four Rías Baixas of Galicia (Spain) and the locations where the remote winds (Silleiro Buoy), local winds (Vigo Harbor), and currents (ADCP in the middle of the Ría de Vigo) were recorded.

Figure 2. Vertical profiles of a) determination coefficient (R2) and b) direction of best correlation (counterclockwise from currents) corresponding to vectorial cross-correlations between filtered-wind and residual-current time series.

Figure 3. Plan view of the geometric relationships between wind and current patterns – depicted with arrows overlaid on a map of the Ría de Vigo. Blue arrows correspond to the principal axes of winds (remote on the left column and local on the right) while the red and green arrows to the directions of the surface (>36 mab) and bottom (<30 mab) layers, respectively, of the residual currents rotated following maximum correlation angles between currents and winds (see text). Panels (a) and (b) represent the general case, with both senses for each arrow direction. Panels (c) and (d) show the arrows for upwelling conditions, and panels (e) and (f) for downwelling.

Figure 4. (a) Lagged vectorial cross-correlation between remote and local winds; negative lag if remote leads local wind. (b) Histogram showing duration of upwelling (blue) and downwelling (green) events determined by periods of bi-directional flow observed in the ADCP record. Every period of time longer that 8h is considered “upwelling” if the residual current in the surface layer (37-42 mab) was >5cm·s-1 outward and simultaneously >5cm·s-1 inward in the bottom layer (16-26 mab). And “downwelling” when current was inward in surface layer and outward in bottom layer. (c) Contour plots with determination coefficients (R2) from vectorial cross-correlations between filtered remote winds and residual currents as a function of depth (mab) and lag (hours); negative lags indicate that the wind leads the current. The black line joins the lags of maximum R2 for each depth. (d) Same for local winds. (e) Histograms represent the relative frequency of 9-day wind-current lags of maximum vectorial cross-correlation between residual currents, in the surface layer (blue bars) and bottom layer (green bars), with remote winds. (f) Same for local winds, but with 6-day window.

Figure 5. Vertical profiles of mean residual along-channel current, positive velocities represent flow into the ria (70º from N) while negative ones exit the estuary. Blue line for upwelling conditions and green for downwelling. Gray shading represents the standard deviation in both cases. The criteria for selecting the profiles to be averaged in upwelling or downwelling conditions were the same as in Figure 4b.



**Figure 1**



**Figure 2**



**Figure 3**



**Figure 4**



**Figure 5**