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The response of large outflows to wind forcing

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ABSTRACT

A numerical model is used to gauge the impact of winds on the evolution of coastal plumes generated by a variety of inlet outflows. The analysis is summarized by a conceptual model that accounts for the formation of surface and bottom mixed layers and tilting of the plume front. It also provides the basis for a two parameter classification of upwelling. The first parameter indicates when a wind event is capable of fully exporting plume waters offshore. The second determines when winds can overcome the plume buoyancy-driven flow. These indices help to explain why larger outflows tend to be less susceptible to upwelling. During an upwelling event, large plumes tend to maintain their structure, while smaller systems are commonly detached and dispersed offshore. The onset of downwelling events often reorganizes large plumes, thus promoting their net downshelf displacement. In contrast smaller systems frequently restart their formation, consequently limiting their downshelf penetration. The addition of long-term fluctuations, superimposed to the synoptic wind forcing, suggests a mechanism for typical seasonal to interannual variability commonly observed for large discharges.

1. Introduction

Coastal outflow plumes are important agents for the transport of organisms, sediments and chemical substances in the coastal ocean (Garvine, 1995; Hill, 1998; Henrichs et al., 2000). Formed by the freshwater supply from estuaries, they typically propagate along the coast as buoyancy-driven flows, but also respond to tides (Garvine, 1999), winds (Lentz and Largier, 2006) and ambient currents (García Berdeal et al., 2001; Fong and Geyer, 2002; Avicola, 2003). Of primary concern here is the impact of alongshelf winds on the development and dispersal of plumes. Observations indicate that plumes' responses are dominantly governed by Ekman dynamics (Fong et al., 1997; Hallock and Marmorino, 2002). The onshore transport associated with *downwelling-favorable* winds tends to keep plumes against the coast, usually promoting their intrusion in the form of narrow coastal jets (Rennie et al., 1999; Moffat and Lentz, 2012). *Upwelling-favorable* winds, on the other hand, promote offshore and upshelf transport. Plumes typically evolve as thin surface layers that mix and stretch offshore (e.g. Fig. 5.4 in Rennie, 1998 and Fig. 9 in Lentz, 2004). Eventually many plumes detach and move offshore (Fong and Geyer, 2001; Houghton et al., 2004).

The role of winds in coastal circulation is of considerable theoretical and practical interest. Chao (1987) demonstrated how highly asymmetric responses occur for upwelling- and downwelling-favorable configurations. Fong and Geyer (2001) developed a conceptual model for the cross-shelf response to upwelling and prediction of the plume leading edge thickness and velocity of propagation. Lentz (2004) explicitly included the process of entrainment to Fong and Geyer's model for the upwelling of plumes. Whitney and Garvine (2005) studied the influence of winds on the alongshelf circulation with an index that compared the strength of buoyancy- and wind-driven currents. Kourafalou et al. (1996a) and Zhang et al. (2009) described the freshwater pathways to the deep ocean.

Most of these modeling studies have scaled the problem for small to medium rivers from ~ 250 to $2500\,m^3\,s^{-1}$. These are the discharges observed for systems like the Chesapeake and Delaware, which typically extend from 100 to 150 km from the bay mouth and 15 to 20 m over a steep inner-shelf (Boicourt, 1973; Wong and Münchow, 1995; Le Vine et al., 1998). Few studies have considered larger discharges, although there are important point-source mid-latitude outflows. Of the top twenty global largest discharges, for example, *fourteen* are located in latitudes greater than 20° (Hovius, 1998). These discharges vary from 7600 to $31\,000\,m^3\,s^{-1}$ and include rivers such as the Columbia, Mississippi and St. Lawrence in North America, the Yenisei in Russia and the Yangtze in China. The Rio de la Plata, for example, discharges $\sim 23\,000\,m^3\,s^{-1}$ and forms a large plume that reaches ~ 60 m depth, extends ~ 100 km offshore and ~ 750 km alongshelf (Guerrero et al., 1997; Möller et al., 2008). Fig. 1 summarizes these scales.

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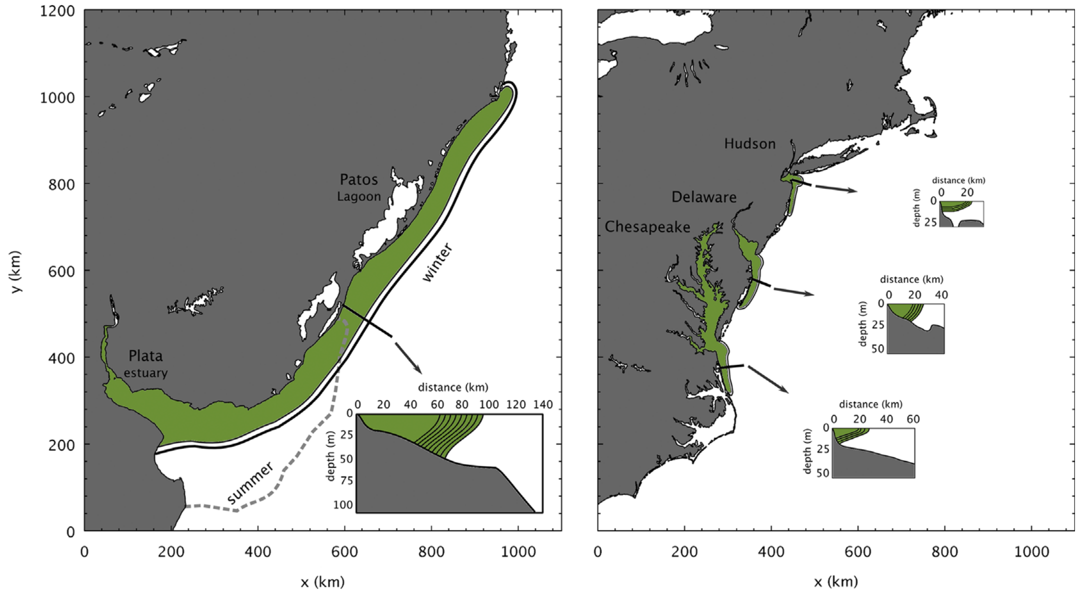

Fig. 1. Data derived cartoon illustrating the differences on the size and structure of plumes derived from smaller and larger outflows. Smaller discharges are shown on the right panel: the Hudson (Chant et al., 2008; Choi and Wilkin, 2007), the Delaware (Wong and Münchow, 1995; Le Vine et al., 1998), and the Chesapeake (Donato and Marmorino, 2002; Lentz and Largier, 2006). The left panel shows Plata plume historical winter (solid line) and summer (dashed line) surface positions and the winter vertical cross-section (Piola et al., 2008a; Möller et al., 2008).

Despite many similarities these coastal currents have markedly different responses to wind. While upwelling-favorable winds were shown to reverse and detach the Delaware (Münchow and Garvine, 1993), observations from Plata suggest that individual upwelling events should rarely destroy the plume (Pimenta et al., 2008). Plata also exhibits large seasonal variability. During the winter, the plume extends for 1200 km from the estuary mouth, while during the summer its typical downshelf intrusion is ~ 400 km (Möller et al., 2008) (Fig. 1). Although this migration has been attributed to winds, the specific role of synoptic and seasonal wind components in the evolution of coastal plumes remains unclear.[1]

Our primary goal here is to quantify the role of the discharge magnitude in the response of coastal plumes to wind forcing. The approach is to perform numerical simulations with a three-dimensional hydrodynamic model implemented to a simplified shelf and estuary bathymetry. Model details are given in Section 2. Motivated by Delaware and Plata observations we study a wide range of plumes that vary from inner-shelf to mid-shelf density fronts. These experiments are described in Section 3. Plume responses to constant upwelling-favorable winds are explored in Section 4, while the theory and conceptual model are presented in Section 5. Varying upwelling and downwelling wind simulations are explored in Section 6, where the responses to synoptic forcing are described for different outflow magnitudes. Section 7 presents a summary and concluding remarks.

2. Model description

The Princeton Ocean Model (POM, Blumberg and Mellor, 1987) is used to study the effect of inlet discharge magnitude on the wind response of coastal plumes. POM is a finite-difference hydrostatic model in terrain-following coordinates that treats non-linear time dependent flows in three-dimensions. This model is widely used in process-oriented and realistic simulations, see Mellor (2004) and Kourafalou et al. (1996b) for the governing equations, numerical schemes or physical parameterizations.

For the present application, we follow a geometry that is typical of continental margins such as the Mid-Atlantic Bight in the US and the Southern Brazilian Shelf. The domain is composed of a steep inner-shelf region and a continental shelf:

$$h_{shelf} = \begin{cases} h_c + \alpha_n y & (y \le y_n) \\ h_n + \alpha (y - y_n) & (y \ge y_n). \end{cases} \tag{1}$$

For the simulations performed the coastal wall depth is set to be $h_c = 4$ m, the inner shelf width to be $y_n = 5$ km and the slope is $\alpha_n = 2.4 \times 10^{-3}$, so that depth h_n is 16 m. The continental shelf has a mild slope, $\alpha = 6.5 \times 10^{-4}$, and the domain is limited to 150 m depth.[2] As the focus is on inner- to mid-shelf plumes, the continental slope and abyssal plain are not included. These should be considered, however, if one attempts to reproduce the dynamics of shelf-break fronts (e.g. Gwarkiewicz and Chapman, 1992).

A steady inlet discharge Q_i is delivered through a channel of 90° exit angle to the coastline, with “round” estuary corners (Fig. 2). The channel has a Gaussian profile given by $h_i = h_i^* e^{-x^2/2\sigma_i^2}$, where x represents the alongshelf coordinate, h_i^* the channel depth and $\sigma_i = b_i/6$, where the channel width is evaluated from

$$b_i = \left(\frac{Q_i}{fRoA_i} \right), \tag{2}$$

here f is the Coriolis parameter, Ro is the Rossby number, and the channel cross-sectional area is approximated by $A_i \sim (\sqrt{2\pi}/6) h_i^* b_i$. Ro is set to be ~ 0.04 to suppress the formation of large bulges (Fong and Geyer, 2002; Pimenta et al., 2011). To generate channels of a similar depth to width ratio, we set $A_i = \delta b_i^2$ with $\delta = 4 \times 10^{-4}$. This results in $b_i = (Q_i/\delta f\ Ro)^{1/3}$. A channel of length $3b_i$ continues offshore and merges to the shelf bottom, thus allowing the development of the estuarine circulation.

At the channel head, the inlet velocity is prescribed by $v_i = Q_i/A_i$. The inlet discharge Q_i ranges from 20 000 up to

[1] Here the term “synoptic” refers to alongshelf wind fluctuations due to the passage high- and low-pressure systems with the temporal scale of days and typical spatial scale of $\sim 10^3$ km. “Seasonal” refers to the migration of basin- to continental-scale systems that cause changes of winds on the scale of months.

[2] A list of variables is available in Appendix A.