Research papers

Intermittent dense water outflows under variable tidal forcing in Shark Bay, Western Australia

Yasha Hetzel a,b,*, Charitha Pattiaratchi a,b, Ryan Lowe b,c

a School of Environmental Systems Engineering (M015), The University of Western Australia, 35 Stirling Hwy, Crawley, WA 6009, Australia
b UWA Oceans Institute, The University of Western Australia, 35 Stirling Hwy, Crawley, WA 6009, Australia
c School of Earth and Environment (M004), The University of Western Australia, 35 Stirling Hwy, Crawley, WA 6009, Australia

A R T I C L E   I N F O

Article history:
Received 5 January 2012
Received in revised form 18 June 2013
Accepted 19 June 2013
Available online 29 June 2013

Keywords:
Gravity currents
Dense water outflow
Inverse estuary
Exchange
Flushing
Shark Bay

A B S T R A C T

Hydrodynamic data (time series of tidal velocities and vertical stratification) were collected during the winter of 2009 in Shark Bay, Western Australia, to document water exchange between the bay and the ocean. The net loss of freshwater through evaporation causes salinity levels in Shark Bay to be higher than the adjacent ocean, leading to its classification as an inverse estuary. The observations revealed pulses of near-bed dense water outflows (velocity $\geq 0.10 \text{ m s}^{-1}$) at weekly to fortnightly intervals, associated with periods of turbulent mixing when tidal velocities and winds were both weak. Although tidal mixing appeared to be the main control on the formation of the outflows, wind mixing during strong wind events was also sufficient to destratify the water column and interrupt the density-driven circulation. These data represent the first direct measurements of exchange flows in the entrance channels of Shark Bay and reveal a mechanism to maintain the balance of salinity as well as contribute to the exchange of material (e.g., larvae) between the bay and the ocean.

© 2013 Elsevier Ltd. All rights reserved.

1. Introduction

In arid climates, such as in Shark Bay Western Australia, density-driven currents can represent an important mechanism for estuary–ocean exchange. In these systems, an excess of evaporation and highly intermittent freshwater input causes the bay waters to become more saline than the adjacent shelf waters. The resulting density gradient drives a circulation that is the reverse, or inverse, of a classical estuarine system, with denser inshore waters flushing out of the bay along the bottom and fresher ocean waters flowing in at the surface (Pritchard, 1967; Nunes and Lennon, 1986; Bowers and Lennon, 1987; Lennon et al., 1987).

In these 'inverse estuary' systems the longitudinal density gradient, and thus density-driven circulation, is often weaker than in classic estuaries with freshwater input. This can increase their vulnerability to ecological problems such as hypoxia and accumulation of pollutants. Inverse estuaries are particularly susceptible to changes in the balance of buoyancy inputs, both naturally and due to human impacts; for example, due to droughts, changing climatic conditions, and the damming of rivers (Largier et al., 1997; Largier, 2010).

Globally, water bodies that exhibit inverse estuarine characteristics are common across a variety of scales and examples include: the Arabian Sea (Banse, 1997), northern Gulf of California, Mexico (Lavin et al., 1998), West Australian coastal waters (Shearman and Brink, 2010; Pattiaratchi et al., 2011), Spencer Gulf, Australia (Nunes et al., 1990); Shark Bay, Australia (Logan and Cebulski, 1970); Laguna San Ignacio, Mexico (Winant and de Velasco, 2003); Hervey Bay, Australia (Ribbe, 2006); Puttalam Bay, Sri Lanka (Arulananthan et al., 1995); and Bahia de Guaymas, Mexico (Valle-Levinson et al., 2001). Largier (2010) provides a review of 'low-inflow' estuaries that seasonally exhibit inverse estuarine circulation, and emphasizes that they are likely even more common than classic estuaries. Perhaps because the arid climates that induce inverse estuarine conditions occur to a large part in areas of low population density and/or in developing countries, there is a comparative void of literature on these systems.

In estuaries the gravitational circulation is driven by a longitudinal density gradient which provides a buoyancy flux for stratification of the water column that is then controlled by mixing (or destratification) effects due to turbulence generated by winds and tidal currents. Enhanced periods of gravitational circulation in inverse estuaries are associated with pulses of dense saline water flowing out of the estuary along the seabed, referred to here as 'dense water outflows'. It is usually during periods of reduced mixing that dense water outflows develop (Linden and Simpson, 2010).
Much knowledge of these processes has been gained through idealized laboratory experiments (e.g. Linden and Simpson, 1986, 1988) and through extensive observational studies in areas such as Spencer Gulf and Gulf St. Vincent in South Australia (Nunes and Lennon, 1986; Nunes and Lennon, 1987; Samarasinghe and Lennon, 1987). In these regions, the inshore salinity increases substantially during the summer due to intense evaporation, which is released as pulses of higher salinity water, particularly during autumn and winter when cooling further increases longitudinal density gradients. The frequency of these pulses are regulated by a fortnightly variation in tidal currents through the spring–neap cycle (Lennon et al., 1987).

In Shark Bay, Western Australia (hereinafter referred to as the Bay; Fig. 1) strong longitudinal salinity gradients have been documented since the 1970s, but the detailed dynamics of its density-driven circulation remain unknown. The ‘Shark Bay Outflow’ was inferred through measurements of sea floor sediments on the shelf (James et al., 1999), hydrographic survey data (Woo et al., 2006), and numerical modeling (Nahas et al., 2005), but little was known about the mechanism or frequency of the release of water from the Bay. In this paper, measurements of current velocity, density, and temperature profiles in one of the main entrance channels to Shark Bay are presented with the aim of describing the dynamics of the export of dense water from the Bay that have previously only been indirectly observed.

2. Study area

Climate, tides and winds are the major physical factors influencing the oceanographic environment of Shark Bay, a large, shallow, hypersaline Bay located along the arid northwest coast of Western Australia between 24 30’ and 26 45’S. Shark Bay is the largest semi-enclosed water body in Australia with a north–south length of ∼250 km, width ∼100 km, and average depth of only 10–20 m. The southern part of the Bay consists of an Eastern Gulf (Hopeless Reach) and Western Gulf (Freycinet Reach) divided by the Peron Peninsula (Fig. 1). Exchange with the Indian Ocean can occur only between the three offshore islands that semi-enclose the Bay. This study focuses on flow through the Naturaliste Channel (width ∼25 km), one of Shark Bay’s two main entrances. Other connections to shelf waters include the Geogrape Channel (width ∼35 km) to the north and much smaller South Passage (width ∼2 km) (Logan and Cebulski, 1970; Burling et al., 2003; Nahas et al., 2005). The slope of the seabed is quite steep through its main channels (Naturaliste and Geogrape) in contrast to the rest of the Bay, which is relatively flat and shallow. Shark Bay is a UNESCO World Heritage site. Here, some of the largest seagrass beds in the world, a diversity of marine mammals, and stromatolite ‘living fossils’ must coexist with a salt mining operation, an aquaculture industry, and important commercial prawn, scallop, snapper, crab, and whiting fisheries (DASET, 1990).

2.1. Climate and geography

Local annual rainfall is ∼200 mm, with most precipitation falling during the passage of winter cold fronts, or occasional summer cyclone events (Logan and Cebulski, 1970). Relatively strong southerly winds (mean ∼8 m s⁻¹) blow consistently during spring and summer (October–March) and are weaker (mean ∼5 m s⁻¹) and more variable during autumn and winter (April–September) (Pattiaratchi et al., 1997; Burling et al., 1999; Woo et al., 2006).

The combination of strong winds and high summer temperatures causes the mean annual evaporation (∼2000 mm) to exceed precipitation (∼200 mm) by a factor of 10, resulting in hypersaline conditions, particularly in the inner reaches. In the innermost portion of the Eastern Gulf, where Faure Sill restricts flow into Hamelin Pool (Fig. 1), the salinity level is ∼65 thus nearly twice that of the adjacent ocean (Burling et al., 1999). Two rivers occasionally empty into the Bay, but are thought to make an insignificant contribution to the mean salinity regime, as they are dry most years and flow only during cyclone events (Logan and Cebulski, 1970). Logan and Cebulski (1970) and later Smith and Atkinson (1983) found that the Bay maintained its salinity structure both seasonally and inter-annually, with generally vertically well-mixed conditions in most of the Bay.

An important feature of the Bay is the presence of density fronts in both Geogrape and Naturaliste Channels that separate Bay water from intruding shelf waters. A weaker frontal feature (Fig. 3b) extends across the entrances to both gulfs and northwards along the eastern shoreline (Logan and Cebulski, 1970; Smith and Atkinson, 1983; Nahas et al., 2005). Nahas et al. (2005) suggested that the location of the frontal systems at the ocean entrances are a result of the balance between tidal mixing and gravitational circulation and may be predicted using the Simpson–Hunter parameter (Simpson and Hunter, 1974). The front across the middle of the Bay is more influenced by wind. The well-defined semi-circular fronts in the Naturaliste and Geogrape Channels are easily identified in satellite sea surface temperature images throughout the year; however, differential heating/cooling of the Bay due to the shallow depths causes a seasonal reversal in temperatures across the fronts (Figs. 2 and 3). In winter, Shark Bay

![Fig. 1. Map of the study area showing the locations of the two moorings (MOOR 1 and MOOR 2) and the location of CTD transect T1 measured in June. The location of the transect measured in July (T2) was similar to T1, though slightly extended on either end. Depth contours at 5,10,15, and 25 m are shown.](image-url)
is cooler than the adjacent ocean whilst in summer it is warmer (Nahas et al., 2005). Part of the reason for such a strong temperature contrast between the Bay and ocean is the presence of the Leeuwin Current— an anomalous poleward-flowing eastern boundary current that brings warm, low salinity, nutrient-poor water south along the West Australian coast (Smith et al., 1991; Woo et al., 2006; Pattiaratchi and Woo, 2009).

2.2. Tides

In addition to buoyancy forces, tidal mixing plays a key role in controlling the formation of dense water outflows by acting against stratification. Shark Bay lies in a transition region along the Western Australia coast between macro (up to 11 m) semidiurnal tides (two cycles per day) in the far north and micro (~0.5 m) diurnal (one cycle per day) tides in the southwest (Pattiaratchi, 2011). The tides within the Bay, described by Burling et al. (2003), are mixed with a range of ~1–1.5 m. Maximum (predicted) tidal currents of ~0.5–1 m s⁻¹ occur near the ocean entrances and across Faure Sill (Fig. 1) (Nahas et al., 2005). The dimensions of the Eastern Gulf result in tidal resonance leading to different tidal regimes within the two gulfs. The Eastern Gulf experiences mainly semidiurnal (the $M_2$ amplitude is doubled due to resonance) tides and the Western Gulf (the focus of this study) is mainly diurnal (Burling et al., 2003).
The diurnal component of the tides is regulated by the tropical month (period of 27.3 days) with the declination of the moon above or below the equator controlling the tidal magnitude. When the moon is at maximum declination, diurnal forcing reaches a maximum resulting in tropic tides with a higher tidal range. When the moon crosses the equator, the diurnal frequency forcing is weakest and equatorial tides with a lower tidal range occur (O’Callaghan et al., 2010). In contrast, the semidiurnal tidal range is related to the lunar cycle that has a period (from new moon to new moon) of 29.5 days. Maximum semidiurnal tidal ranges occur close to the new and full moon, and minimum tidal ranges occur close to the first and last quarters—the spring–neap cycle (Pugh, 1987; O’Callaghan et al., 2010; Pattiaratchi, 2011). In Naturaliste Channel, the tide is dominated by the diurnal components and as a result the maximum tidal range coincides with the tropic tides and is mainly diurnal. Consequently, we will refer to the tropic–equatorial cycle in place of the spring–neap cycle.

2.3. Density-driven circulation

Historical observations of mostly well-mixed regions of Shark Bay concluded that tides and winds were the main drivers of its residual circulation. Logan and Cebulski (1970) were the first to suggest that exchange with the ocean may occur due to density-driven circulation but they stressed the well-mixed nature of the Bay as a potential inhibiting factor, and did not make any direct measurements. However, in an analysis of winter hydrographic survey data, Burling et al. (1999) described the existence of vertical stratification of the water column to the east of Cape Peron (Fig. 1) and noted that formation of dense water outflows was indeed possible. Further analytical studies by Burling et al. (1999) and numerical simulations by Nahas et al. (2005) revealed conditions favorable for the formation of dense outflows. Observations of sea floor sediment properties (James et al., 1999) and water mass characteristics (Woo et al., 2006) outside the Bay have also suggested the existence of what has been termed the ‘Shark Bay Outflow’. However, this was postulated to be more due to long-term average flows rather than a measureable instantaneous flow of saline bottom water. Until now, the absence of velocity measurements within the Bay has made it difficult to reach a conclusion as to the processes responsible for the release of salt from this inverse estuary.

3. Methods

3.1. Observations

The field experiment lasted for 28 days from 25 June to 23 July 2009. This ensured measurements over a complete tropic–equatorial cycle of the tides during the season when outflows were most likely to occur as a result of maximum horizontal density gradients across the Naturaliste Channel. The Leeuwin Current is stronger during the austral winter, thus the shelf waters are warmer and lower in salinity in contrast to the cooler and higher salinity Bay water. The placement of instrument moorings was chosen to best measure flows and temperature stratification in the vicinity of the channel and intruding density fronts, while avoiding areas where prawn and scallop trawl boats were licensed to operate.

The moorings were deployed in two locations (Fig. 1). MOOR 1 was inshore of the two stations (25°29.978’S, 113°15.070’E) and provided the majority of the data used in the analysis. It consisted of a 1200 kHz Teledyne RD Instruments Workhorse ADCP with a pressure sensor. This bottom-mounted ADCP sat flat (< 1° slope) in approximately 16 m of water and collected velocity profiles in 0.5 m bins sampling continuously every 1-s (1 Hz). To measure vertical temperature stratification, a chain of four Sea Bird Electronics SBE39 temperature recorders (thermists) was anchored alongside the ADCP and held vertical by a subsurface buoy. The thermists sampled every 2 min at heights of 1 m, 4.5 m, 7.5 m, and 13.5 m above the seabed. Both the top and bottom thermists had pressure sensors which enabled measurements of the water level as well as the degree of tilt due to drag on the mooring line.

MOOR 2 was located further offshore (25°25.206’S, 113°7.830’E) nearer to the gap between Dirk Hartog and Dorre Islands, in ~17 m of water. The sole instrument was an upward-looking 600 kHz RD Teledyne RD Instruments Workhorse ADCP measuring velocity profiles in 0.5 m bins and sampling at 1-min intervals. Although both moorings were placed in water of similar depths, MOOR 1 was located to the east (inshore) of a gently sloping north–south orientated ridge (height ~4 m) and MOOR 2 was located to the west (offshore) of this rise (Fig. 2).

During the deployment and retrieval of the two moorings, two 35–40 km long transects (Fig. 1) of density profiles were measured using a Sea Bird Electronics SBE 19–PLUS conductivity–temperature–depth (CTD) instrument. Profiles were recorded approximately every 2 km (more closely spaced across the frontal region) in a NW–SE direction from the center of Naturaliste Channel to near Cape Peron and passing by both MOOR 1 and MOOR 2. Transects were designed to capture the variation from the intruding oceanic water to Bay water. CTD transect 1 (T1, Fig. 1: Fig. 2a) was measured the same day instruments were deployed just after the passage of a strong winter cold front and several days of strong winds (up to 12 m s⁻¹), which varied in direction from the north to southwest. In contrast, CTD transect 2 (T2, Figs. 1 and 2b) was measured during a period of calm weather and was extended slightly on either end. To put the measurements into context and relate them to the location of the temperature front, we also examined corresponding MODIS satellite sea surface temperature (SST) data (Fig. 3) available from NASA’s Ocean Color website (http://oceancolor.gsfc.nasa.gov/). Hourly wind data (speed and direction) were obtained from an Australian Bureau of Meteorology weather station at Shark Bay Airport (Fig. 1).

3.2. Data analysis

ADCP data from both instruments were averaged into 5-min intervals after filtering any bad data. Good data were available from ~1.5 m above the seabed to within~1.5 m below the surface and rotated into the principal axes of current variance based on a principal component analysis of the depth-averaged time series (e. g. Emery and Thomson, 1997). All discussion of currents focuses on the flow along the major axis, with positive values denoting flow into the Bay. A short section of the time series (~8 h on 19 July) from the MOOR 2 dataset was omitted from analyses due to errors likely associated with the mooring rope becoming tangled around the ADCP.

Harmonic tidal analysis (e.g. Foreman, 1977) was performed using the T-Tide MATLAB toolbox (Pawlowicz et al., 2002) for current velocity and water levels with the MOOR 1 and MOOR 2 ADCP data. Inference was used to separate the S2/K2 and K1/P1 constituents based on parameters obtained from the TPXO7.2 tide model (Egbert and Erofeeva, 2002); however these constituents were minor contributors to the overall tidal signal. The significant constituents were then used to reconstitute year-long time series of velocity and water levels based on the full tidal signal (~17 constituents) as well as for only the semidiurnal and diurnal constituents. These diurnal and semidiurnal predicted time series enabled us to determine the individual contributions of the semidiurnal and diurnal tides.
Table 1

<table>
<thead>
<tr>
<th>Frequency</th>
<th>Depth</th>
<th>Surface</th>
<th>Mid</th>
<th>Bottom</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magnitude, $u$</td>
<td>Inclination</td>
<td>Phase, $\phi$</td>
<td>Magnitude, $u$</td>
<td>Inclination</td>
</tr>
<tr>
<td>Current velocity</td>
<td>$3.0$</td>
<td>$0.03$</td>
<td>$3.5$</td>
<td>$0.01$</td>
</tr>
<tr>
<td>MOOR 1, 24-day mean</td>
<td>$44$</td>
<td>$0.03$</td>
<td>$48$</td>
<td>$0.01$</td>
</tr>
<tr>
<td>Subtidal</td>
<td>$1$</td>
<td>$15$</td>
<td>$35$</td>
<td>$10$</td>
</tr>
<tr>
<td>Tidal</td>
<td>$1$</td>
<td>$15$</td>
<td>$35$</td>
<td>$10$</td>
</tr>
<tr>
<td>Mix</td>
<td>$1$</td>
<td>$15$</td>
<td>$35$</td>
<td>$10$</td>
</tr>
</tbody>
</table>

For all analyses, bottom currents were defined as those recorded in the first 4 bins from $-1.5$ to $-3.5$ m above the seabed and surface currents recorded $-1.5$ to $-3.5$ m below the surface. To demonstrate trajectories of the movement of water parcels near the surface and bottom over time, progressive vector diagrams were constructed by summing the instantaneous velocities recorded by the ADCPs. Subtidal currents were investigated by low-pass filtering the 1-h averaged ADCP time series using the PL64 filter (Beardsley et al., 1985) with a half-power period of 38 h. The filtered data were rotated to the principal axes of subtidal depth-averaged current variance based on a principal component analysis. Variability in the structure of the current profiles and vertical temperature stratification were compared with the local wind speed cubed ($W^3$) and bottom current speed cubed ($u_b^3$), as these are proportional to available mixing energy from the wind and tidal currents, respectively (e.g. Nunes and Lennon, 1987; Nunes Vaz et al., 1989; Simpson et al., 1990).

The balance between the major destratifying and stratifying influences was also explored following the approach of Nunes Vaz et al. (1989), Rippeth and Simpson (1996), and Nahas et al. (2005). This approach considers the destratifying influence of tidal and wind mixing and the stratifying influence of gravitational circulation (analogous to the dense water outflows) on the water column by considering the rate of energy imparted to a unit volume of water with units of $J m^{-3} s^{-1}$ (Nunes Vaz et al., 1989). The three dominant terms from Nahas et al. (2005) were used here:

$$\frac{4\epsilon \kappa_{DP} u_b^3}{3\pi \delta} + \frac{\delta \kappa_{DP} W^3}{\rho g \kappa_{mix}} = \frac{1}{320 \rho g h^2} \left( \frac{\delta h}{\kappa_{mix}} \right)^2$$

(1)

For this analysis, the following typical Shark Bay winter values (e.g. Nunes et al., 2005) were assumed: mean depth, $h=11$ m; air density, $\rho_a=1.2$ kg m$^{-3}$; mean seawater density, $\rho_w=1024.5$ kg m$^{-3}$; drag coefficient for bottom stresses, $\kappa_b=2.5 \times 10^{-3}$; tidal mixing efficiency, $\epsilon=3.7 \times 10^{-3}$; wind mixing efficiency, $\delta=3 \times 10^{-3}$; near-bed tidal velocity, $u_b$ (from observations) with $u_b(mean)=0.12$ m s$^{-1}$; wind speed, $W$ (from observations) with $W_{mean}=4$ m s$^{-1}$; vertical eddy diffusivity, $K_{mix}=1 \times 10^{-2}$ m$^2$ s$^{-1}$ (Nunes and Lennon, 1987); and drag coefficient for surface wind stress (e.g. Pugh, 1987), $\kappa_s=0.03(0.63 + 0.066 W^{1/3})/1000$.

The horizontal density gradient (from observations—see below) was $1.4$ kg m$^{-3}$ over a distance of $35$ km ($4 \times 10^{-5}$ kg m$^{-4}$)—this value covered the range from typical ocean water to Bay water over the CTD transect and is similar to larger scale horizontal density gradients from the ocean to the inner gulfs of Shark Bay described by Logan and Cebulski (1970). The gravitational circulation term in Eq. (1) was derived assuming that the vertical eddy diffusivity ($K_{mix}$) was independent of depth, a case only valid for reasonably well-mixed conditions (Nunes Vaz et al., 1989). Nahas et al. (2005) found this simplification to be valid for Shark Bay to describe the relative (order of magnitude) balance between mixing and stratification processes. Although the gravitational circulation term does not explicitly include a term for the surface buoyancy flux, this is accounted for in the gravitational circulation term through the horizontal density gradient. This assumption is justified as the longitudinal density gradient, dominated by salinity, over the length of the system remains quasi-steady in Shark Bay during winter over the monthly timescale examined here (Logan and Cebulski, 1970; Nahas et al., 2005). Nahas et al. (2005), Nunes Vaz et al. (1989), and Nunes and Lennon (1987) used the same mixing efficiency for tide ($\epsilon$) and wind ($\delta$), i.e. $\epsilon=\delta=3.7 \times 10^{-3}$. However, Nunes Vaz et al. (1989) noted that the correct value for wind mixing efficiency ($\delta$) is not clear and depends on the stratification. For this study, it was found that the value of the wind mixing efficiency ($\delta$) had to be reduced when...
compared to $\epsilon$ to explain changes to the observed stratification. However, the relative magnitude of the wind and tidal mixing terms remained within an order of magnitude despite subtle variations in the value of $\delta$ used.

Some caution should be used when interpreting the values for the balance between the destratifying and stratifying terms as they are sensitive to the mean water depth ($h$) chosen to represent the system (not a local value), with greater water depth decreasing the influence of wind and tidal mixing whilst strengthening the gravitational circulation. The gravitational circulation term, however, is most sensitive to changes in depth since it is represented in Eq. (1) as $h^4$. These limitations should not present a problem here, as the intention is to describe the relative balance between dominant processes across the system rather than exact values for a particular location.

4. Results

4.1. Hydrographic observations

Offshore SST, as shown by satellite imagery (Fig. 3a), was warmest at the beginning of the experiment (∼24°C) due to the stronger flow of the Leeuwin Current during winter (Godfrey and Ridgway, 1985). The warm water intruding into the Bay formed a temperature front near MOOR 1 as was observed in the CTD measurements (Fig. 2). The front would be transported with the flood and ebb currents and other external forcing (e.g. wind) but daily satellite SST images confirmed that it remained relatively close to MOOR 1 for the duration of the experiment. In July the decrease in SST of 1°C measured by the CTD occurred over the entire area, with an overall decrease of ∼2–3°C evident in the shallower inshore regions of the Bay (Fig. 3b). The offshore cooling was likely due to the seasonal weakening of the Leeuwin Current, while the temperature drop inside the Bay may be attributed to radiative cooling. This cooling of the inner Bay waters would have encouraged near-bed outflows by further increasing the density of the more saline inshore waters of the Bay.

The first CTD survey on 25 June 2009 (Fig. 2a–c) showed generally well-mixed conditions with a decrease in temperature ($\Delta T$=−2.9°C) and increase in salinity ($\Delta S$=0.6) from the channel into the Bay (toward the southeast) corresponding to a density increase of −1.25 kg m$^{-3}$. The maximum horizontal gradients were found around 5 km and 20 km from the start of the transect. We hypothesize that the generally lack of vertical stratification was due to mixing related to stronger winds and tidal currents that occurred during the previous week. When the transect was repeated (Fig. 2d–f), 28 days later on 23 July under calm weather conditions, a similar horizontal density gradient ($\Delta \rho=1.4$ kg m$^{-3}$) was measured but with a more defined front and some vertical stratification. Water temperatures both inshore and offshore, as shown by the measurements as well as the satellite data, were 1°C cooler by this time and a protruding tongue of colder more saline water extended along the bottom into the eastern portion of the transect near Cape Peron.

4.2. Current observations

4.2.1. Tidal variability

The major characteristics of the tidal currents are summarized in Table 1 for the dominant tidal constituents (K1, O1, M2, S2) at various depths. Tidal constituents for water levels are shown in Table 2 along with phase differences between currents and water levels (∼125°) that indicated the tide enters the Bay as a mixed progressive wave. The experiment occurred over three periods of tropic (spring) tides and two equatorial (neap) tides (Fig. 4a). The tropic stage in early July corresponded to lunar apogee and thus tidal forcing was substantially weaker during this period than at the start and end of the experiment.

The major axis currents at MOOR 1 (Fig. 4a) were aligned with the 101° (clockwise from north) axis and exhibited a maximum velocity of ∼0.4 m s$^{-1}$ during the ebb tide on 25 June. Mean depth-averaged current speeds over the experiment were 0.17 m s$^{-1}$. The tropic–equatorial variability of the tidal cycle was reflected in the current measurements, with intensified velocities recorded around the following dates: 25–30 June, 7–10 July, and 21–23 July. Velocity profiles exhibited only little vertical structure during these periods suggesting that tidal mixing was sufficient to produce well-mixed conditions. In contrast, during intervening (neap) periods two-layer flows developed and persisted for as long as 20 h with up to ∼4 m thick outflows along the bottom (e.g. on 29 June; Fig. 5). MOOR 2 (Fig. 4b) recorded stronger more unidirectional flow due to the restriction of water passing through the channel between Dirk Hartog and Dorre islands. The major axis of the depth-averaged flow was along the 103° (clockwise from north) axis, with a maximum depth-averaged major-axis velocity of ∼0.65 m s$^{-1}$ also during the ebb tide on 25 June. Mean depth-averaged current speeds at MOOR 2 were ∼0.22 m s$^{-1}$. In contrast to MOOR 1, there were no obvious periods of two-layer flows on an hourly timescale during periods of weak tidal forcing. The major differences in the flow characteristics recorded by the two instruments were likely related to bathymetric variations between the two sites and are more evident at subtidal frequencies described in Section 4.2.2.

For both MOOR 1 and MOOR 2, the form factor ($F_n$) for tidal currents (Table 1) indicated that the semidiurnal constituents were slightly dominant over the diurnal ($F_n=0.7$). In contrast the diurnal signal dominated the water levels ($F_n=1.1$; Table 2). The breakdown of the combined velocity signal into semidiurnal and diurnal velocities (Fig. 6) illustrates the contrasting influence of the slightly out-of-phase diurnal and semidiurnal forcing. During the experiment the maximum/minimum current speeds were slightly offset from periods of maximum/minimum tidal range. For this reason the tidal height range is not the best indicator of when tidal currents reach peak velocity.

4.2.2. Subtidal variability

Progressive vector diagrams (Fig. 7) confirmed that the periods of two-layer flow (e.g. Figs. 4 and 5) had influence over longer timescales. Over 28 days, residuals at both mooring locations were directed into the Bay at the surface and out of the Bay near the bottom. Although the trajectories for both sites had a southerly component rather than directly toward the open ocean, closer examination of the bathymetry at the sites indicated that the bottom flows would still be directed offshore under the force of gravity. This is consistent with residual flows patterns in previous numerical model results (e.g. Fig. 5b in Nahas et al., 2005). Both moorings indicated a similar pattern, mainly differing in the angle of the trajectories. The inner mooring (MOOR 1) showed a north–south component that was possibly related to topographic steering (Fig. 7b). Flow at MOOR 2 (Fig. 7a) was aligned more along the east–west axis, perhaps because there were no bathymetric obstructions on its offshore side, as there were for MOOR 1. For both moorings, the progressive movement of the bottom currents out of the Bay reached a maximum during periods of weaker tidal flows.

The inverse estuary two-layer flow regime was further confirmed by the orientation of the 28-day mean current velocity (Table 1) at MOOR 1, with mean surface currents directed to the NE (56°) and mean bottom current to the SSW (212°). At MOOR 2 the mean surface current flowed to the ESE (98°) and the mean bottom current to the SW (225°). In addition to topographic
steering, the southerly component of the offshore-directed mean bottom currents could be due to the origin of the water exiting the Bay at depth. Numerical modeling work by Nahas et al. (2005) and model results by the authors of the present study (unpublished) show near-bed flows originating in the Eastern Gulf, turning southwest around Cape Peron and passing near the moorings sites before exiting the Bay. Another possible source of dense water is the depression just south of MOOR 1 but the direction of the flow suggests that the Eastern Gulf is the more likely source.

The subtidal current data (Fig. 8) revealed that the two-layer flow regime identified in the progressive vector diagram (Fig. 7) and the 28-day mean (Table 1) was dominated by three ‘outflow events’ in June, on 6 July, and on 19 July 2009.

Table 2

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Amplitude (a_n) (m)</th>
<th>Phase (\phi_n) (deg)</th>
<th>(\Delta \phi) ((\phi_v - \phi_n))</th>
<th>Form factor ((F_n [(K1+O1)/ (M2+S2)])</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water level</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MOOR 1</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Diurnal—K1</td>
<td>0.20</td>
<td>300</td>
<td>133</td>
<td>1.10</td>
</tr>
<tr>
<td>Diurnal—O1</td>
<td>0.14</td>
<td>284</td>
<td>116</td>
<td></td>
</tr>
<tr>
<td>Semidiurnal—M2</td>
<td>0.21</td>
<td>325</td>
<td>131</td>
<td></td>
</tr>
<tr>
<td>Semidiurnal—S2</td>
<td>0.09</td>
<td>24</td>
<td>137</td>
<td></td>
</tr>
<tr>
<td>MOOR 2</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Diurnal—K1</td>
<td>0.20</td>
<td>293</td>
<td>130</td>
<td>1.14</td>
</tr>
<tr>
<td>Diurnal—O1</td>
<td>0.13</td>
<td>277</td>
<td>116</td>
<td></td>
</tr>
<tr>
<td>Semidiurnal—M2</td>
<td>0.20</td>
<td>306</td>
<td>119</td>
<td></td>
</tr>
<tr>
<td>Semidiurnal—S2</td>
<td>0.09</td>
<td>5</td>
<td>126</td>
<td></td>
</tr>
</tbody>
</table>

Fig. 4. Major axis current velocities (positive is flow into the Bay) for (a) MOOR 1 and (b) MOOR 2. Data has been averaged to 5 min intervals. The white areas represent data excluded from the analysis and the surface water level is shown with a black line.

Fig. 5. Major axis current velocities (positive is flow into the Bay) for MOOR 1 on 29 June showing two-layer flows persisting for over two tidal cycles. The zero velocity contour is shown as a thick black contour line. The white areas represent data excluded from the analysis and the surface water level is shown with a thin black line.
that lasted for 2–3 days with speeds \( \sim 0.10 \text{ m s}^{-1} \). The outflow events were defined as periods where vertical temperature stratification coincided with enhanced offshore-directed near-bed currents and inflow near the surface. These outflow events ((I),(II), and (III) in Fig. 8) are presented in detail below to describe the wind and tidal states operating during each event.

Fig. 6. Predicted time series of water levels (a, b) and bottom tidal current speeds cubed (c, d) based on based on constituents derived from harmonic analysis of ADCP current and water level data recorded at MOOR 1. (a) and (c) were predicted using the full tidal signal with the significant constituents from the t-tide analysis. Separate predictions of the water level (b) and currents based only on the significant diurnal constituents (red/thick line) and the significant semidiurnal constituents (in blue/thin line). The start and end of the experiment are shown with gray vertical lines and the phase and declination of the moon are shown above. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Fig. 7. Progressive vector diagrams for MOOR 2 (a) and MOOR 1 (b). Filled circles have been plotted for each day and labeled every 5 days. Bottom currents are shown in red and move toward the west while surface currents are shown in blue moving toward the east. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
More attention is given to site MOOR 1 as the ADCP data is more complete and the thermistor chain was moored here. Wind speed, and thus available turbulent mixing energy at the surface (expected to be proportional to $W^3$—Fig. 8a), varied throughout the experiment on a 4–7 day cycle due to the passage of weather systems. The tidal range (Fig. 8b) was a maximum around 25 June, 5 July, and 23 July and a minimum around 31 June and 13 July. The tidal current speed, and thus available tidal mixing (related to $|U_b|^3$ in Fig. 8c), was slightly offset from the periods of maximum tidal range (see Section 4.2.1). This in turn regulated the stratification of the water column as recorded by the thermistor chain (Fig. 8d). Three main outflow events may be identified at MOOR 1 in the major-axis subtidal velocity (Fig. 8e). One outflow event (II) was evident at MOOR 2 (Fig. 8f), which was located offshore of the small cross channel ridge (see Figs. 1 and 2). This provided some level of confidence that the dense water flows actually exited the Bay as depth increased offshore of MOOR 2, with no other topographic barriers. The formation, characteristics, and shut down of each of these outflow events are discussed below.

4.2.2.1. Outflow Event I (29–31 June). This outflow event at MOOR 1 was restricted to the bottom 4 m and was of short duration (∼32 h) but was the most intense of the three events, with a sustained outflow velocity of ∼0.10 m s$^{-1}$. Offshore-directed flow at MOOR 2 occurred on 28 June, but it was short-lived and a two-layer structure did not develop. This period was characterized by relatively weak tidal currents of order 0.18 m s$^{-1}$, which enabled stratification to develop near the seabed at MOOR 1 but stronger winds resulted in the surface mixed layer deepening to within ∼4 m of the bottom. This was observed in the vertical temperature time-series record where surface temperatures were markedly higher than bottom temperatures but at intermediate depths were similar to surface temperatures (Fig. 8d). The destratification coincided with a series of strong wind events (Fig. 8a) and, as the tidal currents remained weak, this suggested that the main cause for the breakdown of stratification was vertical mixing by the wind.

4.2.2.2. Outflow Event II (5–7 July). This outflow was the thickest (∼8 m) of the three events with offshore-directed currents ∼0.08 m s$^{-1}$, extending to almost 50% of the water depth at
MOOR 1 (Fig. 8e). The event was also clearly seen at MOOR 2 on the offshore side of the ridge with subtidal outflow velocities \(-0.08 \text{ m s}^{-1}\) near the bottom and offshore directed flow also covering more than half of the water column (Fig. 8f). During this period slight changes in mixing intensity caused formation/shutdown of the outflows. At MOOR 1 the water column became stratified for a period of \(-63 \text{ h}\) after several days of calm wind and weak tidal current conditions (Fig. 8a,c,d). The short-duration wind event on 6 July was insufficient to fully mix the water column but the outflow and stratification weakened over a period of several days when increasing wind and tidal currents were responsible for vertical mixing of the water column.

4.2.2.3. Outflow Event III (17–20 July). Subsequent to approximately 7 days of mixed conditions resulting from sustained winds and elevated tidal current speeds, the water column became stratified (Fig. 8d) when both winds and tidal currents weakened to the lowest levels recorded during the experiment (Fig. 8a,c). The coinciding bottom outflow at MOOR 1 had the longest duration of the three events observed (lasting \(-84 \text{ h}\)). During 19 July, a storm front with strong westerly winds up to 10 m s\(^{-1}\) mixed the water column and shut down the outflow. This wind event occurred just before tidal current velocities reached the highest levels measured during our field experiment (Fig. 8c). Unfortunately, the missing data at MOOR 2 restricted our ability to determine how strong the outflow was when it reached the offshore site.

Each of the three events described above had varying contributions from the destratifying forces of tide and wind. The result was a weekly to fortnightly cycle of weakening turbulence that caused the system to alternate between well-mixed to stratified conditions with the generation of dense water outflows. Although there was evidence that wind mixing was sufficient to mix the water column and shut down the dense water outflows, it was difficult to separate the effects of wind mixing due to the sporadic nature of the strong wind events and overlap with the more predictable cycle of tidal mixing. Our analyses have focused more on the effects of tidal mixing despite the periodic importance of wind mixing.

5. Discussion

Longitudinal density gradients in inverse estuaries are the driving force of near-bed dense currents that are modulated by tide and wind-generated turbulence. The majority of previous field studies conducted in inverse estuaries have focused on environments with a well-defined spring–neap tidal cycle (Lennon et al., 1987; Nunes and Lennon, 1987; Lavin et al., 1998). The circulation observed during this experiment in Shark Bay, while consistent with previous accounts in other inverse estuaries, indicated some differences in the timing of dense water outflows related to the different tidal forcing.

Field measurements of currents and stratification obtained in an entrance to Shark Bay showed dense water outflows up to 8 m thick (\(-50\% \text{ of water depth}\)) flowing out of the Bay during winter months when the semidiurnal and diurnal components of the tide were out of phase. The outflows occurred at weekly to fortnightly intervals when the winds and tidal currents were both weak.

Analysis of the field data can be extended to make predictions, based on the interactions between the diurnal and semidiurnal components, of when dense outflows are more likely to occur in Shark Bay. Fig. 9 extends Fig. 6 (described in Section 4.2.1) to show the predicted tidal forcing over an entire year. The cycle of weak/strong tidal currents that regulates the dense outflows repeats throughout the year such that periods of weak currents can occur during all seasons. However, the phase shift through the year of the semidiurnal and diurnal components (Fig. 9b,d) causes a seasonal variation in the tidal range (Fig. 9a) and current speeds (Fig. 9c) that is greatest \(-1–2 \text{ months}\) after the equinoxes and solstices when the semidiurnal and diurnal components are in-phase and out-of-phase. As a result, peak tidal currents occur around May and November when semidiurnal and diurnal constituents are in-phase and minimum currents occur around August and February when the constituents are out-of-phase. This suggests that tidal conditions around August and February are most favorable for dense water outflows. However, February experiences very strong southerly winds due to strong sea breezes and thus these dense outflows may not exist in February due to consistent turbulent wind mixing over the shallow depths of the Bay. For example, Pattiaratchi et al. (2011) found that along the south-west Australian inner shelf the strong winds vertically mixed the water column to depths greater than those existing in Shark Bay.

Estimates of the relative balance between the stratifying influence of gravitational circulation (dense water outflows) and the destratifying influence of wind and tidal mixing (Section 3.2) supported our observation that outflow formation was modulated primarily by tidal mixing although strong wind events could also inhibit their formation. Using mean values from the field experiment, Eq. (1) yielded the balance:

\[
\frac{6.3 \times 10^{-7}}{\text{Tidal mixing}} + \frac{5.7 \times 10^{-7}}{\text{Wind mixing}} \approx \frac{6.9 \times 10^{-7}}{\text{Gravitational circulation}} \quad \left[ \frac{\text{J m}^{-3} \text{s}^{-1}}{} \right]
\]  

The calculated wind and tidal mixing terms were of the same order of magnitude as the strength of the gravitational circulation. This implies that each of these processes are capable of destroying stratification induced by gravitational circulation if maintained for a sufficient period of time, or when the two processes act in unison. The field observations in Shark Bay support this view as do the findings of Nunes Vaz et al. (1989) for Spencer Gulf in South Australia. Nahas et al. (2005) found the destratifying influence of tidal mixing to be an order of magnitude higher than wind mixing in Shark Bay, but they used a mean tidal velocity that was several times greater than we observed (0.5 m s\(^{-1}\) > 0.12 m s\(^{-1}\)). This would have caused an over-estimation of the relative importance of tidal mixing compared to wind mixing. We hypothesize that the influences of tidal mixing and wind mixing are of similar importance for destratification, with the main difference being that strong wind mixing events are in general more sporadic and shorter in duration than the predictable periods of elevated levels of tidal mixing during the winter months.

This relationship is illustrated in Fig. 10, where the magnitudes of the wind and tidal mixing terms both deviate above and below the energy associated with typical mean gravitational circulation over the 28 day period. Periods of vertical stratification measured at MOOR 1 correspond to the days when both wind and tidal mixing drop below the dashed line representing gravitational circulation. The range of values for both wind and tidal mixing are within an order of magnitude of each other with tidal mixing values sustained for longer periods above and below the theoretical threshold of gravitational circulation. The magnitude of the wind mixing term does not vary quite as much and is harder to interpret, especially considering the uncertainties related to the value chosen for the wind mixing efficiency (\(\bar{\alpha}\)). It is clear that additional work needs to be undertaken to define the wind mixing efficiency, although this is unlikely to change the general balance described here. Locally, each of these three processes will vary with water depth. In shallow areas the balance will shift toward the destratifying processes (wind/tidal mixing) while deeper areas will favor stratification (gravitational circulation).

The results presented in this study move towards filling a major gap in the knowledge of the oceanography of Shark Bay. The dense
water outflows identified provide one mechanism to explain the Shark Bay Outflow and the data suggest that it is a regular occurrence on somewhat predictable intervals during the winter. The longitudinal density gradient that drives the benthic outflows is primarily due to the high salinity levels in the shallow inner regions. As a result, conditions are most favorable for outflow formation during the winter when cool water temperatures in the inner Bay further increase density gradients and winds are weakest.

In contrast, summer heating of surface layers would be less likely to lead to vertically stratified conditions due to the dominance of evaporation and wind mixing. Woo et al. (2006), however, also recorded traces of high salinity water outside the northern entrance to the Bay. In the summer, wind-driven circulation could act as the dominant summer exchange mechanism as has been previously proposed (Logan and Cebulski, 1970; Burling et al., 1999; Nahas et al., 2005). The stronger salinity signal recorded outside the northern entrance...
by Woo et al. (2006) supports this hypothesis that saline water would be advected northward out of the Bay by southerly winds. This question, however, is beyond the scope of this work, as we have focused on the inter-tidal timescales during the winter, at only one entrance.

The inhibiting effect of turbulence on the formation of subsurface dense water flows similar to those described here has been extensively documented in the laboratory (Linden and Simpson, 1986, 1988; Simpson and Linden, 1989) and to a lesser extent in the field. Comparisons between the results presented here and studies in the South Australian Gulfs (Lennon et al., 1987; Nunes and Lennon, 1987; Samarasinghe and Lennon, 1987; Nunes Vaz et al., 1989) and the upper Gulf of California, Mexico (Lavin et al., 1998), are most relevant to the present study, where a similar phenomenon occurs, albeit with subtle variations in timing due to differences in the mix of the important tidal constituents. Similar to our study in Shark Bay, investigators in the Gulf of California and South Australian studies found that near-bed outflows of similar magnitude form on a fortnightly frequency during neap tides and are inhibited during spring tides due to increased tidal mixing. If calm conditions coincide with this period of weak tidal currents, the dense water formed through evaporation and cooling in the head of the Gulfs is allowed to flow below the less-dense, lower salinity offshore water. This generally occurs during the autumn and winter when the density gradient is enhanced by cooling of Gulf waters. In Spencer Gulf, South Australia, the dense saline water flows at speeds ~0.10 m s⁻¹ when it reaches geostrophic equilibrium and is transported off the shelf (Lennon et al., 1987). The observed outflows in Shark Bay reached similar speeds at a weekly to fortnightly frequency during winter, but additional work is needed to discover how far the dense water outflows travel before they are mixed on the shelf and cease to flow.

In both the Gulf of California and the South Australian Gulfs there is a strong spring–neap tidal cycle coinciding with the phases of the moon. In the upper Gulf of California the mostly semi-diurnal tides can vary up to 5 m over the spring–neap cycle (Lavin and Marinone, 2003). In the South Australian Gulfs, the range variation is less but an anomalous near-equality of the lunar and solar semi-diurnal components causes an extreme modulation of the spring–neap cycle with tidal range/mixing dropping to near zero during the neaps, locally called the ‘dodge’ tide (Nunes and Lennon, 1986; Nunes et al., 1990). In contrast, the tides in Shark Bay are less extreme, and more mixed, highlighting the importance of the relationship between the diurnal and the semi-diurnal components. Despite this, the fundamental controlling mechanism remains the same and the result is a quasi-periodic outflow at a similar frequency to those observed in the South Australian Gulfs and the upper Gulf of California.

The explanation for the timing of the outflows presented here is somewhat simplified and does not explore the possibility that a certain balance between mixing and tidal advection might be required to transport the dense water out of depressions or over sills (e.g. Gordon et al., 2004). This could cause a lag between our simple prediction of when dense water outflows should occur and when they actually do occur. Our observations did not provide evidence for this, and a more extensive field campaign or numerical modeling would be needed to address this question.

The density-driven circulation in inverse estuaries such as Shark Bay is an important mechanism controlling the balance of salt and dispersing pollutants, as the dense outflows provide more effective transport than diffusive processes. In Shark Bay, the nearbed outflows observed are likely an important part of the delicate balance of the ecosystem, having both positive effects (i.e., maintaining salinity levels, transporting larvae, eggs, and nutrients within the Bay) and perhaps sometimes negative effects. For example, commercially fished wild scallops spawn during the winter months when dense outflows are most likely to occur. Hydrodynamic flushing of larvae has been proposed as an explanation for dramatic variability in stocks from year to year (Caputi et al., 1996; Kangas et al., 2006; Lenanton et al., 2009). If this is the case and larvae are lost from the system due to transport by dense water outflows, greater understanding of the dynamics of dense water outflows could have useful applications for management of the fishery. Knowledge of whether dense water outflows occur in the other main entrance and their seasonal variability would help us to understand their relative importance for circulation, Bayshelf exchange, and larval dispersal.

6. Conclusions

Field measurements of current velocities and stratification in one of the major entrances to Shark Bay, a large inverse estuary, showed that periodic stratification results in the formation of dense water outflows on a weekly to fortnightly frequency, during the winter. The dense water outflows occur when periods of weak tidal currents and wind speeds coincide, allowing buoyancy forcing caused by longitudinal density gradients to overcome the destabilizing effects of wind and tidal mixing. The processes described in Shark Bay are fundamentally the same as those that have been observed in similar inverse estuarine environments, although the timing of the outflow events in relation to tidal forcing is somewhat modified due to a different balance of tidal components. The dense outflows observed explain one mechanism for exchange, and give insight into the temporal variability of such outflow events.

Acknowledgments

The project was funded by a Fisheries Research and Development Corporation (FRDC) Grant (for the field investigations) and The University of Western Australia through the International Postgraduate Research Scholarship to YH. Special thanks to Peter Holtermann and Arnoldo Valle-Levinson who assisted with the field work, Ivan Haigh for help with the tidal analysis, and Mervi Kangas and Arani Chandrapavan at the Western Australian Department of Fisheries for their input on the biological implications of the work and the movements of the commercial trawl boats in the area. Finally, we would like to thank the anonymous reviewers whose detailed comments and suggestions provided valuable insight and contributed greatly to the final manuscript.

References

DASSET. 1990. Nomination of Shark Bay, Western Australia by the Government of Australia for inclusion in the world heritage list. The Department of the Arts, Sport, the Environment, Tourism.Territories.