Oceanic inflow from the Coral Sea into the Great Barrier Reef

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Long-term current meter data from the continental shelf region of the Great Barrier Reef show that there exists a zone of oceanic inflow onto the shelf. This oceanic inflow splits into two branches on meeting the continental shelf slope, resulting in two net longshore currents on the slope, one to the north and the other to the south of the separation point. In 1981 this separation point was located between 17°S and 18°S. This circulation was successfully predicted using a depth-averaged two-dimensional model in which the regional sea level gradient is explicitly added in the momentum equations. The resulting circulation on the continental shelf is controlled by an oceanic inflow of 0.58 Sv, spread over 500 km of the shelf edge both north and south of the separation point. The inflow appears measurably impeded by the presence of coral reefs, with >50% of the inflow occurring in a 150 km long area where reef density is small. Satellite images confirm this spatial variability. Longshore currents on the shelf generated by the inflow are modulated by the wind and tides, which can deflect the mean current away from areas of high reef density and generate localized outflows to the Coral Sea. Oceanic inflow is believed to be important because it flushes the shelf even in the absence of wind; it controls the dominant direction of across-shelf and along-shelf spread of spawn material from reefs; it makes it possible for upwelled water to spread quickly over the GBR shelf; it may also protect coral reefs by preventing river plumes from spreading onto the outer shelf.

Keywords: water circulation; shelf dynamics; mass transport; numerical model; East Australian Current; Great Barrier Reef; Australia.

Introduction

The Great Barrier Reef (GBR) is located on the continental shelf and slope of Australia’s northeastern coastline [Figure 1(a)], at the western margin of the Coral Sea. The net circulation over the GBR shelf appears to be controlled both by the local wind and by the circulation of the adjoining Coral Sea (Wolanski, 1994). The wind-driven circulation has received most attention from researchers focusing on the longshore currents. The cross-shore currents are mainly due to the oceanic influence and have received minimal attention. From geostrophic calculations from their CTD data as well as historical CTD data from Scully-Power (1973) and Church (1987), Andrews and Clegg (1989) found that an oceanic jet in the Coral Sea impinges on the shelf of the central region of the GBR, probably between 16°S and 19°S. They demonstrated also that this cross-shelf jet splits in two branches in the Coral Sea offshore from the continental shelf edge, one flowing north and the other flowing south from the bifurcation point. Details on the width of the jet and location of the bifurcation point were not known due to the paucity of data. Hughes (unpublished data) later modelled the circulation in the Coral Sea. Though the forcing at the open boundary conditions were poorly known, the model reproduced the general observations of Andrews and Clegg (1989); the impinging jet was found to be the South Equatorial Current (SEC). This current was predicted to split into two branches on meeting the continental slope. These branches are the northward flowing Hiri current, and the southward flowing East Australian Current (EAC), both of which are located offshore from the shelf edge. The transports of the Hiri current and the EAC have been estimated at 6 Sv northwards, and 6–14 Sv southwards, respectively. The EAC propagates along the continental shelf slope, with typical speeds of 0.3 m s\(^{-1}\) in the surface 200 m of outer shelf and shelf break waters of the central GBR (Church & Boland, 1983). Hughes (unpublished data) predicted that the bifurcation point of the SEC lies near 14°S. This location is now known, from geostrophic calculations from CTD observations, to...
migrate as far south as 20°S (Scully-Power, 1973; Church, 1987). Little is known if this bifurcating current also penetrates on to the continental shelf.

This circulation along the slope presumably also generates an inflow on the GBR shelf itself. The along-shelf component of this circulation is well known. It has been attributed to the EAC over the slope generating a longshore regional pressure gradient on the shelf (Middleton, 1987). In turn this surface slope induces a southward net flow on the GBR shelf, which displays strong inter-annual and seasonal fluctuations due to variability in the circulation in the Coral Sea (Wolanski & Pickard, 1985).

By contrast the cross-shelf component of this circulation remains unknown, but it is this component that controls the flushing of the GBR by the Coral Sea, and hence the residence time of water on the GBR, which is biologically important. In December 1995, near Lizard Island (14°7’S) Wolanski (pers. comm.) observed a net shoreward advection of coral eggs after mass spawning during slight southeasterly winds. This flow regime persisted with no tidal reversals in direction for 5 days, and provided no possibility of a seaward spread of coral eggs. In 1986 and 1987, a net shoreward spread of coral eggs, with occasional tidal reversals, was observed for several days after mass spawning near Bowden Reef (~19°S; see Wolanski et al., 1989; Oliver et al., 1992). On 19 February 2000, Wolanski (pers. comm.) observed a net onshore flow on the outer shelf northeast of Cairns at 16°7’S. This flow did not reverse with the tide for approximately 12 h, during which there was an absence of any significant wind forcing. This onshore flow protected the coral reefs by inhibiting the offshore spread of river plumes (King et al., 2001). Upwelling on the outer GBR is usually transient and geographically localized; the onshore inflow rapidly distributes the upwelled nutrients over most of the shelf width (Andrews & Gentien, 1982; Andrews, 1983b; Andrews & Furnas, 1986).

Low frequency flow on the GBR continental shelf is also highly modulated by the local wind over the shelf (Godfrey, 1973a, b; Wolanski & Bennet, 1983; Wolanski & Pickard, 1985; King & Wolanski, 1992). Andrews and Furnas (1986) suggested that tidal flows do not significantly contribute to long-term transport

Figure 1. (a) Locality map of central Great Barrier Reef. The model domain is indicated by the enclosed rectangle. Sites of field observations of wind, sea level and current used to force and calibrated the model are shown. (b) Model domain.
processes in the central region of the GBR. However, more recent studies found that the rugged topography generates large tidally averaged residual currents within the matrix of reefs (King & Wolanski, 1996; Wolanski & Spagnol, 2000). Thus, long-term, advective transport in the central GBR is controlled by the low frequency shelf currents generated by the combined influence of the wind, residual tidal currents and the Coral Sea circulation, including a zone of inflow.

In this paper, we use available current meter data to infer the oceanic inflow location and magnitude. We also calculate the resulting net circulation over the Great Barrier Reef. We calculate this effect using a numerical model where we also include the effect of the tides and the local wind. We show that the location of the oceanic inflow zone is largely controlled by the bathymetry and that this inflow controls much of the net circulation over the GBR. The predictions of net circulation on the GBR compare favourably with current meter observations.

**Materials and methods**

*Field data*

Observations of shelf currents and wind were obtained from Wolanski and Pickard (1985) who recorded data at temporal and spatial scales large enough to resolve the long-term, regional circulation features of the central GBR shelf. In particular, a subset of the Wolanski and Pickard data was used which spanned the austral winter from April to August 1981. During this period, currents were observed at northern, central, and southern locations within our study region at mooring sites near Lizard Island, Green Island, and Cape Upstart, respectively, while wind was recorded at Rib Reef (Figure 1(a)).

Observed currents and wind were decomposed into along- and across-shelf components. For currents, the along-shelf direction at each site was found from principal component analysis of the raw data and compared with those interpreted from navigational charts where topographical steering was thought to have affected the current direction. For wind, the mean along-shelf direction between Lizard Island and Bowen was used. Along-shelf currents and wind were then low pass filtered to remove fluctuations with periods less than 1.25 days (Godin, 1972). Time series plots of these data are shown in Figure 2.

The low-frequency longshore currents were strongly correlated with the low-frequency longshore wind component (Wolanski & Bennet, 1983; Wolanski & Pickard, 1985; Burragge et al., 1991). In addition, a net current also prevailed (Figure 2). At the Lizard Island mooring, the longshore flow was predominantly northward, with the magnitude of the flow modulated by the longshore southeast wind (Figure 2). At the Green Island mooring, the direction of the longshore base flow during low wind conditions was less clear, however during south-easterly wind events the flow was towards the north. Finally, at the Cape Upstart mooring, the base flow was southwards during low wind events and only reversed to northwards flow during strong south-easterly winds.

Regression analysis was used to approximate linear relationships between low-frequency, longshore wind at Rib Reef and low-frequency longshore currents at the three mooring sites. The results of this analysis are presented in Table 1. Correlation coefficients between wind and longshore currents were significant in all cases. The sign and magnitude of the constant term determined from the regression analysis (c in Table 1) revealed the direction and relative magnitude of the base flow generated by the regional sea level gradients due to the EAC (Godfrey, 1973b) in the absence of wind. From Table 1 it can be seen that a positive (northwards) base flow existed at the Lizard Island mooring, while negative (southwards) base flows existed at both the Green Island and Cape Upstart moorings. The magnitude of the flow at the Green Island mooring was approximately one fifth of that at the Cape Upstart mooring. The change of the longshore net current between the southern and northern locations indicated that an oceanic inflow had occurred in between these stations.

*Hydrodynamic model*

The depth-averaged, two-dimensional model of King and Wolanski (1996) was implemented to study the inflow from the Coral Sea and the resultant circulation throughout the central GBR. A depth-averaged, barotropic model was considered appropriate for this application as the shelf waters are generally well mixed throughout the year and flow is primarily horizontal (Wolanski, 1994). The model domain covered the continental shelf from Bowen, in the south, to Lizard Island in the north, and extended into the western Coral Sea (see Figure 1(b)). The computational grid for the model was a regular mesh of 2.0 x 2.0 km spatial resolution, and total grid dimensions of 343 x 138 computational points. The mesh size was sufficiently small to resolve the topographical influence of the reefs in the study region (Wolanski et al., 1989; Wolanski & King, 1990; King, 1992; King & Wolanski, 1996). Localized, reef-induced three-dimensional circulation features are known to
occur at much smaller scales (≈ hundreds of meters; Wolanski & Hamner, 1988; Deleersnijder et al., 1992; Wolanski et al., 1996) and were not resolved by this model. The grid was oriented with the x-axis directed in the mean along-shelf direction within the model domain (approximately 35° counter-clockwise from magnetic North).

To include the influence of regional sea level gradients, the elevation of the sea surface above mean sea level, \( \eta \), was defined as being composed of a mean elevation, \( \bar{\eta} \), plus a time varying fluctuation, \( \eta' \), about this mean:

\[
\eta = \bar{\eta} + \eta'
\]  

(1)

Decomposition of the sea surface elevation into mean and fluctuating components allowed \( \eta' \) to be used as the prognostic variable for sea surface elevation, and presented an opportunity for regional sea level gradients, \( \partial \eta / \partial x \) and \( \partial \eta / \partial y \), to be additional model forcing
Table 1. Linear regression analysis of the response of low-frequency, longshore currents, \( V_{Ls} \), to low-frequency, longshore wind at Rib Reef; \( W_{Ls} \). \( R \) is the correlation coefficient. Lag times (\( V_{Ls} \) lags \( W_{Ls} \)) were calculated from the lag frequency, longshore currents, \( V_{Ls} \), to low-frequency, longshore winds, \( W_{Ls} \).

<table>
<thead>
<tr>
<th>Location of mooring site</th>
<th>Linear regression ( V_{Ls} = m \times W_{Ls} + c )</th>
<th>Lag (hours)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lizard Island</td>
<td>( 1.92 \times 10^{-2} )</td>
<td>+0.053</td>
</tr>
<tr>
<td>Green Island</td>
<td>( 2.82 \times 10^{-2} )</td>
<td>-0.020</td>
</tr>
<tr>
<td>Cape Upstart</td>
<td>( 2.02 \times 10^{-2} )</td>
<td>-0.107</td>
</tr>
</tbody>
</table>

parameters, prescribed as independent values at each computational point and not just at the open boundaries.

Thus, the depth-averaged, barotropic equations of motion were represented as follows:

\[
\frac{\partial U H}{\partial t} + \frac{\partial U^2 H}{\partial x} + \frac{\partial U V H}{\partial y} - f V H + g H \left( \frac{\partial (\eta + \eta')}{\partial x} \right) = 0
\]

\[
+ \frac{g U |U|}{C^2} \frac{\tau_{xx}}{\rho} - \beta \nabla^2 (U H) = 0 \tag{2}
\]

\[
\frac{\partial V H}{\partial t} + \frac{\partial U V H}{\partial x} + \frac{\partial V^2 H}{\partial y} - f U H + g H \left( \frac{\partial (\eta + \eta')}{\partial y} \right) = 0
\]

\[
+ \frac{g V |U|}{C^2} \frac{\tau_{xy}}{\rho} - \beta \nabla^2 (V H) = 0 \tag{3}
\]

\[
\frac{\partial \eta}{\partial t} + \frac{\partial U H}{\partial x} + \frac{\partial V H}{\partial y} = 0 \tag{4}
\]

where \( t \) is time; \( x, y \) are the orthogonal horizontal axis; \( h \) is water depth below mean sea level; \( H \) is total water depth (\( h + \eta \)); \( U \) and \( V \) are depth averaged velocity components in the \( x \) and \( y \) directions, respectively; \( \mathbf{U} = (U, V) \) is the depth averaged horizontal velocity vector; \( f \) is the Coriolis parameter; \( g \) is acceleration due to gravity; \( C \) is the Chezy coefficient; \( C = H^{1/6}/n \) where \( n \) is the Manning coefficient; \( \tau_{xx} \) and \( \tau_{xy} \) are wind stress components in the \( x \) and \( y \) directions, respectively; \( \rho \) is fluid density; \( \beta \) is horizontal eddy viscosity; and \( \nabla^2 \) is the Laplacian operator \( \nabla^2 = (\partial^2 / \partial x^2 + \partial^2 / \partial y^2) \).

The preceding equations were discretized using a finite-difference approach and solved implicitly.

The open model boundaries to the northwest, northeast and southeast were forced by sea surface elevation, \( \eta \). At the closed coastal boundary and at island and emergent reef boundaries in the interior of the domain, zero transport perpendicular to the boundary was specified.

An examination of the tidal constituents along the northwestern boundary showed that the phase and amplitude of each constituent were relatively constant across the shelf from Cape Flattery (coastal) to Lizard Island (mid-shelf) and Carter Reef (shelf break). Therefore the forcing data for the entire northwestern boundary was determined by using observed tidal constituents at Lizard Island to define the sea surface elevation.

King and Wolanski (1996) showed that the tidal wave propagation in the southern region of the model domain is hampered by the dense reef matrix in this region, resulting in a large change in the phase of the tides across the reef matrix to the shelf edge [between Charity and Lynx reefs in Figure 1(a)], but only a small change in phase across the reef-free inner shelf. Based on this finding, the forcing data for the southeastern boundary was determined by linear interpolation of the sea surface elevation across the inner shelf, then from the reef matrix to the shelf edge. Observed tidal constituents from Charity Reef were applied to define the sea surface elevation at the eastern corner of the model grid. Tidal constituents for the coastal extremity of the southeastern boundary at Bowen were known. Sea surface elevation along the seaward boundary was determined by a linear interpolation between eastern extremities of the northwestern and southeastern boundaries.

Wind forcing for the entire model domain was determined by the observations from Rib Reef. This was considered appropriate as during the southeast trade the wind field in this region of the GBR is highly coherent over distances in excess of 1000 km (Wolanski, 1982).

Tidal and wind data were easily incorporated into the model, and the model itself, given an adequate representation of the bathymetry, determined reef/tidal current interactions and residual currents (King & Wolanski, 1996). Regional sea level gradients, however, were not known explicitly. Such gradients cannot be measured using tide gauge data due to the large variability in both strength and location. Satellite altimetry can provide difference in mean sea level height at regional scales, but the resolution of present sensors is insufficient for the Great Barrier Reef (C. Koblinsky, pers. comm.). Instead, these slopes were inferred as follows. Analysis of the observed low frequency, longshore currents presented in Table 1 identified at least two distinct regions with net flows in opposite directions: a northward net flow in the
northern region of the model domain; and a southward net flow in the southern region of the model domain. The change in sign of the net flow suggested that oceanic inflow occurred between latitude 14.7°S and 16.75°S, which is consistent with the known meandering of the SEC bifurcation point (Church, 1987). This net flow regime was represented in the model by three regions with differing sea level gradients: regions of northwards and southwards flow to the north and south of Green Island, respectively, were separated by a region with zero sea surface gradient in the vicinity of Green Island. These regions are schematically represented in Figure 3. The region with zero sea surface gradients on the shelf represented the area immediately inshore of the bifurcation point on the continental slope, and defined a zone of inflow from the Coral Sea from the observations of Wolanski (pers. comm.). Forcing by regional sea surface gradients was only applied on the shelf where bottom friction maintains numerical stability.

Figure 3. Schematic representation of the regional sea surface gradients used to force the model. Transects 1–6 show locations of transects for transport calculations. Sea surface gradients in the northern (southern) region of the model domain drive water northward (southward).
The magnitudes of the sea level gradients in each region, relative to each other, were obtained from the constant term from the regression analysis (Table 1). The absolute values of the gradients, however, were determined by 'trial and error', whereby the magnitude of the regional gradients, and their extent and location, were adjusted until there was adequate agreement between predicted and observed low frequency currents at the three mooring sites.

The introduction of regional sea level gradients as a forcing mechanism in the model necessitated the adjustment of the sea surface elevation boundary forcing data to maintain sea surface continuity at the model boundaries. The magnitude of regional longshore and cross-shelf gradients were used to determine the sea level offsets at the relevant model boundaries, given their distance from points of inflection of the relevant gradients. These offsets were then added to the relevant boundary elevation data.

The model was run 100 days from 0000 h, 12 April 1981, with a time step between prognostic calculations of 3 min. Calibration of the model involved tuning the location, and magnitude of the three regional sea level gradients until a satisfactory comparison was reached between observed and predicted low frequency, longshore currents at the Lizard Island, Green Island and Cape Upstart mooring sites. The model was then run for the four forcing scenarios: Run no. 1 was forced only by wind and tidal elevation at the boundaries; Run no. 2 was forced by the wind, tides and regional sea surface gradients as outlined in Figure 3, and will be hereafter referred to as ‘net’ forcing; Run no. 3 was forced only by tidal elevation and regional sea surface gradients; Run no. 4 forcing was similar to Run no. 2 except that the magnitude of regional sea surface gradients was doubled.

### Results

Low-frequency, longshore currents predicted by the model at the mooring sites near Lizard Island, Green Island and Cape Upstart were compared with field observations at these locations. This comparison is shown in Figure 2 and Table 2. Computed currents for Run no. 1, with no forcing by regional sea level gradients, significantly under-predicted the observed southerly flow at the Cape Upstart mooring [Figure 4(a)]. Indeed, the predicted flow is almost entirely northward, in response only to wind forcing [see also Figure 5(a)]. At the Green Island and Lizard Island mooring sites, the divergence of the predicted flow from the field data was less obvious, given the variability in the field data.

Additional forcing by regional sea surface gradients was introduced in Run no. 2, and predicted net low-frequency, longshore currents compared qualitatively well with field observations. The improvement in the predictions is most obvious at the Cape Upstart mooring site [see also Table 2 and Figure 4(b)]. Predicted base longshore currents at the Cape Upstart and Green Island mooring sites showed net increases, compared to Run no. 1, of 0·059 m s\(^{-1}\) and 0·027 m s\(^{-1}\), respectively, in the southerly direction, averaged over the 100 day model run. At the Lizard Island mooring, there was an average increase of 0·032 m s\(^{-1}\) in the northwards base longshore flow. The net flow field was clearly controlled by the oceanic inflow [Figure 5(b)]. Doubling the regional sea level gradient (compare Run nos 2 and 4) improved the predictions in the southern region [Figure 2(c)] but degraded the model in the central region [Figure 2(b); see also Table 2].

Given the limitations of the data, Run no. 2 was taken as an adequate calibration of the model forced by regional sea surface gradients for the investigation of the inflow from the Coral Sea and the net transport that this inflow drives.

Cumulative volume fluxes across five cross-shelf, and 1 longshore transect were calculated to examine the long-term transport of water mass on the GBR shelf. The five cross-shelf transects, shown in Figure 3 as transects (1)–(5), extended from the coastal

### Table 2. Observed and predicted base longshore currents (m s\(^{-1}\)) at mooring sites on the GBR continental shelf for 100 days starting 12 April 1981. Base flow was determined from linear regression analysis of the response of observed and predicted low frequency, longshore currents to low frequency wind. A positive (negative) longshore flow indicates flow to the north (south)

<table>
<thead>
<tr>
<th>Current meter</th>
<th>Observed</th>
<th>Run no. 1</th>
<th>Run no. 2</th>
<th>Run no. 3</th>
<th>Run no. 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lizard Island</td>
<td>0·053</td>
<td>0·009</td>
<td>0·041</td>
<td>0·011</td>
<td>0·008</td>
</tr>
<tr>
<td>Green Island</td>
<td>0·020</td>
<td>0·036</td>
<td>0·024</td>
<td>0·048</td>
<td>0·024</td>
</tr>
<tr>
<td>Cape Upstart</td>
<td>0·107</td>
<td>0·008</td>
<td>0·051</td>
<td>0·052</td>
<td>0·137</td>
</tr>
</tbody>
</table>
boundary to the shelf edge, which was taken as the 100 m contour. The shore-parallel transect (6) extended along the 100 m contour across the defined ‘inflow’ region with zero longshore sea surface gradient.

As can be seen in Figure 5(b) and Table 2, there was a net southward transport south of transect (3) and a net northward transport, north of transect 4. The bifurcation region between transects 3 and 4 was characterized by negligible net longshore transport. Average transports of 0·22 Sv to the south and 0·05 Sv to the north were calculated across transects (2) and (5), respectively, and there was a constant transport of 0·048 Sv into the GBR from the Coral Sea across the shore-parallel transect (6). The magnitude of the inflow across transect (6) was an order of magnitude smaller than the fluxes calculated across cross-shelf transects (2) and (5), indicating that oceanic inflow occurs over a much longer stretch of the shelf edge.

The spatial distribution of the cross-shelf oceanic inflow into the GBR shelf is shown in Figure 6. The largest inflow occurred between model longshore grid positions 50 and 130, roughly between Rib and Bowden reefs. In this region the density of coral reefs on the outer shelf is the least sparse, and thus there is little topographical impediment to oceanic inflow. However, south of this location, the density of coral reefs is much higher, and is known to impede exchange between the inner and outer shelf (King & Wolanski, 1996). Note that just upstream of Old Reef the model predicted a small zone of outflow from the GBR into the Coral Sea; this effect is due to steering by coral reef assemblages and was previously suggested by Spagnol et al. (2001). North from 15·5°S, the outer edge of the GBR shelf is dominated by a series of very dense Ribbon Reefs separated by deep narrow channels. Cross-shelf edge exchanges in this region are limited to these deep channels, which occupy only about 10% of the along-shelf distance. The existence of this barrier apparently explains the small mass flux calculated for the regions north of the prescribed inflow zone.

From the inflow profiles shown in Figure 6, the volume transport into the region of the central GBR represented by the model was estimated to be ~0·58 Sv for Run no. 2. For comparison, volume transport for Run nos 1, 3 and 4 were estimated at ~0·008 Sv, ~0·56 Sv and ~1·10 Sv respectively.

Synoptic distributions of net currents are shown in Figure 7. As can be seen from Figure 7(a), the average current over the 100 days of Run no. 1 showed an unrealistic, northward flow over the entire shelf. With the net forcing applied in Run no. 2, the 100 day averaged currents [Figure 7(b)] showed a northwards flow in the northern region of our domain, and a southwards flow seaward of the mid shelf in the southern region. In the shallower inner shelf of the southern region, average flows were northwards, with a smaller magnitude than the southwards flow observed offshore. This result was consistent with the ‘decoupling’ of inner shelf currents from the mid to outer shelf currents, which isolates inner from outer shelf waters (King & Wolanski, 1992).

As described earlier, the model boundaries are forced by sea surface elevation which is adjusted according to the magnitude of regional longshore and cross-shelf sea surface gradients to maintain sea surface continuity at the boundaries. However, the flux through the boundaries is free to vary in response
to the circulation within the model domain. A comparison of Figures 7(a) and (b) reveals the changes in the net circulation near the continental margin in the Coral Sea in response to the forcing regimes used in Run nos 1 and 2. Adjustment of the boundary forcing data for Run no. 2 [Figure 7(b)] resulted in a circulation in the far western Coral Sea which was in qualitative agreement with that described by Andrews and Clegg (1989), and Hughes (unpublished data): a westward flowing oceanic jet that meets the shelf at \( \sim 16^\circ \text{S} \), and splits into northward and southward currents flowing parallel to the shelf.

Figures 7(c) and (d) show synoptic views of net currents for the case of strong southeast trade winds and calm weather, respectively. The net southward flow on the shelf prevailed only in the southernmost region during strong winds, but extended to over half of the domain in calm weather.

**Discussion**

Low frequency currents on the shelf are generated by the oceanic inflow of Coral Sea water onto the GBR shelf (Godfrey, 1973b), by the wind (King & Wolanski, 1992), and by non-tidal residual currents in the reef matrix (King & Wolanski, 1996; Spagnol et al., 2001). The net effect of these three distinctly different forcing mechanisms on the circulation of the central GBR shelf was investigated through the development of a numerical model. Forcing by winds and tidal elevation at the boundaries was readily incorporated in the model using traditional techniques. The influence of the oceanic inflow was parameterized in the model by incorporating regional sea surface gradients as an additional forcing mechanism at each computational point and not just at the open boundaries as was done by King and Wolanski (1992) who were, as a result, unable to force the model simultaneously with wind, tides and oceanic forcing.

The location and direction of regional sea surface gradients were estimated from an analysis of existing long-term current observations. The magnitudes of the sea surface gradients were calibrated by matching observed and predicted low-frequency, longshore currents at three spatially distinct mooring sites. Thus, the location and magnitude of the sea surface gradients were not precisely determined, which introduces some level of uncertainty into the predicted circulation and magnitude of the resultant oceanic inflow onto the GBR. Notwithstanding, the relative agreement of observed and predicted flows at the Lizard Island, Green Island and Cape Upstart mooring sites for model Run no. 2 [Figures 2 and 4(b)] does give confidence to the estimated magnitudes of both the sea surface gradients and the oceanic inflow.

Sea surface gradients of approximately \( 7.2 \times 10^{-8} \) and \( 6.9 \times 10^{-8} \) prevailed in the northern and southern regions of the model domain, respectively, during the period of the model run. The resultant circulation produced a transport of approximately 0.58 Sv from the Coral Sea into the central section of the GBR. This estimate is approximately one order of magnitude higher than previous estimates for the oceanic inflow into the GBR.
magnitude less than estimates of the southwards transport of 6–14 Sv off the shelf (Church, 1987), which is realistic given the existence of the reef matrix which inhibits cross-shelf exchange (Wolanski & Bennet, 1983). Variability in the exchange volumes due to variability of the Coral Sea circulation can be estimated from comparison of the predicted inflows for Run nos 1, 2 and 4. For these three runs, each with significantly different forcing parameters, the magnitude of predicted low frequency, longshore flow on the shelf at the three mooring sites remained within the limits of the field data, indicating that the transport of the inflow may indeed range from being negligible, to exceeding 1 Sv. Our understanding of this variability is not clear from the present study and requires further investigation in future studies.

Analysis of the along- and across-shelf fluxes throughout the central GBR revealed that the flux in the southern region could not be reconciled against the magnitude of the volume flux from the Coral sea in the region of the bifurcation. Estimates of the spatial characteristics of the cross-shelf edge flux show that more than 50% of the inflow occurs through a region of low reef density between Rib and Bowden reefs.

Andrews (1983b) reported observations of sustained inflow near Rib Reef, from current meter records spanning from October 1980 to June 1982. Mean flow through the reef matrix near Glow and Needle Reefs on the outer shelf exhibited a persistent onshore component, while flow near John Brewer Reef, on the mid-shelf, was directed more along-shelf (Table 3). Observed mean currents on the shelf during winter were an equivalent order of magnitude to those computed in Run no. 2. Andrews (1983b) argued that the onshore flow reflects the intrusion of upwelled water across the shelf. Upwelling of water from the deeper continental slope on the outer GBR, although generally transient and localized, has been shown to span hundreds of kilometres along the shelf break (Andrews, 1983a), with observed onshore advective speeds of up to 14–21 cm s⁻¹ on the mid to outer shelf (Andrews & Furnas, 1986). However, the present study indicates that the onshore flow is also a result of, and may in fact be dominated by, oceanic inflow due to the larger scale circulation in the Coral Sea. Table 3 also shows cross-shelf flow calculated from the trajectory of 2 Lagrangian drifters released from near Lizard Island in December 1995, and tracked for 3 days (Wolanski, unpublished data). These observations show the influence of oceanic inflow in the northern region of the central GBR, however at a smaller magnitude than those observed further south by Andrews (1983a), due to the existence of the dense Ribbon Reefs in the northern GBR. The present study shows that net circulation within the central GBR would rapidly advect nutrient-enhanced upwelled water over the GBR shelf.

Oceanic inflow onto the central GBR is clearly evident in a satellite image of the chlorophyll concentration in this region at 01:43:51, 14 July 2000 GTM (Figure 8). For interpretation of Figure 8 it should be noted that lighter areas are chlorophyll enhanced or
turbid water generally associated with shelf and coastal regions; darker areas indicate water of oceanic origin with low chlorophyll concentrations. Darker ‘plumes’ that penetrate shoreward from the shelf edge indicate zones of oceanic inflow. In the northern region of the central GBR, north of approximately 16°S, inflows are limited to very distinct channels between the otherwise highly dense reef matrices. South of 16°S, where the spatial density of the reef matrix decreases, oceanic inflows become more spatially frequent, with a major zone of inflow evident between Rib Reef and Bowden Reef. Further south from Bowden Reef there is minimal inflow across the outer shelf inflow due to the presence of the dense reef matrix on the outer shelf. On the mid shelf however, oceanic water originating from inflows further north appears as a dark plume with distinct frontal features. A second image collected at 02:06:10, 15 July 2000 GMT (not shown) revealed that the oceanic/coastal-water front south of Bowden Reef had penetrated approximately 16·5 km further southwards, indicating an average speed of approximately 0·18 m s⁻¹. Wind speeds recorded nearby at an automatic weather station on Davies Reef during the interval between the two images were generally light (mean 3·38 m s⁻¹, minimum 0·00 m s⁻¹, maximum 5·58 m s⁻¹) and direction was from the Northeast to Southeast quadrants. The magnitude of the shelf flow inferred from the satellite images is consistent with that observed at Cape Upstart for periods of similar meteorological conditions during the austral winter of 1981, and also with flows predicted by the model in the present study [Figure 2(c); see also Table 3]. Further, the spatial variability of the observed oceanic inflows is consistent with that revealed by the numerical model [Figures 6 and 7(b), respectively].

It is also noted from Figure 8 that there can exist quite distinct cross-shelf fronts between coastal waters and waters of oceanic origin. This observation again suggests that there is little across shelf mixing in this region which supports the findings of both this present study, and those of King and Wolanski (1992). Oceanic inflows appear to isolate the mid-shelf and outer-shelf from inner shelf waters by restricting the offshore movement of water on the inner shelf. As a consequence, reefs within inflow zones on the mid and outer shelf may be more protected from the pollutant threats originating from inner shelf waters.

The complex topography of the GBR introduces major spatial and temporal variability in the net circulation of this region (King & Wolanski, 1992, 1996; Wolanski & Spagnol, 2000; Spagnol et al., 2001). The findings of this study also demonstrate that the spatial variability of the topography also influences the location and magnitude of exchange between the
Observed and computed longshore and cross-shelf currents (m s⁻¹). Observed mean currents at Needle Reef, Glow Reef and John Brewer Reef moorings are from Andrews (1983b). Observed cross-shelf current at Lizard Island is the mean current calculated from Lagrangian drifter data (Wolanski, unpubl. data). Longshore flow on the southern shelf was estimated from satellite imagery (see Figure 8). Computed mean currents for Run no. 2 are the average of 100 days starting 12 April 1981. Positive (negative) longshore flow indicates flow to the north (south), and positive (negative) cross-shelf flow indicates flow directed onshore (offshore).

<table>
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Coral Sea and GBR, due to regional forcing from the EAC.

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References


Figure 8. SeaWiFS satellite image of chlorophyll a concentrations in the central Great Barrier Reef, at 01:43:51, 14 July 2000 GMT. Darker areas indicate water of oceanic origin with low chlorophyll concentrations. Lighter areas are either chlorophyll rich shelf waters or case-2 coastal water. The solid black contour line delineates waters of coastal and oceanic origin. Black areas are land or cloud. The model domain of the present study is indicated by the enclosed rectangle. The image was received and processed by the Australian Institute of Marine Science, courtesy of Orbimage.


