OCEANOGRAPHIC PROCESSES ON THE NINGALOO CORAL REEF, WESTERN AUSTRALIA

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CHAPTER 1 INTRODUCTION

The Ningaloo Reef tract lies on the western coast of Australia between 21°47' and 24°S and along longitude 113°30'E (Figures 1.1 and 1.2). The reef tract runs parallel to the coastline from Gnarraloo Bay to Point Murat, a distance of some 280 km (Figure 1.3) and consists of a barrier reef 1-6 km offshore (average 2.5 km), backed by a shallow, sedimentary lagoon (mean depth at AHD of about 2 m) with occasional patch and nearshore platform reefs.

Coral reefs occur along the central and northern coast of Western Australia, but ecological and physical studies have only been carried out in the Dampier Archipelago about 300 km northeast of North West Cape, and at the Abrolhos Islands approximately 800 km south of North West Cape (e.g. Chittleborough 1983; Hatcher and Walker 1984).

Extensive coral reef development does not occur south of the Abrolhos Islands, although a small coral reef occurs at Rottnest Island, and reef building corals are found as far south as Esperance (Figure 1.1).

The Ningaloo Reef tract is the largest fringing coral reef in Australia, providing easy access for visitors to experience a rich diversity of marine life. Increased recreational usage in recent years has led to a proposal to establish a marine park under the management of the Department of Conservation and Land Management, to ensure the sustainability of these resources. Research is needed to develop the knowledge upon which effective management can be based. Effective management is dependent upon an understanding of the key ecological processes of the system, which relies in part on a basic understanding of

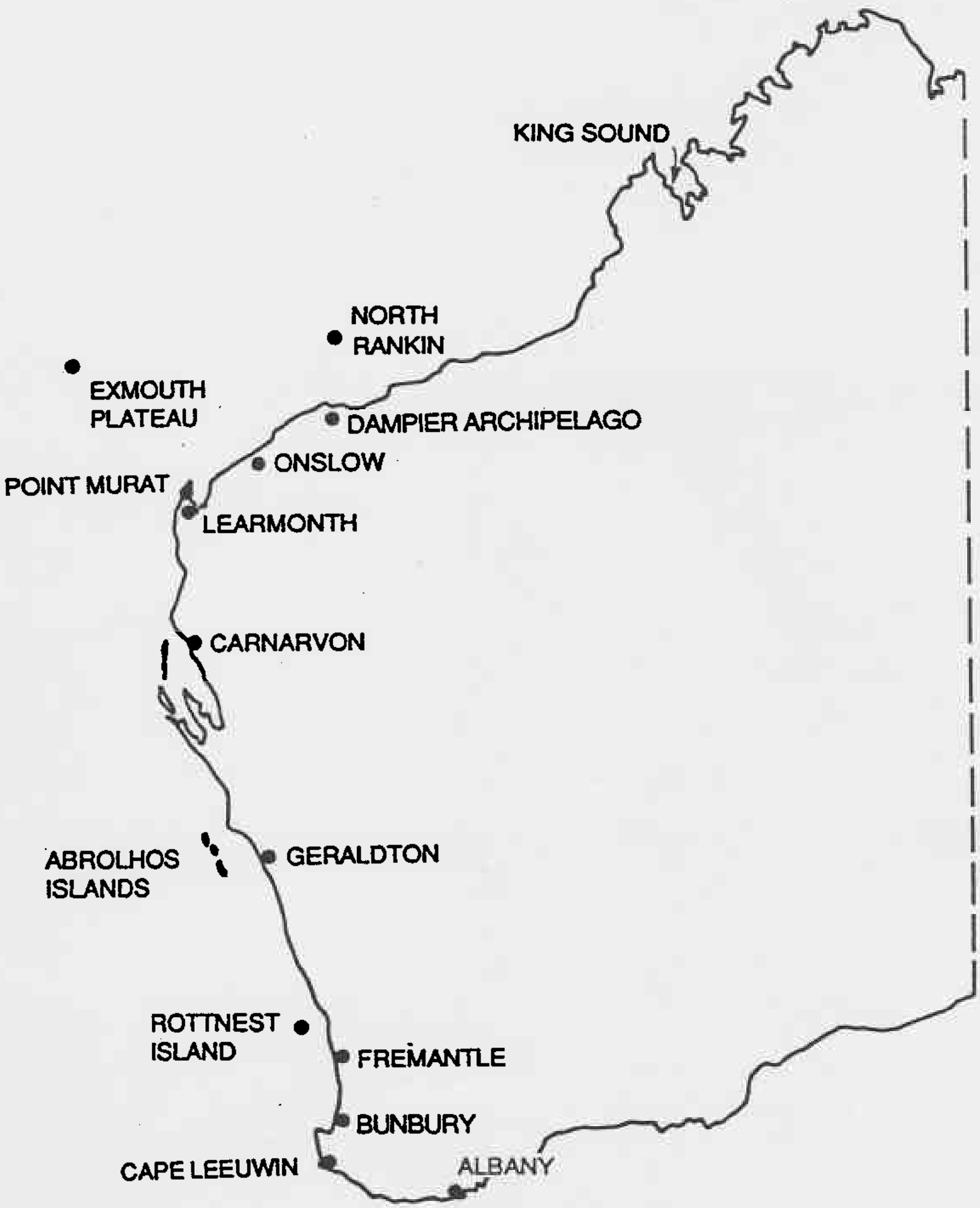

KING SOUND
NORTH
RANKIN
EXMOUTH
PLATEAU
DAMPIER ARCHIPELAGO
ONSLOW
POINT MURAT
LEARMONTH
CARNARVON
ABROLHOS
ISLANDS
GERALDTON
ROTTNEST
ISLAND
FREMANTLE
BUNBURY
CAPE LEEUWIN
ALBANY

Figure 1.1

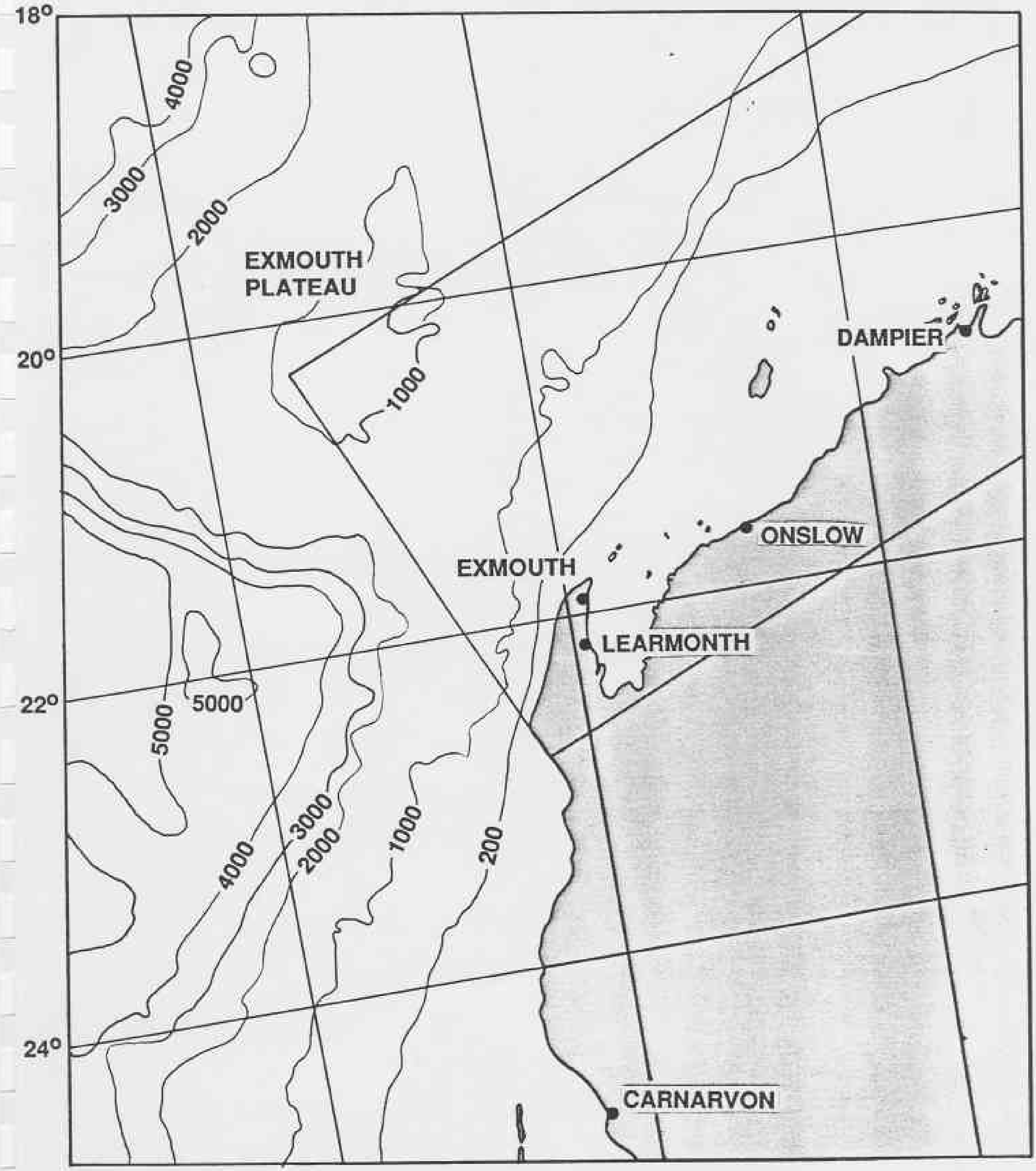

18°
20°
22°
24°
4000
3000
2000
1000
EXMOUTH
PLATEAU
5000
5000
5000
4000
3000
2000
1000
200
DAMPIER
ONSLOW
EXMOUTH
LEARMONTH
CARNARVON

Figure 1.2

the physical environment. So far, the Ningaloo Reef tract has been the subject of only very limited studies, described in the reports of Meagher (1980) and May et al. (1983). More recently Simpson and Masini (1986) have collected some tidal and seawater temperature data at a number of stations within the lagoon (Figure 1.3). These are particularly relevant to the prediction of the timing of mass spawning of scleractinian corals (Simpson 1985; 1986). They have also emphasised the general need to assemble a basic understanding of the physical processes which operate within the reef and lagoon.

This report attempts to lay a foundation for a future programme of physical studies by using existing knowledge of processes on coral reefs and applying this specifically to the Ningaloo Reef tract. It also draws on regional oceanographic information, much of which comes from the studies conducted on the North West Shelf by companies involved in oil and gas exploration and extraction. The report identifies areas of physical research which will contribute to an understanding of the structure and function of the Ningaloo Reef ecosystem.

Figure 1.2 shows the regional bathymetry of the continental shelf near North West Cape. The shelf (defined as the region between the coast and the 200 m contour) becomes very narrow between North West Cape and Point Cloates. The width of the shelf over this piece of straight coastline is about 10 km, bringing the shelf-break closer to the coast than at any other location in Australia. South of Point Cloates the orientation of the coast changes so that the width of the shelf increases to about 50 km at Amherst Point.

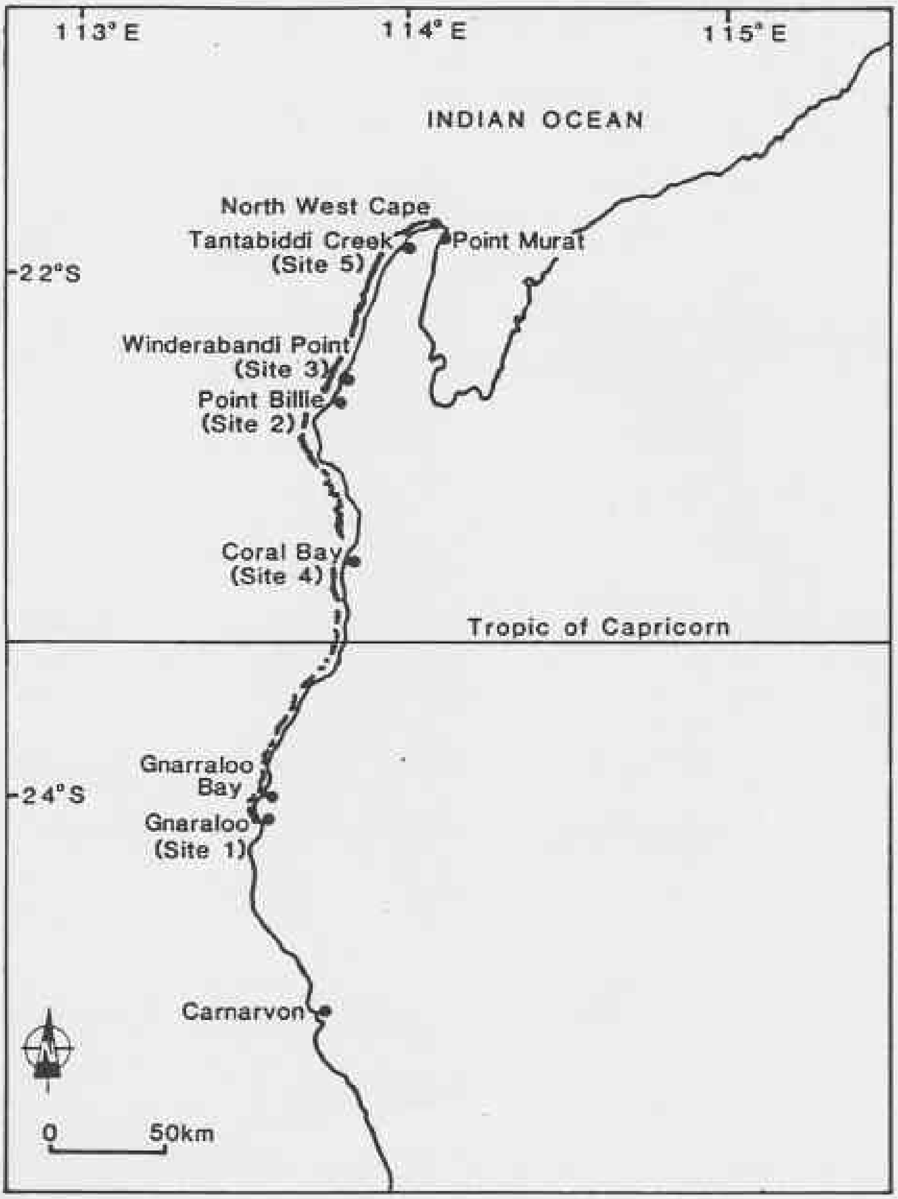

113° E
114° E
115° E
INDIAN OCEAN
North West Cape
Tantabiddi Creek
(Site 5)
Point Murat
22°S
Winderabandi Point
(Site 3)
Point Billie
(Site 2)
Coral Bay
(Site 4)
Tropic of Capricorn
Gnarraloo
Bay
24°S
Gnaraloo
(Site 1)
Carnarvon
0
50km

Figure 1.3

The reef follows the line of the coast as its orientation changes at Point Cloates and remains inside the 30 m isobath (Figure 1.4). However, there is a region between Point Cloates and Point Maud where the lagoon reaches its maximum width of about 6 km. In fact, the bathymetry suggests that the Ningaloo Reef tract can be divided into three sectors.

- Northern sector. North West Cape to Point Cloates (~ 120 km) The lagoon is less than 3 km wide and parallel to a straight coast. The shelf break is also parallel to the shore and the shelf is about 10 km wide.

- Central sector. Point Cloates to Point Maud. The lagoon is some 6 km wide and 50 km long and has the structure of a long embayment with a major break in the reef at its southern end near Point Maud.

- Southern sector. Point Maud to Gnarraloo Bay (90 km). The reef at Amherst Point is very scattered and a definite structure is only evident some 35 km south of Point Maud, at Pelican Point. In this sector the lagoon is about 1 km wide.

The northern sector of Ningaloo Reef tract appears to be especially interesting from the viewpoint of physical studies. It comprises 120 km of straight reef-line, close to both the shelf-break and shore, with a very shallow lagoon. Lengths of straight barrier reef so close to the shore are comparatively rare on the continental shelf. Studies of this sector form a valuable contrast of scale to the extensive work on the Great Barrier Reef where the lagoon is of the order of 100 km wide. Most of the discussion in this report is based on ideas which although particularly relevant to the northern sector are applicable to the entire length of Ningaloo Reef tract.

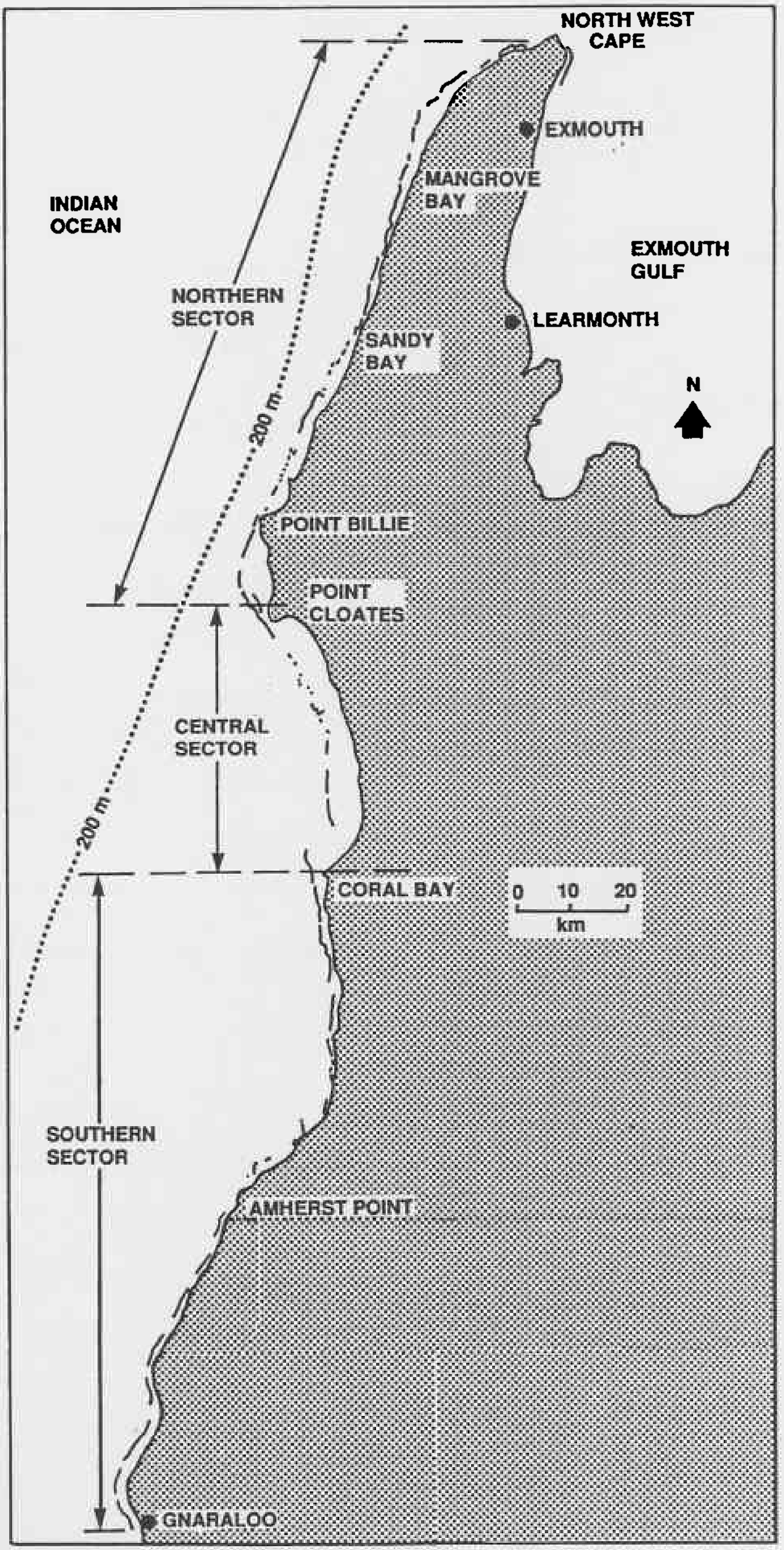

NORTH WEST CAPE
EXMOUTH
MANGROVE BAY
INDIAN OCEAN
EXMOUTH GULF
NORTHERN SECTOR
LEARMONTH
SANDY BAY
N
200 m
POINT BILLIE
POINT CLOATES
CENTRAL SECTOR
200 m
CORAL BAY
0 10 20
km
SOUTHERN SECTOR
AMHERST POINT
GNARALOO

Figure 1.4

CHAPTER 2 REGIONAL OCEANOGRAPHY

2.1 Currents

The continental shelf in the region of the Ningaloo Reef tract is very narrow (approximately 10 km). To the north the shelf expands to nearly 400 km width and the coast runs towards the northeast. This region is called the North West Shelf and has been the subject of considerable oceanographic study in relation to oil and gas programmes. South of Ningaloo the coast runs to the south with a maximum shelf width of 70 km.

The large-scale currents off the west coast of Australia are complex. Most textbooks refer to a northward flowing current within 1000 km of the coast; the West Australian Current. There are two types of evidence which support the existence of this current. The first is the millions of observations of ship drift made for more than a century by merchant ships (Australian Pilot 1972). The second is that wide slow equatorward currents on the eastern sides of oceans are a consequence of the wind circulation around subtropical high pressure systems. In the South Indian Ocean the subtropical ocean gyre consists of the westward South Equatorial Current near 10°S, the narrow strong poleward Agulhas Current near Africa, the eastward West Wind Drift near 40°S and a wide weak equatorward flow over most of the eastern portion of the gyre. Predicted maximum surface speeds in summer are 0.2 to 0.4 m s^{-1} and the currents are somewhat weaker in winter.

The strength of the West Australian Current shows considerable annual variability and it is often displaced away from the continent in summer by a large cyclonic eddy with southward flow near the coast (Cresswell 1977; Cresswell and Golding 1980).

During winter a southward flow called the Leeuwin Current occurs near the shelf-break. This current is a relatively recent discovery (Cresswell and Golding 1980) and considerable data on its surface manifestation has been obtained from the advanced very high resolution radiometer (AVHRR) of the NOAA satellite. According to Holloway and Nye (1985) the current originates on the North West Shelf and there reaches a maximum speed of 0.25 m s^{-1} over the shelf-break. The current seems to strengthen as it passes North West Cape and then flows as a band of lower salinity, warm, tropical water. The current is some 50 km wide and 200 m deep. It flows mainly over the shelf, and shelf-break, and continues around Cape Leeuwin into the Bight. There is some evidence for an equatorward undercurrent containing colder higher salinity water. The Leeuwin Current is believed to be due to a steric height gradient along the coast of about 0.5 m over 14° latitude caused by differential heating in the tropics (Thompson and Cresswell 1983). This is opposed by wind stress which is toward the equator. The overall effect of the wind stress depends on the depth of mixing of the momentum. In summer, and in water less than 30 m depth, the wind is dominant and a northward flow occurs. However in winter the mixed layer becomes much deeper and the pressure gradient dominates the surface region. The seasonality of the Leeuwin Current is poorly documented. This is partly due to the AVHRR information being degraded by surface solar heating in the non-winter months and cloud cover in the north.

2.2 Temperatures

There is a considerable variation of coastal sea temperature along the west coast of Australia. These vary from maximum mean monthly summer values of 30.4°C at King Sound on the northwest coast to 20.9°C at Albany on the southern coast (Figure 1.1). Winter minima are 24.3°C at King Sound and 17.4°C at Albany. At Exmouth the range is 26.6°C (summer) to 22.9°C (winter) (Pearce 1986).

2.3 Upwelling

Upwelling is a process in which subsurface water reaches the surface. It is a conspicuous phenomenon along the eastern boundaries of most oceans due to the prevailing equatorward winds. These produce equatorward surface currents which move offshore due to the Coriolis force so that deeper water moves toward the surface; referred to as Ekman transport. Thus the isotherms curve upward and there is a region of colder water in a narrow band near the coast. This supply of subsurface water is often nutrient rich, and results in increased biological productivity.

There are no indications of permanent upwelling on the western coast of Australia. Coastal temperatures are not depressed and isotherms tend to curve downwards. It is likely that the Leeuwin Current is a major reason for this lack of upwelling. However upwelling is evident on the northwest coast during winter, although the cause seems unclear, and primary production is high (Holloway et al. 1985).

2.4 Stratification

During summer there is a strong temperature gradient from surface to bottom over most of the shelf off the western and northwestern coastline of Western Australia. On the North West Shelf near the shelf-break, a well-defined thermocline does not exist in summer over most of the shelf and temperature increases fairly smoothly through the water column. Late in summer, and into winter, a surface mixed layer develops so that this vertical gradient of temperature is eroded from the surface downwards, i.e. a surface mixed layer deepens to fill the entire water column. Nearshore a thermocline may exist which collapses in winter. Along the west coast the situation is greatly complicated by the Leeuwin Current in winter, and by the seabreeeze near the coast in summer.

2.5 Tides

Table 2.1 shows the amplitudes (h) and phases (g) of the most important astronomical tidal constituents at a number of locations along the west and northwest coast of Australia. Also shown are the frequency in cycles per year of each component, and the value of the form factor F defined as the ratio of the sum of amplitudes of the diurnal to semidiurnal components. An accepted classification of tidal regimes is:

F is 0 to 0.25: Semidiurnal tide. Two high and two low waters of nearly the same height occur each day. The time of high water follows in nearly constant time interval after the moon's transit of the local meridian. The mean range at spring tide is given by $2(M_2 + S_2)$.

F is 0.25 to 1.5: Mixed, predominantly semidiurnal tide. Two high and two low waters occur each day, but with large inequalities in range and time. They assume their maxima when the declination of the moon has passed its maximum value. The mean spring range is $2(M_2 + S_2)$.

F is 1.5 to 3.0: Mixed, predominantly diurnal tide. At times only one high water occurs each day, namely after the extremes of the moon's declination. Otherwise, there are two high waters each day which, however, show large inequalities in range and time, especially when the moon has passed over the equator. The mean spring range is $2(K_1 + O_1)$.

F is greater than 3.0: Diurnal tide. Only one high water occurs each day. At neap tide, when the moon has crossed the equatorial plane, two high waters may occur. The mean spring range is $2(K_1 + O_1)$.

Table 2.1 Amplitude (h) and phase (g) of astronomical tides at various locations (shown in Figure 1.1) on the east coast of Australia. Also shown is the form factor F at each location, and the frequency of each tidal component (from Australian National Tide Tables, published by the Hydrographer RAN)

	Amplitude (h) metres				Phase (g) degrees				
Location	M2	S2	K1	01	M2	S2	K1	01	Form Factor (F)
Bunbury	0.05	0.05	0.16	0.11	291	300	300	289	2.7
Fremantle	0.04	0.04	0.16	0.12	286	302	308	290	3.5
Geraldton	0.07	0.05	0.17	0.12	288	309	301	285	2.4
Carnarvon	0.30	0.14	0.20	0.13	304	006	285	273	0.8
Point Murat	0.49	0.27	0.18	0.13	314	026	302	281	0.4
Learmonth	0.66	0.36	0.19	0.14	312	024	292	281	0.3
Onslow	0.91	0.53	0.20	0.15	321	033	290	280	0.2
Cycles per year									
	705	730	366	339					

Thus the tide on the west Australian coast changes from 'mixed predominantly diurnal' at Bunbury to 'diurnal' at Fremantle and then becomes 'mixed predominantly semidiurnal' at Carnarvon, and progresses to 'semidiurnal' at Onslow on the North West Shelf.

Below Carnarvon the semidiurnal tide is most evident away from the diurnal spring tide. Thus the tidal record has the appearance of diurnal springs with neaps showing semidiurnal tides. Above Carnarvon the diurnal tide tends to be more pronounced away from the semidiurnal springs.

The Ningaloo Reef tract lies between the standard ports of Carnarvon and Point Murat (Figure 1.3). Pressure sensor deployments along the Ningaloo Reef tract indicate that the tides within the lagoon along most of the tract are better correlated in phase and amplitude with the predicted tides at Carnarvon than at Point Murat (Simpson and Masini 1986). The tides at Ningaloo are mixed, predominantly semi-diurnal with a form factor near 0.8 and a maximum range at springs of about 2 m.

Apart from the tidal components discussed here there is a range of smaller and lower frequency components. These alter the effective amplitude and phase of the major components.

An important annual variation in mean sea-level occurs in the southern part of the west coast. It is of amplitude 0.1 m with zero occurring in October and March and the elevation being positive in winter. The annual variation seems to be absent at North West Cape but is present on the North West Shelf with opposite phase, i.e. elevation is negative in winter.

2.6 Internal Waves

Internal waves are gravitational oscillations of the thermocline separating bottom denser water from the upper lighter water. In fact, it is not necessary to have a sharp interface between the two fluids for internal waves to occur; any variation of density through the water column will support internal waves.

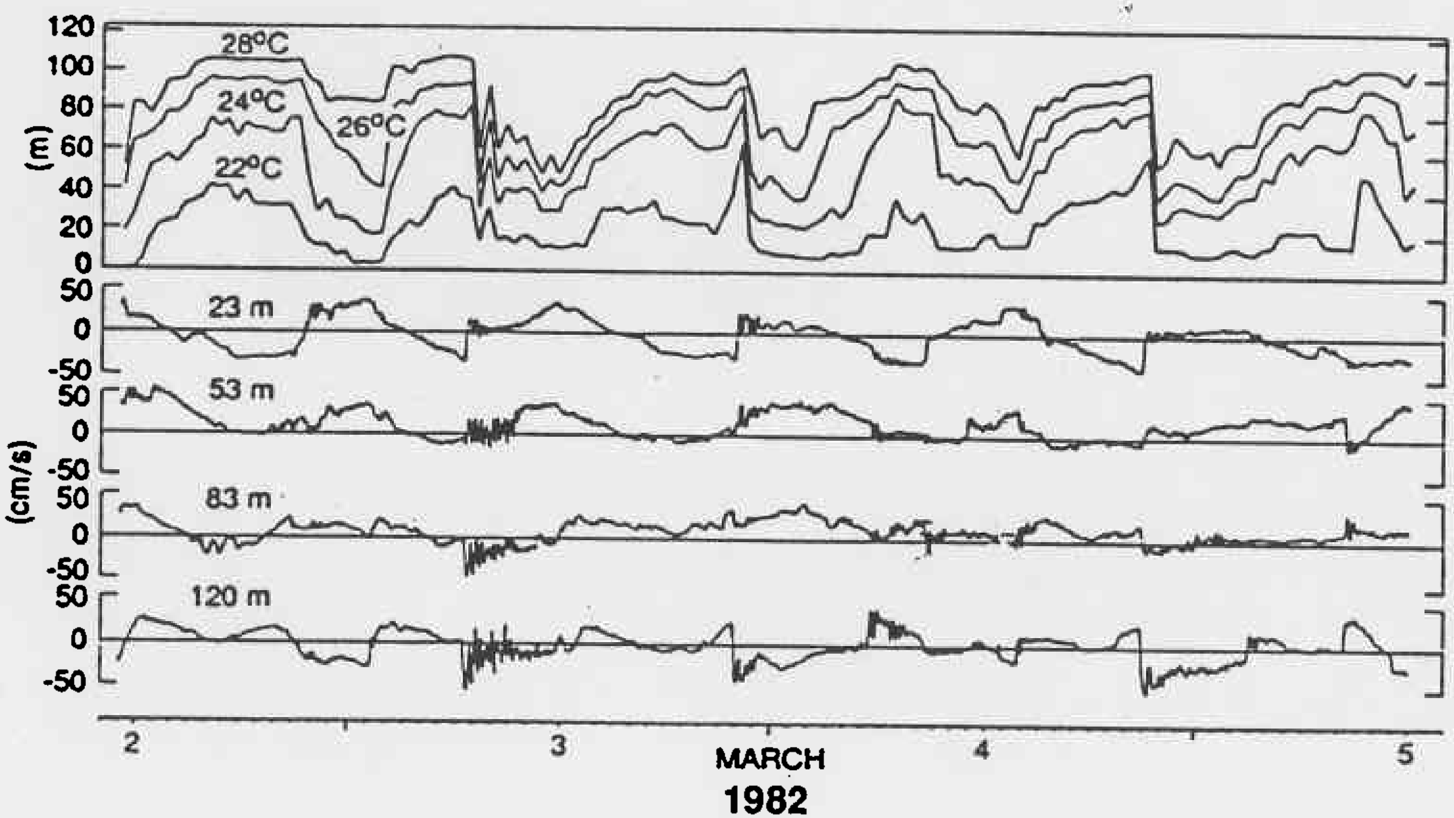
The waves have frequencies which vary between the local inertial frequency (period of 31.9 hours at Ningaloo Reef) and the depth-averaged buoyancy frequency

$$\left(\alpha\ g\ \frac{dT}{dz}\right)^{1/2}$$

where T(z) is the temperature profile through the water column and $\alpha \sim 10^{-4}$°C is the thermal expansion coefficient. Usually this frequency is of order 0.01 to 0.1 s^{-1} in a stratified sea, and the internal waves of interest are of much lower frequency. Maximum vertical displacements occur in the interior of the water column and maximum horizontal velocities near the surface and bottom.

Wyrtki (1962) reported that internal waves produced vertical displacements of isotherms of tens of metres off the west coast and this has been studied in detail by Holloway (1983; 1986a) on the North West Shelf. The internal (or baroclinic) waves are forced by the astronomical tide and are referred to as the baroclinic tide because they have the same frequency components as the ordinary (or barotropic) tide. However the baroclinic tide is not coherent and appears as a set of waves of random phase which seem to propagate onshore with typical phase speed of 0.4 m s^{-1}, i.e. taking 4 days to cover 140 km from the shelf-break to shore at Dampier. They have a typical wavelength of 20 km and are rapidly damped in amplitude as they cross the shelf. They are assumed to be generated by the action of the barotropic tide flowing over the shelf-break but this generation region does not appear to be very localised.

Holloway (1983) found vertical displacements of up to 30 m in water of depth less than 100 m. The resulting baroclinic currents reached



120
100
80
60
40
20
0
(m)
28°C
24°C
26°C
22°C
50
0
-50
(cm/s)
23 m
53 m
83 m
120 m
2
3
4
5
MARCH
1982

Figure 2.1

magnitudes of 0.2 m s^{-1} in the cross-shelf direction. Baines (1981) has also identified internal waves on the North West Shelf from Landsat imagery, finding packets of waves with wavelengths from 300 to 1000 m, 25 km between packets and propagation speeds of 0.5 to 1.0 m s^{-1}.

Holloway (1986a) has developed the idea of a climate for the baroclinic tide. This gives a monthly assessment of the mean amplitude of the (semidiurnal) internal tide. The climatology has its origins in a number of physical factors which have seasonal variation: the stratification or buoyancy frequency is the most important. The amplitude of the baroclinic tide is maximum in the period November to April and falls off by about a factor of 2 to become minimum in June to September. This is essentially a result of the buoyancy frequency becoming small as the shelf waters become vertically mixed in winter.

Internal waves on the North West Shelf have very large amplitude in relation to the water depth. In this sense the waves are often close to 'breaking'. Holloway (1986b) finds that the waves show 'internal jumps'. Figure 2.1 shows a time series from March 2 to 5, 1982, at the North Rankin mooring near the shelf-break off Dampier, of the heights of three isotherms.

It is possible that such large waves may break the surface and so release colder bottom water into the shallow lagoon of the Ningaloo Reef tract.

2.7 Exmouth Plateau

The Exmouth Plateau is a shallow region lying ~ 250 km offshore from Exmouth. It is the closest area to the Ningaloo Reef tract which has received detailed oceanographic study (Holloway 1986c).

Like the Ningaloo Reef tract, the plateau has mixed diurnal and semidiurnal tides. However tidal currents seem to be mainly semidiurnal with only weak currents at the diurnal frequency. This is consistent with observations on the west coast (e.g. Hearn et al. 1985) where the semidiurnal tide is much smaller and tidal currents are of order 0.02 m s^{-1} in the open ocean. The semidiurnal tidal currents are both barotropic and baroclinic.

Holloway (1986c) found strong inertial period currents and high energy bursts of short period (~ 30 minutes) internal waves. A salinity and temperature survey conducted from July to November 1978 showed the absence of a well-mixed surface layer; temperature fell fairly steadily from 24°C at the surface to 10°C at 300 m depth. Salinity was close to 35°/oo but with a slight maximum of 35.6°/oo near 200 m depth.

2.8 Coastal Circulation

Currents on the continental margin are strongly influenced by meteorological forces. There are several components to this forcing: momentum transfer through the wind stress; wind stirring of the water column; pressure adjustments; heat fluxes, and mass (water) transfer through evaporation and precipitation. These forces act on a host of different space and time scales through various meteorological phenomena. In this section some discussion is presented of quasi-steady currents generated by wind forcing.

Surface wind stress in a coastal region produces a velocity within the water column which varies with depth. Surface water always moves essentially with the wind. However below the surface the direction and speed of the water is controlled by a balance between the downward flux of momentum from the wind, the inertia of the water column, pressure gradients arising from the wind-induced motion, bottom friction, and Coriolis forces due to the earth's rotation. Given this variety of forces it is necessary to consider special cases in order to allow a tractable framework for analysis.

Coastal ocean circulation is therefore studied using a water body, either of uniform density, or as a finite number of layers of different densities. Only the former case (barotropic motion) is considered here. In this case the depth-averaged velocity, which determines the net transport of water, can be derived from the balance of wind stress, bottom friction, and the horizontal pressure gradient due to the slope of surface elevation of the water. The effect of wind-stress and surface elevation are separable and the current can be expressed as a sum of a wind-current and a slope-current. However, the depth-averaged current must satisfy the constraints of continuity.

The currents induced by a wind stress acting on a coastal sea are determined by two major considerations. The first of these is the alongshore variability of the system, and involves changes in wind-stress, coastal direction, and offshore bathymetry. The second is the offshore bathymetry. The simplest possible case of an infinite straight coast with constant water depth and wind stress was solved by Ekman in 1905. Although these solutions formed a major step in our basic understanding of coastal oceanography, the circulation created by wind stress on a real coast is controlled by bathymetry and alongshore variations.

Csanady (1978; 1980) has considered the coastal circulation which arises in a vertically-mixed ocean as a result of cross-slope and alongshore wind stress which varies in magnitude in an alongshore direction. This introduces an alongshore length scale L_{along} over which the wind stress varies and an offshore length L_{off} which is defined by

$$L_{off} = \frac{2\ r}{f\ s}$$

where r is the friction velocity, f the Coriolis parameter, and s is the mean offshore slope of the shelf. Typical values of L_{off} are of order 10 km.

If $L_{along} >> L_{off}$ the effects of the alongshore wind is generally dominant whilst for $L_{along} << L_{off}$ the offshore wind is the more important. For most meteorological systems L_{along} would be expected to be of the order of 1000 km so that alongshore wind controls circulation. Csanady (1978) shows that this produces an 'arrested or trapped topographic wave' with nearshore flow in the direction of the wind up to an offshore 'trapping distance' of

$$(L_{off} \cdot L_{along})^{1/2}$$

i.e. typically less than 100 km. Beyond this 'coastal boundary layer' cross-shore Ekman flow occurs. Associated with the wave is a sea-level variation which is maximal at the shore.

Figure 2.2a shows a trapped wave due to a variation in magnitude and sign of the alongshore wind. The figure shows regions of high pressure (sea-level raised) and low pressure (sea-level reduced). These provide a

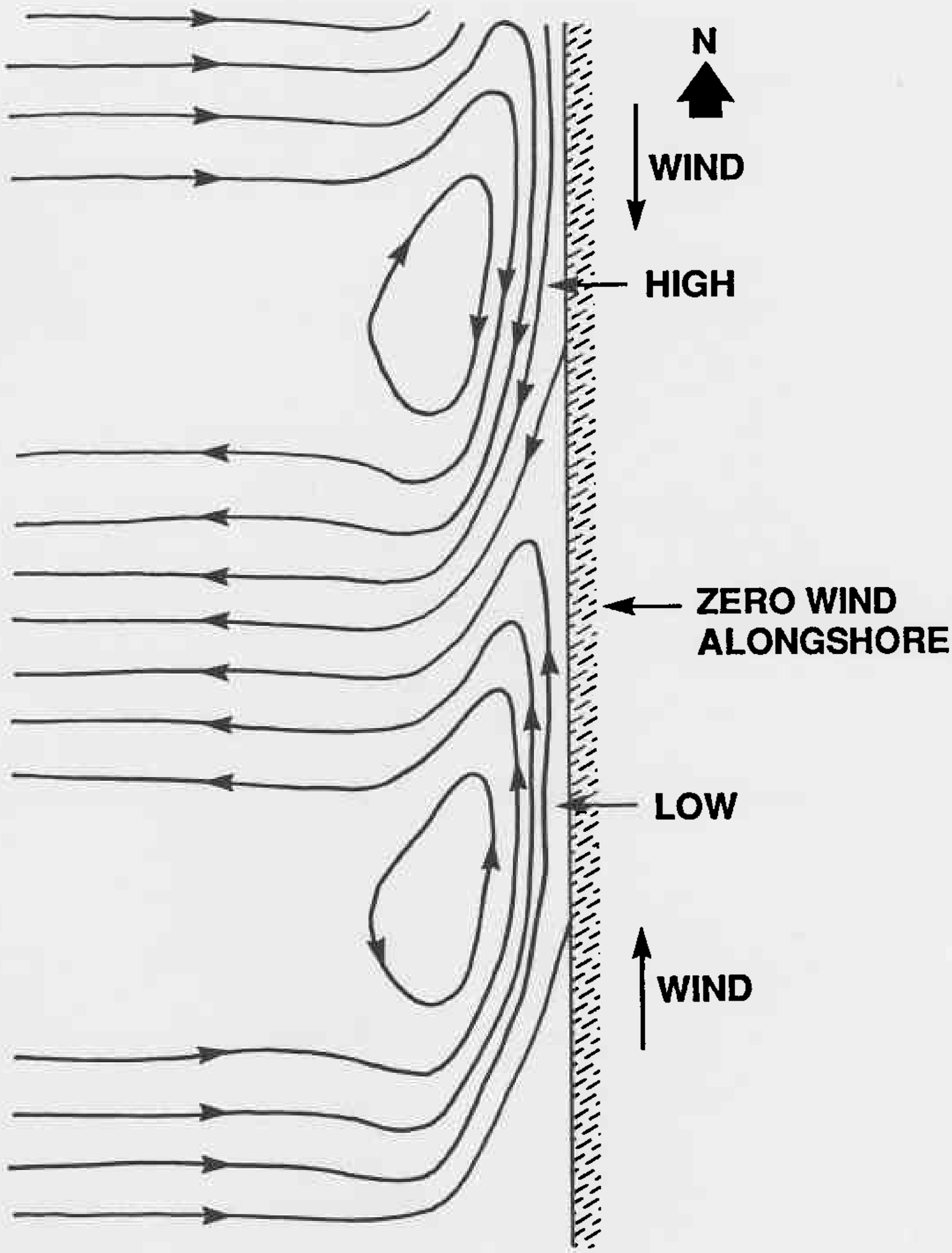


Figure 2.2a

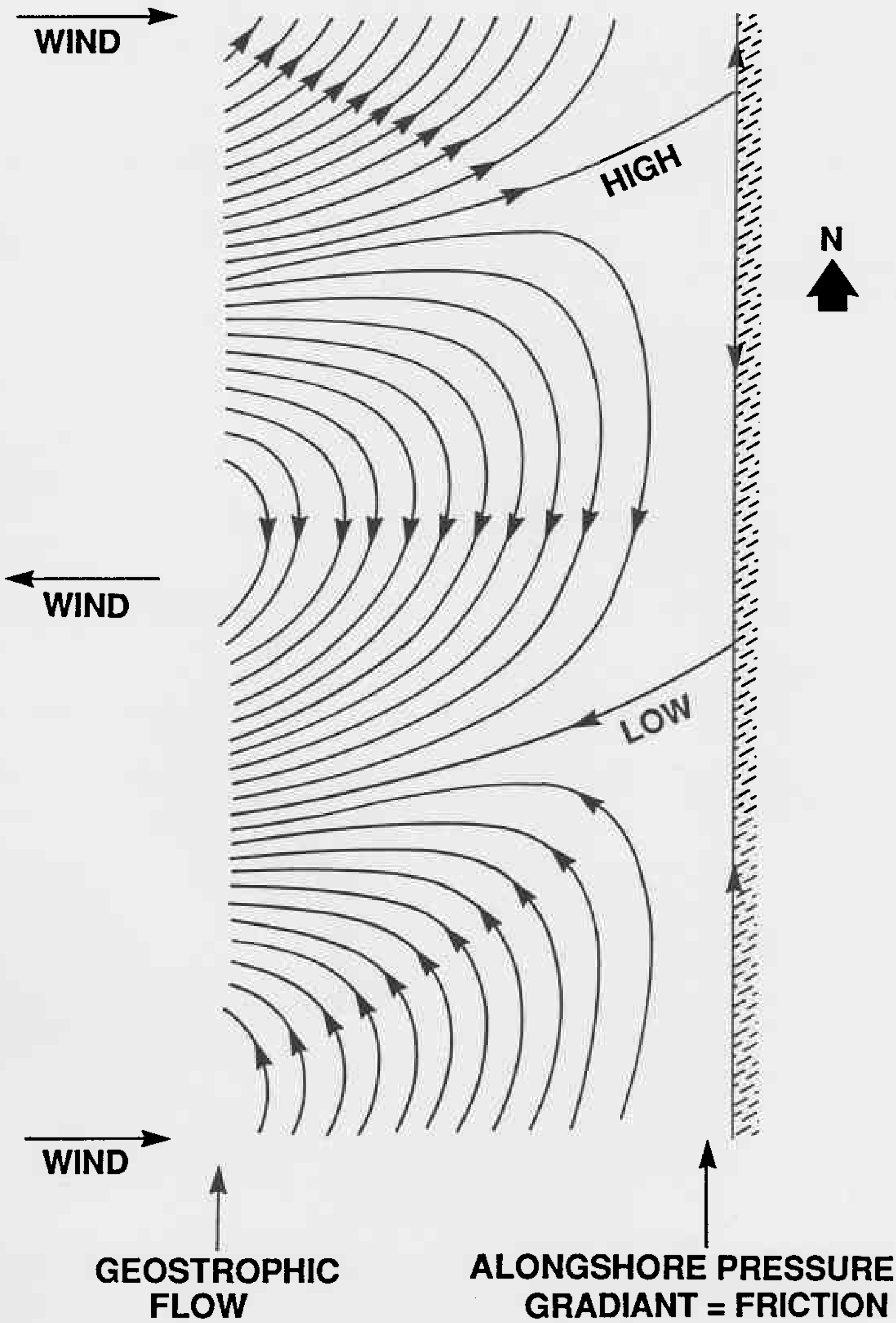

WIND
HIGH
N
WIND
LOW
WIND
GEOSTROPHIC
FLOW
ALONGSHORE PRESSURE
GRADIANT = FRICTION

Figure 2.2b

pressure gradient which balances the Coriolis force due to near-shore flow parallel to the shore. It is expected that this circulation and surface elevation will migrate along the coast with the meteorological system. A nearshore current of 0.1 m s^{-1} would typically be associated with elevation changes of order a few centimetres.

In the case where L_{along} is less than L_{off}, or of a similar size to L_{off}, the cross-shore winds become effective in raising the sea-level at the coast. This is shown by Figure 2.2b. The changes in water elevation produce pressure gradients to oppose the cross-shore winds at the coast. However the location of the maxima and minima in the elevation are displaced from the points of maximum wind stress. The trapping distance is the same as for alongshore winds and within the coastal boundary layer the currents again flow parallel to shore. Further offshore, geostrophic balance occurs with flow perpendicular to the wind stress.

Sea-level changes produced by cross-shore winds are typically of the order of 10 cm, i.e. an order larger than for longshore winds. They migrate with meteorological systems. However they are usually only important for storm conditions on a straight coast, or in the vicinity of sudden changes in the direction of a coast or its bathymetry.

Cyclones (Section 4.6) are occasional occurrences on the North West Shelf. They have spatial scales for wind variation of the order of 100 km; the core of the cyclone is much smaller (~ 30 km) and winds are maximal at the edge of this core. Cyclones also produce currents through the spatial gradient of their atmospheric pressure. However cyclones are infrequent.

It is shown in Section 3.2 that winds near the Ningaloo Reef tract may have a spatial variation closer to 100 km than 1000 km because of the change in coastal orientation at North West Cape.

The value of L_{off} for the shelf near the central and southern sectors of the Ningaloo Reef tract is typically 10 km. However the shelf is much steeper for the northern sector and L_{off} reduces to a value of the order of 1 km. It therefore seems that alongshore winds will be dominant over the entire length of coast in which the reef is situated. Typical values for the trapping distance will be 10 to 30 km.

It must be emphasised that these features refer to large-scale circulation on the shelf outside the reef edge; circulation within the lagoon is discussed in Section 4.3.

CHAPTER 3 CLIMATE

3.1 Introduction

The annual variation in the climate of Western Australia is a result of the movement of a major atmospheric circulation knowns as the Hadley cell. This consists of warm air rising in the tropics and then moving poleward to descend in the sub-tropics. This descending air creates a region of high pressure known as the sub-tropical ridge. It forms a belt of warm dry air around the globe and is associated with the worlds major deserts.

In the southern summer the meteorological equator migrates southward and the sub-tropical ridge is located between 35° and 40°S. In winter there is a corresponding northward movement with the ridge between 25 and 30°S. As a result, northern Australia experiences dry conditions during winter whilst in summer it lies within the tropical low pressure system.

The equatorward movement of surface air has a component towards the west, due to the rotation of the earth, and over the ocean this creates the southeast trade winds. These tend to dominate the prevailing wind conditions at North West Cape for most of the year. In winter the northward movement of the ridge leads to the prevailing winds becoming more easterly. However the Cape does not experience the summer northwesterly winds and monsoon conditions, which occur further north.

3.2 Winds

Wind data are collected by the Bureau of Meteorology at Carnarvon, Learmonth, and Exmouth. Carnarvon is 130 km south of the southern boundary of the Ningaloo Reef (Gnarraloo Bay) and is representative of the southern, and perhaps central, sectors of the reef. Learmonth and Exmouth are close to the northern sector of the reef but are separated from the reef by the Cape Range mountains which lie along the centre of Exmouth Peninsular. The peninsular is some 25 km in width and the range reaches 300 m above sea-level. Furthermore certain local aspects of winds at these stations are evidently controlled by air-sea interactions within Exmouth Gulf. Thus no reliable data exists for wind conditions on the Ningaloo Reef tract, especially the northern sector.

Wind statistics have been developed here from data supplied by the Bureau of Meteorology.

Tables 3.1 and 3.2 show mean monthly wind speeds and directions based on 09.00 and 15.00 hours data at Carnarvon and Learmonth. The means are based on the wind run, and wind velocity vector. An equivalent mean wind for the surface stress vector (which is the velocity vector multiplied by wind speed) is also included. The tables present the measured wind speed and no correction has been applied for anemometer elevation (4 and 5 m at Carnarvon and Learmonth, respectively). The data is a mean over 39 years at Carnarvon and over 9 years at Learmonth. Wind run is slightly lower at Learmouth than Carnarvon. However the mean wind velocity and mean wind stress is much lower at Learmonth because the direction of the wind is much more variable.

Table 3.1 shows that the mean annual wind (stress) direction at Carnarvon is 194°. At 09.00 (Table 3.3) the direction is 152°, and at 15.00 hours (Table 3.4) it is 205°. This change is due to the seabreeze which produces a diurnal cycling of the wind direction from southeast to southwest. The effect is observed along the entire west and southwest coast both at coastal stations and some hundreds of kilometres inland. A seabreeze is identified on approximately 70% of afternoons (Southern 1979).

At Learmonth (Tables 3.2, 3.5 and 3.6) the mean annual wind (stress) direction is 178° with a change from 171° at 09.00 hours to 229° at 15.00 hours. The wind direction at Learmonth is rotated to the west relative to Carnarvon; the 09.00 winds at Learmonth are more southeasterly and the 15.00 winds are much more westerly. However the 15.00 wind speed is much weaker on average, and as a result, the mean annual wind direction at Learmonth is similar to that at Carnarvon.

The weaker southwesterly seabreeze at Learmonth is actually a result of two partially compensating afternoon wind situations. Tables 3.7 and 3.8 show the annual occurrence matrices at Carnarvon and Learmonth for 09.00 and 15.00 hours. At Carnarvon the wind directions at 09.00 hours lie mainly between south and east and at 15.00 the direction is south to southwest. At Learmonth the wind direction at 09.00 is also between south and east. However the direction of the wind at Learmonth at 15.00 hours lies either south to west or east to north. Inspection of the occurrence matrix at Learmonth at 15.00 hours shows that the afternoon northeasterly wind is weaker than the southwesterly seabreeze. It occurs on 56% of days with a mean speed of 16.8 km hr^{-1} (4.7 m s^{-1}) whilst the southwesterly occurs on 40% of days with a mean speed of 22.7 km hr^{-1} (6.3 m s^{-1}).

Table 3.1 Monthly and annual mean wind speed and direction at Carnarvon (39 years of data) based on 09.00 and 15.00 hours observations

	Mean Wind Run	Mean Wind Velocity		Mean Wind Stress	
	Speed $m\ s^{-1}$	Speed $m\ s^{-1}$	Direction Degrees	Speed $m\ s^{-1}$	Direction Degrees
January	7.17	5.31	196	6.69	193
February	6.67	4.69	189	6.12	187
March	6.37	4.86	184	6.07	183
April	5.61	3.81	177	5.16	176
May	5.19	2.61	157	4.12	156
June	4.51	1.30	152	2.71	156
July	4.69	1.83	155	3.30	158
August	5.39	3.00	168	4.59	168
September	6.39	4.31	181	5.87	179
October	6.94	4.88	191	6.29	188
November	7.40	5.83	196	7.00	194
December	7.30	5.56	198	6.83	196
ANNUAL	6.14	3.89	184	5.51	184

Table 3.2 Monthly and annual mean wind speed and direction at Learmonth (9 years of data) based on 09.00 and 15.00 hours observations

	Mean Wind Run	Mean Wind Velocity		Mean Wind Stress	
	Speed $m\ s^{-1}$	Speed $m\ s^{-1}$	Direction Degrees	Speed $m\ s^{-1}$	Direction Degrees
January	5.97	2.21	213	4.36	216
February	5.29	1.07	207	2.77	215
March	4.89	1.42	134	2.79	140
April	4.25	1.14	139	2.36	150
May	4.22	1.94	112	3.09	113
June	4.12	1.73	103	2.89	94
July	4.28	1.69	119	3.00	118
August	4.41	1.42	128	2.63	134
September	5.33	2.37	158	3.98	165
October	5.77	2.25	182	4.12	190
November	6.16	3.14	199	4.92	203
December	5.99	3.05	203	4.87	205
ANNUAL	5.06	1.53	165	3.19	178

Table 3.3 Monthly and annual mean wind speed and direction at Carnarvon (39 years of data) at 09.00 hours

	Mean Wind Run	Mean Wind Velocity		Mean Wind Stress	
	Speed $m\ s^{-1}$	Speed $m\ s^{-1}$	Direction Degrees	Speed $m\ s^{-1}$	Direction Degrees
January	6.65	4.78	169	6.31	167
February	6.26	4.32	159	5.87	159
March	6.06	4.72	159	5.90	158
April	5.09	3.68	142	4.93	141
May	4.67	3.30	111	4.44	112
June	3.92	2.16	99	3.26	105
July	4.02	2.48	106	3.57	109
August	4.80	3.29	123	4.62	125
September	5.69	4.08	146	5.60	144
October	6.41	4.44	161	5.96	157
November	6.78	5.42	172	6.60	169
December	6.75	5.04	175	6.44	171
ANNUAL	5.59	3.65	150	5.24	152

Table 3.4 Monthly and annual mean wind speed and direction at Carnarvon (39 years of data) at 15.00 hours

	Mean Wind Run	Mean Wind Velocity		Mean Wind Stress	
	Speed $m\ s^{-1}$	Speed $m\ s^{-1}$	Direction Degrees	Speed $m\ s^{-1}$	Direction Degrees
January	7.68	6.65	214	7.50	211
February	7.67	5.99	210	6.87	207
March	6.66	5.74	204	6.63	201
April	6.11	5.06	202	5.97	198
May	5.68	3.63	194	4.86	190
June	5.10	2.12	204	3.38	201
July	5.33	2.72	197	4.02	193
August	5.99	4.37	200	5.50	197
September	7.08	5.78	205	6.82	201
October	7.48	6.35	212	7.23	209
November	7.97	6.94	213	7.77	210
December	7.85	6.79	215	7.62	213
ANNUAL	6.67	5.15	207	6.31	205

Table 3.5 Monthly and annual mean wind speed and direction at Learmonth (9 years of data) at 09.00

	Mean Wind Run	Mean Wind Velocity		Mean Wind Stress	
	Speed $m\ s^{-1}$	Speed $m\ s^{-1}$	Direction Degrees	Speed $m\ s^{-1}$	Direction Degrees
January	5.42	3.45	189	4.88	192
February	4.95	3.09	186	4.29	188
March	4.70	3.38	168	4.38	168
April	4.03	2.87	169	3.75	170
May	3.87	2.76	147	3.62	144
June	3.60	2.43	146	3.31	133
July	3.90	2.71	162	3.69	154
August	4.14	2.65	158	3.62	159
September	5.34	4.34	163	5.31	164
October	5.28	4.02	173	5.22	174
November	5.74	4.33	177	5.46	179
December	5.53	4.25	184	5.33	185
ANNUAL	4.71	3.27	170	4.39	171

Table 3.6 Monthly and annual mean wind speed and direction at Learmonth (9 years of data) at 15.00 hours

	Mean Wind Run	Mean Wind Velocity		Mean Wind Stress	
	Speed $m\ s^{-1}$	Speed $m\ s^{-1}$	Direction Degrees	Speed $m\ s^{-1}$	Direction Degrees
January	6.50	1.87	259	4.33	245
February	5.62	1.25	150	2.83	130
March	5.07	1.78	228	2.92	225
April	4.45	1.31	225	2.15	222
May	4.51	2.13	252	3.17	257
June	4.55	2.17	244	3.18	239
July	4.57	2.07	254	3.12	258
August	4.65	1.37	247	2.32	252
September	5.32	0.64	117	1.97	172
October	6.26	0.86	233	3.25	233
November	6.60	2.79	235	4.95	233
December	6.42	2.52	234	4.79	229
ANNUAL	5.38	0.26	236	1.68	229

Table 3.7 Annual occurrence matrices at Carnarvon

09.00 hours
Speed (km/hour)

Direction	1-5	6-10	11-20	21-30	31-40	40+	
N	0.4	1.1	1.8	0.5	0.0	0.0	3.8
NE	0.8	1.8	2.7	1.3	0.4	0.0	6.9
E	1.1	2.8	6.3	5.0	2.2	0.4	17.8
SE	1.0	2.6	8.3	8.7	4.6	1.1	26.2
S	0.5	1.9	8.6	13.1	7.2	1.9	33.2
SW	0.0	1.5	3.9	2.2	0.4	0.0	8.0
W	0.0	0.8	1.6	0.5	0.0	0.0	2.9
NW	0.0	0.3	0.7	0.1	0.0	0.0	1.1
	3.7	12.8	33.7	31.4	14.8	3.4	100

15.00 hours
Speed (km/hour)

Direction	1-5	6-10	11-20	21-30	31-40	40+	
N	0.0	0.1	0.4	0.3	0.2	0.1	1.0
NE	0.0	0.0	0.3	0.3	0.0	0.0	0.5
E	0.0	0.2	1.1	1.0	0.2	0.0	2.5
SE	0.0	0.2	0.9	1.4	0.6	0.0	3.1
S	0.0	0.4	6.1	15.8	12.1	2.4	36.8
SW	0.1	1.5	11.1	18.7	8.9	1.4	41.8
W	0.0	1.1	4.8	4.4	0.8	0.0	11.1
NW	0.0	0.1	1.1	1.3	0.8	0.0	3.3
	0.0	3.6	25.9	43.1	23.5	3.9	100

Table 3.8 Occurrence matrices at Learmonth

09.00 hours
Speed (km/hour)

Direction	1-5	6-10	11-20	21-30	31-40	40+	
N	0.1	0.2	1.2	0.3	0.0	0.0	1.8
NE	0.4	0.7	2.4	1.2	0.1	0.0	4.9
E	0.7	2.2	5.1	1.4	0.3	0.0	9.7
SE	1.3	3.0	8.2	4.3	1.6	0.0	18.5
S	2.2	5.9	20.1	15.9	4.3	0.2	48.6
SW	0.6	1.7	5.2	4.8	0.8	0.0	13.1
W	0.2	0.4	1.4	1.0	0.0	0.0	3.0
NW	0.0	0.1	0.3	0.0	0.0	0.0	0.4
	5.5	14.2	44.0	29.0	7.1	0.2	100

15.00 hours
Speed (km/hour)

Direction	1-5	6-10	11-20	21-30	31-40	40+	
N	0.3	1.9	9.4	6.1	1.3	0.2	19.2
NE	0.6	2.8	13.2	5.3	1.0	0.2	23.1
E	0.6	2.9	9.4	0.9	0.0	0.0	13.7
SE	0.2	0.7	4.0	2.9	0.5	0.0	8.2
S	0.0	0.2	2.9	5.4	2.3	0.3	11.0
SW	0.0	0.0	2.2	6.5	3.6	0.5	12.7
W	0.0	0.3	2.0	5.7	2.1	0.1	10.2
NW	0.0	0.2	1.0	0.6	0.1	0.0	1.9
	1.6	8.9	44.0	33.3	10.8	1.3	100

The northeasterly is evidently a seabreeze from Exmouth Gulf. Learmonth is subject to both the southwesterly seabreeze from the west coast and this additional seabreeze from the northeast. This is a fairly complex situation and either effect may be dominant on a particular day. There is also the possibility of a transition from one seabreeze effect to the other during one day.

The Ningaloo Reef tract is likely to be partially sheltered from the northeasterly by the Cape Range. However the southwesterly will clearly be affected by the Gulf and Cape Range. It is anticipated that the net strength of the southwesterly will be much lower on the northern sector of the Ningaloo Reef than on the southern or central sectors which are further from both Exmouth Gulf and the northwest coast. This very important change in wind conditions over a distance of order 100 km is likely to have important effects on large scale circulation on the shelf as mentioned in Section 2.8.

3.3 Rainfall, Evaporation and Solar Radiation

Ningaloo Reef lies in the arid region of Western Australia which covers all of the northern parts of the west coast and extends inland to the central, eastern and southern parts of the state. Mean annual rainfall lies between 250 and 300 mm. There is however considerable variability in rainfall between years. Much of the rain results from storms and cyclones and this enhances the variance of the rainfall. The low terrestrial runoff from this arid region has, without doubt, aided the development of a major barrier reef so close to the continent.

Table 3.9 shows mean and median monthly rainfall at Learmonth based on 86 years of data. Rain occurs on comparatively few days mainly between February and July. There are no rivers in the region and surface water enters the reef area through creeks.

Evaporation at Ningaloo station is typical of the region and greatly exceeds mean rainfall throughout the year. Based on 14 years of data from class A pan evaporometers the mean annual evaporation is about 2700 mm. Although such data may not be strictly applicable to a coastal lagoon it provides a reasonable estimate. Table 3.10 gives monthly means.

The annual global radiation at Ningaloo based on data for 1968 to 1974 is 265 J s^{-1} m^{-2}. The is close to the maximum for the whole of Australia. The monthly means for January, April, July and October are given in Table 3.11.

Table 3.9 Mean and median monthly rainfall in mm, and number of rain days at Learmonth based on 1898 to 1984 data

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Mean	18	30	33	17	49	48	28	16	3	2	1	2
Median	2	9	5	3	37	37	20	10	0	0	0	0
Rain days	2	2	2	1	4	5	4	3	1	1	0	0

Table 3.10 Estimated mean monthly pan evaporation at Ningaloo station in mm based on 1967 to 1981 data

Jan.	Feb.	Mar.	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
350	320	300	200	180	120	135	170	240	300	300	360

Table 3.11 Estimated mean monthly global radiation at Ningaloo Reef in J s^{-1} m^{-2} based on data from 1968 to 1974

January	April	July	October	Annual
336	240	168	319	265

CHAPTER 4 OCEANOGRAPHIC PROCESSES ON THE REEF AND IN THE LAGOON

The following sections describe component parts of the oceanographic processes for the Ningaloo Reef tract and its lagoon. Their separation, although somewhat artificial in a practical sense, helps to clarify the nature of the individual processes. The overall dynamics can then be synthesised for any combination of driving forces.

4.1 Wave Pumping

Figure 4.1 shows a typical cross-section of Ningaloo Reef taken normal to the shore. It consists of a seaward reef slope, a reef crest, a reef flat, a lagoon and the shore. Within the lagoon there are further shallow patch reefs and occasional platform reefs along the shoreline.

The seaward reef slope has not been examined in detail but presumably has a similar structure to that of other high energy coral reefs. The main physical function of the seaward reef slope is to dissipate the energy of incident waves. In this respect it serves as a very efficient breakwater. The dissipation process is achieved by grooves which cut the seaward face of the reef at fairly regular intervals. The modal wave length of this spatial array of grooves is usually about 10 m or more.

The grooves commence at a depth of some tens of metres and run up the reef slope into and through the surf zone. At the inner end they may end abruptly or form surge channels. Munk and Sargent (1948) first pointed out that the starting point of these grooves corresponds to the depth at which wave action becomes significant. The threshold velocity for movement of sand of 1 mm diameter is about 0.3 m s^{-1}. For the swell and wind waves

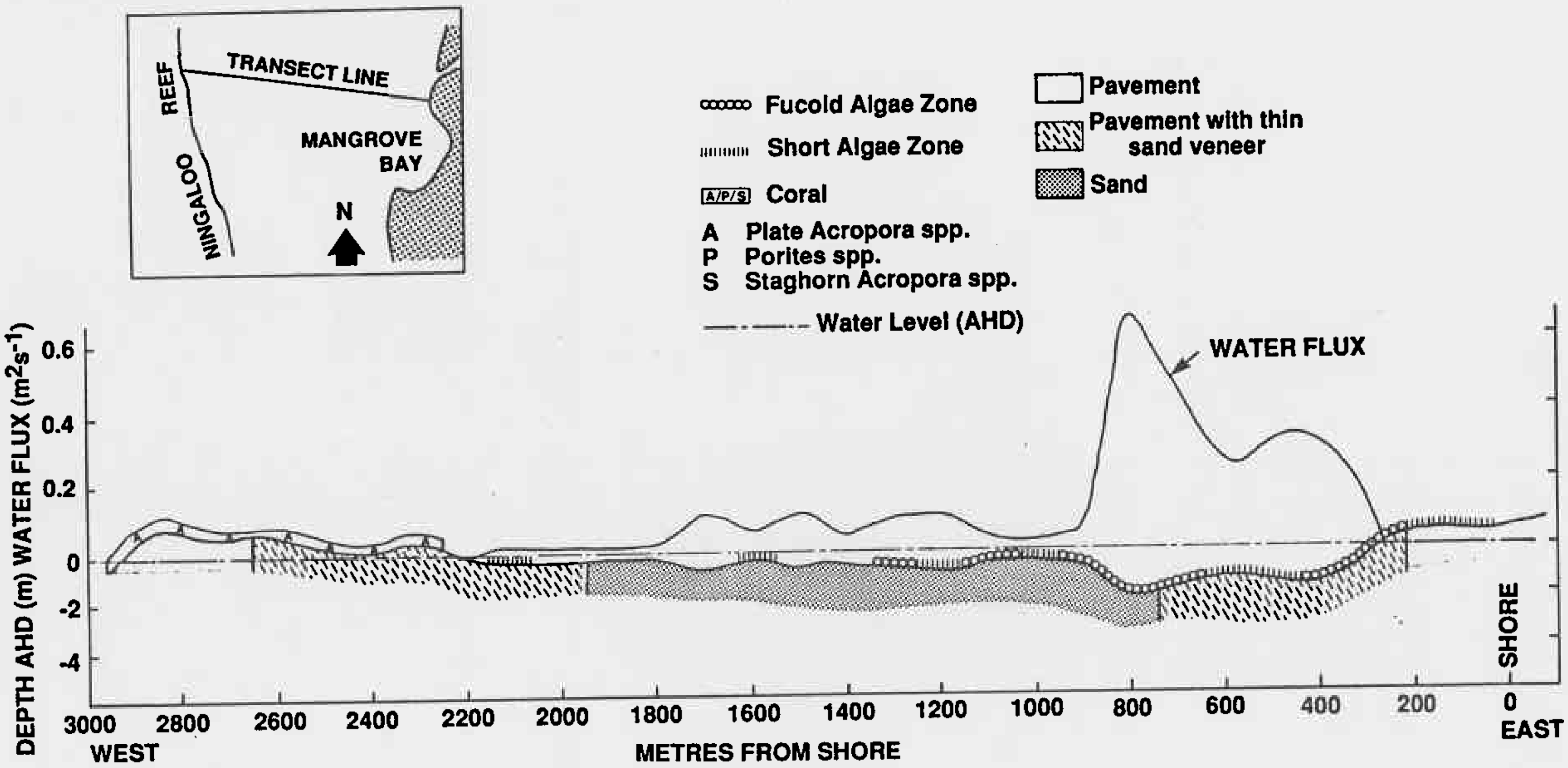

TRANSECT LINE
REEF
NINGALOO
MANGROVE BAY
N
Fucoid Algae Zone
Short Algae Zone
A/P/S Coral
A Plate Acropora spp.
P Porites spp.
S Staghorn Acropora spp.
Water Level (AHD)
Pavement
Pavement with thin sand veneer
Sand
WATER FLUX
SHORE
DEPTH AHD (m) WATER FLUX (m^2s^{-1})
0.6
0.4
0.2
0
-2
-4
3000
2800
2600
2400
2200
2000
1800
1600
1400
1200
1000
800
600
400
200
0
WEST
EAST
METRES FROM SHORE

Figure 4.1

typical of the west coast this orbital velocity is reached at a depth of about 20 m. This is typical of most coral reefs throughout the tropics.

Inspection of aerial photographs of the Ningaloo Reef shows structure in front of the surf-zone which probably corresponds to grooves. The lines are estimated to be of the order of 10 to 20 m long. According to Roberts et al. (1975) the spatial-frequency distribution of the quasi-periodic groove and spur structures have a similar shape for most reefs. However the modal wavelength becomes shorter as the energy of impinging waves increases. Grooves are usually normal to the reef line and only occur on those parts of the reef where this normal lies within a dominant part of the annual wave rose.

The Ningaloo Reef has an outward normal which mostly lies in the quadrant northwest-to-southwest. However portions of the reef do face north and south. It would appear from the aerial photographs that grooves are absent from the northward facing reef fronts which is consistent with the direction of wind and swell waves.

Coral reefs usually contain surge channels on the reef crest which are essentially extensions of the grooves. The channels help to dissipate wave energy by a process of destructive interference between successive waves. The action of the grooves and channels is to remove a major fraction α of the wave energy through frictional processes of various types.

A fraction of the incoming flux of water proceeds to cross the reef crest and enter the lagoon. This appears as a fairly constant flow and does not contain any appreciable component at the wave frequency. The dynamics of the flow are controlled by a wave set-up on the reef crest. This elevates the time averaged water level on the reef to above the level for the open ocean. The elevation Δh is such that if the incoming wave energy flux is E

$$(1 - \alpha)E = \rho g \, \Delta h \, Q$$

where Q is the volume flux (normal to the reef) passing over the reef crest, ρ is the density of seawater and g the acceleration due to gravity.

The elevation Δh serves to drive water over the reef flats into the lagoon. The usual situation is that water exits from the lagoon through breaks in the reef. This flow is also at least partially driven by the wave set-up Δh so that the entire lagoon has a water level elevated with respect to the ocean. Water enters the lagoon by transferring a part of its kinetic energy into potential energy. Water exits the lagoon by converting potential energy back to kinetic energy just as in normal pipe flow.

The elevation decreases along the current trajectories through the lagoon to provide the force necessary to overcome bottom friction. Neglecting Coriolis forces, and any tidal or wind effects, the force balance is simply

$$-g \, h \frac{d \, \Delta h}{dx} - uh \frac{du}{dx} = C_D \, u^2$$

where x is a coordinate measured along the trajectory, h is the water depth, and u is the depth-mean velocity along the x-direction. C_D is the bottom drag coefficient.

Because h is minimal over the reef flat, u must be maximal here in order to maintain the continuity of volume flux. Hence, the frictional term is largest over this region and Δh decreases rapidly. Away from the reef flat, Δh changes more slowly and the final reduction of Δh to zero occurs across breaks in the reef which serve as current exits to the ocean.

Flows which are driven by a balance between pressure forces and bottom friction tend to be strongest in deep channels. This is simply because the total force on the water column due to pressure gradients, i.e. elevation change, is proportional to h. Figure 4.1 shows a typical bathymetric cross-section from the shore to the reef. A channel of depth 2 m is evident close to the shore. Such channels are not uncommon in coastal lagoons behind coral, and other, barrier reefs and are referred to by a variety of names including 'moat', 'gutter', and 'drainage channel'. Certainly, currents within such channels are often found to be substantial and clearly evident to the eye, i.e. 0.1 to 0.5 m s^{-1}. They tend to form by a natural process of sand scour through reefs fringing the shore. Casual observation by one of the authors (CJH) shows that such currents will typically move, and eject, coral debris of diameter up to about 0.15 m. The dominance of water motion through these channels is due to the pressure-friction balance which has been described.

The motion of water within the channels can be easily demonstrated mathematically. This is most simply achieved by a purely alongshore model. In the real coral reef circulation problem, inward current trajectories exist everywhere along the reef. They converge to form an increasing alongshore current which then exits through a break in the reef. However, the simple alongshore model assumes this alongshore current to be fully formed and neglects any further across-reef contribution. It also assumes that the 'next' break in the reef is far removed from the region of interest.

Let the y-axis be parallel to the shore with the x-axis drawn cross-shore. In the alongshore model the alongshore current u and depth h are functions of x only and there is no cross-shore current. The elevation Δh is therefore constant with respect to x, i.e. there is no cross-shore pressure gradient. In fact, Δh varies linearly with y and the force balance is

$$g\, h(x) \frac{d\, \Delta h}{dy} = C_D\, u^2(x)$$

for a quadratic bottom friction law with drag coefficient C_D. This equation indicates that u is proportional to the square-root of h. It is usually better to replace the quadratic law by a linear form so that the force balance gives

$$u_o(x) = \frac{g\, h_o(x)\ d\ \Delta h/dy}{C_D\, u_f}$$

where the subscript zero implies a particular value of water level and u_f is a 'background' velocity due to wave and other processes. Thus u is proportional to h and total transport varies as the square of h.

Figure 4.1 gives a curve of total volume transport derived using this square-law (with an elevation change of 0.05 m in 10 km) for the bathymetric section which is shown and water level at AHD. The dominance of the channel (defined as the region offshore to 1000 m) is evident with about 70% of the volume flow occurring in the channel. The process is very similar to that encountered in the wind flushing of a bay or harbour by a deep channel (Hearn et al. 1986).

Direct observations of channel flow have been made by Marsh et al. (1981) in two lagoons in Guam which resemble fairly closely the topography of the Ningaloo Reef tract. One of these is shown in Figure 4.2. An important difference from Ningaloo is that the lagoon is closed by rocky headlands at each of its ends. However, the lagoon is otherwise similar with very shallow water (of order 1 m depth at Guam and 2 m at Ningaloo) and a nearshore channel (or 'moat'). The tidal range is 0.4 m to 0.9 m (neaps and springs, respectively) at Guam which is also about a factor of 2 lower than at Ningaloo (0.8 m and 2.0 m, respectively).

Figure 4.2 shows the current trajectories (solid arrows) deduced by dye studies in Tumon Bay, Guam. Surf driven water moves across most portions of the reef margins and flows shoreward gradually changing direction; it flows as an alongshore current in the deep channel adjacent to the shore. The alongshore current eventually exits through the main break in the reef (San Vitores Channel). There are also minor breaks in the reef which carry outflow only during heavy surf conditions.

Current speeds in Tumon Bay moat ranged up to 0.6 m s^{-1} and landward flows on the outer reef flat reached 0.3 m s^{-1}. Flows through San Vitores Channel were 1 m s^{-1} under heavy (but not maximal) surf. These velocites are not untypical of currents across coral reefs (e.g. Maragos 1978; Frith 1983).

Marsh et al. (1981) found that a large (but rather variable) percentage of the volume transport occurred within the moat; a typical value for Tumon Bay is 40%.

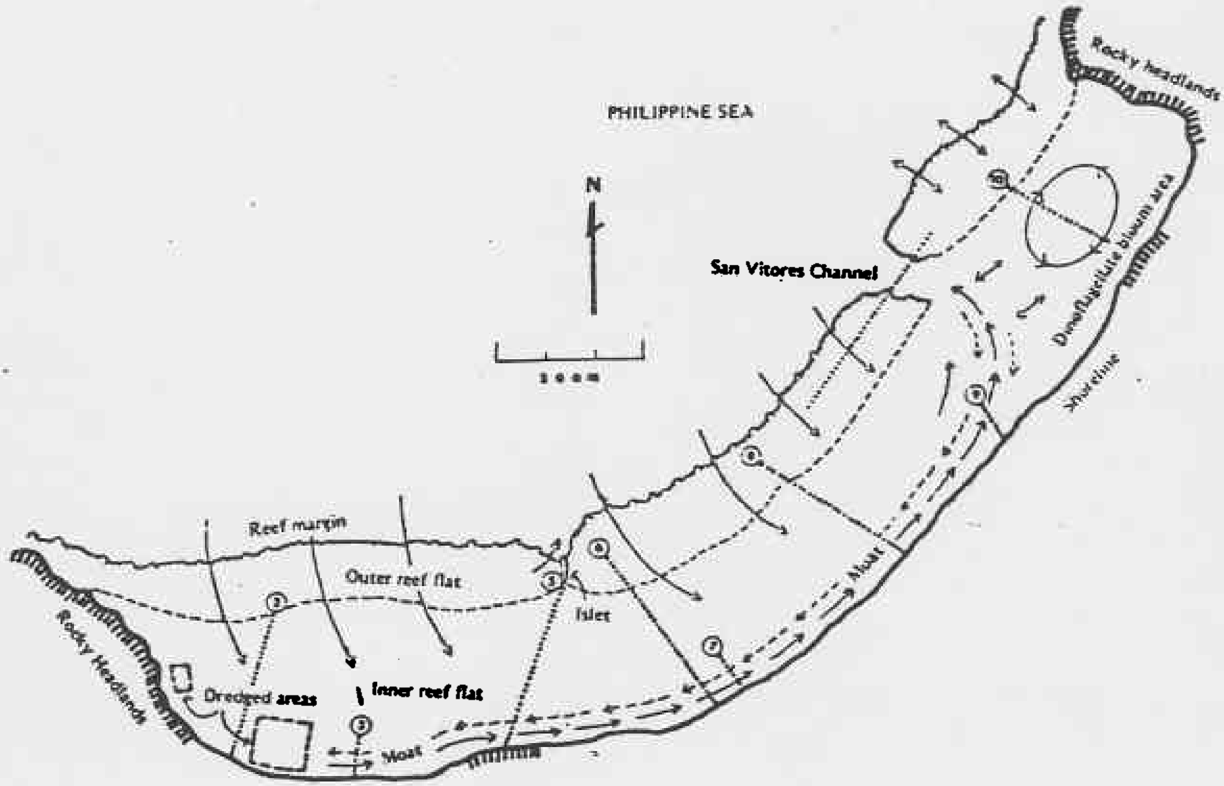

PHILIPPINE SEA
N
Rocky headlands
San Vitores Channel
Dinoflagellate bloom area
Shoreline
Reef margin
Outer reef flat
Islet
Moat
Rocky Headlands
Dredged areas
Inner reef flat
Moat

Figure 4.2

It is very evident, both from the simple alongshore model developed here, and the data of Marsh et al. (1981), that the fraction of transport in the channel (or moat) is a function of tidal height. If the water depth h(x) is expressed as

$$h(x) = h_o(x) + h_{tide} + \Delta h$$

where h_o is the depth relative to a datum (say, AHD), such that it is positive at all x, and h_{tide} is the water level relative to the datum, u(x) from the previous equation becomes

$$u(x) = A(h_o(x) + h_{tide})$$

where

$$A = (g\ d\ \Delta h/dy)/C_D\ u_f$$

This relation only applies for $h_{tide} + \Delta h > -h_o(x)$ and u(x) is otherwise zero. However, note that the constant A is proportional to d $\Delta h/dy$ which may change with tidal height. The equation shows that the profile of u(x) is proportional to the actual water depth h(x); transport is proportional to its square. Thus if, for example, water level in Figure 4.1 drops from AHD to AHD - 0.5 m the fraction of transport in the channel increases to almost 100%.

Tidal variation has two additional effects. The first concerns currents induced by the ebb and flood of water between the ocean and lagoon. This is treated in the next section. The second involves the variation of wave-pumped volume transport (into the lagoon) with tidal height. This is an extremely non-linear relationship. Certainly, there is some critical water level at which the barrier reef crest becomes totally dry (for a given height of wave field). The quantity α, introduced earlier (to denote the fraction of incident wave energy removed by the reef), then becomes unity and Q is zero.

For a situation in which the reef is well covered by the incident waves, the relationship between Q and Δh (in the energy flux equation) is probably controlled by the lagoon. This involves the variation of friction along the current trajectories through the lagoon and out to the ocean via the breaks in the reef. Marsh et al. (1981) collected data which show large variations in volume transport (up to 2 orders of magnitude) with tidal height. However, it is clearly a problem which can only be rationalised by the application of a numerical model to the real bathymetry. This would include the incident wave energy and tidal height (in the ocean) as its major parameters. The model would then proceed to calculate time-dependent currents throughout the system including the effects of drying of reefs and variation of friction with water depth and sea-state. The model would then be calibrated, and verified, by comparison with field data.

Although the detailed application of numerical models to wave-pumping of currents in shallow coral barrier reef coastal lagoons lies as a future research project it seems that the basic physics of the processes is already understood. Breaking waves on the reef crest drive water through the lagoon by creating an elevated water level. Water enters across the reef crest and moves shoreward. Near to the shore, it joins an alongshore current flowing through a deep channel (or moat). These channels eventually move offshore to exit the lagoon through the breaks in the reef. These exit currents are similar to the 'rips' generated by littoral currents flowing along beaches due to breaking waves on the shore. The strength of the exit currents increases with the surf state.

Let β be the fraction of the reef line which is occupied by breaks. Measurements on aerial photographs of the northern sector of Ningaloo Reef suggest that $\beta \sim 0.15$ over the section from Tantabiddi Creek to Point Billie. This is based on breaks being identified as regions without visible surf at the time of the flight (water level was close to mean sea level).

Thus if u_{reef} is the current across the reef crest, and h_{reef} is the mean water depth over the reef, the exit current through the breaks assuming no tidally or wind driven flow is

$$u_{wave} = \frac{h_{reef}\ u_{reef}}{h_{break}\ \beta}$$

where h_{break} is the water depth in the breaks. Taking u_{reef} to be 0.5 m s^{-1}, as a typical value, and $h_{break} \sim 2$ m gives

$$u_{wave} = 1.7\ h_{reef}$$

so that at high water springs, when $h_{reef} \sim 1$ m, u_{wave} may reach 1.7 m s^{-1} if our value of u_{reef} is reasonable.

Coral reefs which lie along coasts in which there are no natural control of breaks (such as river exits) may well assume a value of β which is determined by break-through of exit currents. Thus if β is initially very small any natural small breach in the reef will incur a large exit current. This will enlarge the break until the value of u_{wave} falls below the threshold for physical rupture of the reef structure.

There are other controls on the development of reef structure which can often determine the value of β. For a coastal lagoon, one (which is not operative at Ningaloo) is groundwater discharge at low water which places the coral under stress. β must be large enough to allow adequate flushing of this freshwater from the lagoon if the reef is to survive.

Figure 4.3 shows an aerial photograph of a portion of Ningaloo Reef. Clearly evident in the photograph is a structure of lines running across the reef towards the shore. The lines are about normal to the direction of the barrier reef and usually terminate in one of the darker strips running alongshore. These strips can be identified as the deep channels which have been the subject of discussion in this section. Near to breaks in the reef, the lines do not necessarily terminate in one of the channels but swing through an arc of 180° and exit through the break.

It is apparent that the lines are following the same routes that have been established for the current trajectories. The lines must then be maps of the current trajectories which are created by processes of erosion through the reef and scour over the mobile sandy substrates.

The material which cuts these lines through the system is probably produced biologically and removed by wave energy. The lines serve as a long-time average of the bottom current trajectories. The lines virtually cover the whole lagoon up to the coast and this is simply a consequence of the proximity of the coast to the barrier reef. Similar lines within a lagoon of the size of the Great Barrier Reef, off northeastern Australia, would map only a fraction of its lagoonal currents.

The lines represent the tracks of particles originating on the reef crest and therefore are presumably specific to wave-pumped currents. This, together with the natural long-time averaging which is inherent in the formation of the lines, could make the line pattern superior to other

Figure 4.3

current trajectory data (for example, from dye or drogues) for understanding and modelling wave-pumped circulation.

The separation of the lines can be resolved down to a few metres in the photograph of Figure 4.3. However, a proper spectral decomposition would certainly be of value as it could contain important information on the wave processes on the reef.

An important feature of these lines on the Ningaloo Reef is that they suggest a northward alongshore motion in the channels. Individual lines tend to swing northward as they meet the channels. Also, the lines often run more northeasterly than in a direction strictly normal to the reef line.

The origin of this preferred direction presumably lies in the wave rose which is biased towards swell and wind waves from the south. It may also reflect direct wind-driven currents from a dominantly southerly wind. It is possible that the channels themselves tend to direct the current northward and have been formed in this orientation by a history of scour and erosion.

The channel-forming processes must include drift currents formed by breaking waves on the shore and its fringing reefs. These result from wave propagation through the breaks in the reef and are also responsible for the formation of embayments and headlands. The processes are illustrated in Figure 4.4. Apart from diffraction through holes in the reef there are processes of refraction over the reef crest which are important at high water.

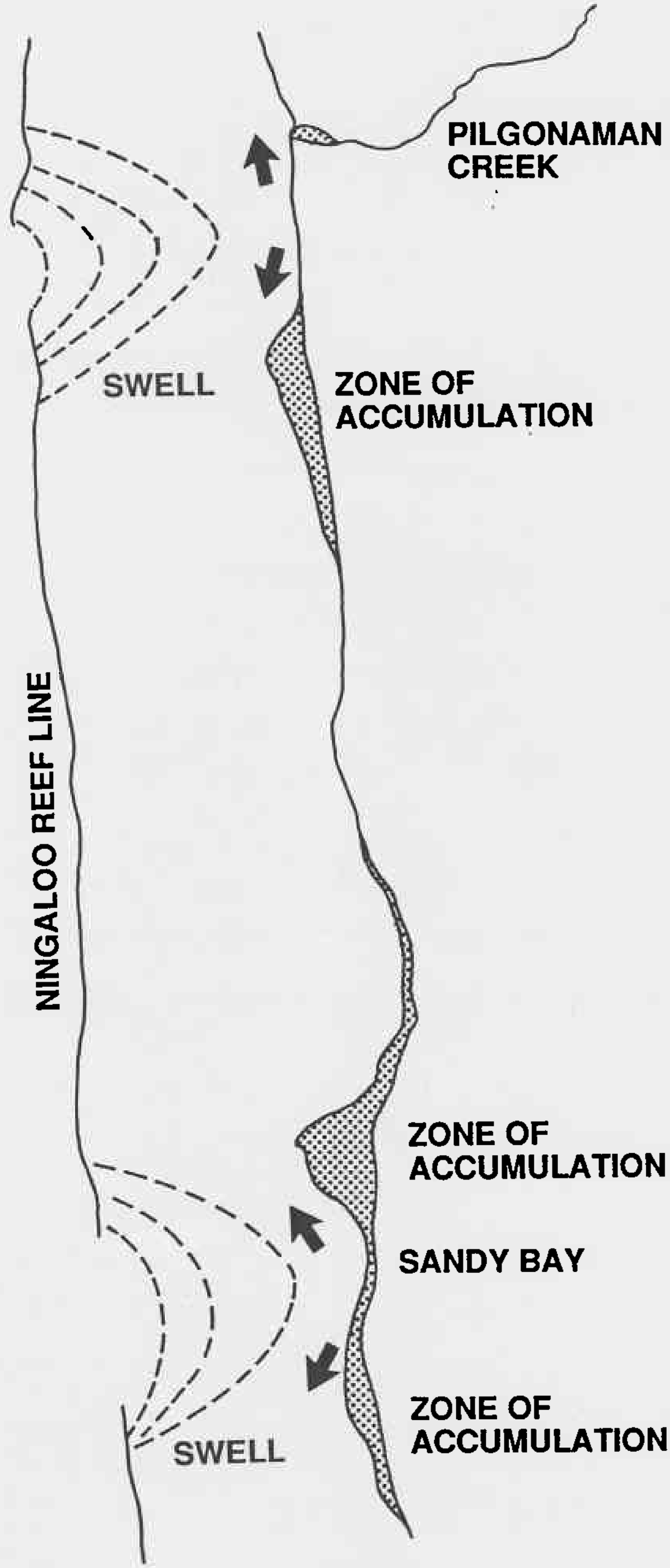

PILGONAMAN CREEK
SWELL
ZONE OF ACCUMULATION
NINGALOO REEF LINE
ZONE OF ACCUMULATION
SANDY BAY
ZONE OF ACCUMULATION
SWELL

Figure 4.4

4.2 Tidal Currents

Tidal elevation changes in the ocean produce corresponding changes in the lagoon with some attenuation and phase lag. In a long coastal lagoon like the Ningaloo Reef lagoon there are two aspects to the tidal currents. The first is the propagation of an alongshore tidal wave through the lagoon with only secondary cross-shore currents. The second involves cross-reef processes acting to exchange water between the lagoon and open ocean; the alongshore wave propagates in the deep water beyond the reef. In this second tidal mode there are, of course, alongshore currents in the lagoon since the cross-reef flow is not spatially uniform but controlled by local topography. The second mode would occur for example in Tumon Bay (Figure 4.2) where the lagoon is closed at its ends by headlands.

Because the Ningaloo Reef lagoon is very shallow the speed c of a free gravitational wave is low, e.g. for $h = 1$ m, $c \sim 3$ m s^{-1}. Hence the transit time from one end of the lagoon to the other (~ 200 km) would be about 18 hours. Additionally the wave would be highly damped. It therefore seems reasonable to suppose that this tidal mode plays a negligible role in Ningaloo Reef lagoon.

Thus given a local change in the level of the ocean, currents cross the reef line due to pressure forces created by a difference in water elevation between the lagoon and ocean. The major problem is to determine whether these tidal currents enter predominantly through the breaks in the reef or across the reef crest.

At low water there is only one route, and that is through the breaks, because the reef crest is dry. This is also the condition for which tidal currents are most evident simply because wave pumping has ceased; higher velocities are also necessary to carry the volume fluxes through the shallow water.

Inside the lagoon the tidal flow is controlled by a balance of pressure and bottom friction. It therefore favours the same channels as the wave-pumped flow. The difference is a change of direction with water entering (rather than leaving) through the reef breaks. This is followed by a flow into the lagoon through the channels. The dashed arrows in Figure 4.2 show this reversed flow for Tumon Bay, Guam, at low tide. Measurements made by Marsh et al. (1981) show that the reversed flow is similar to the normal flow in that it occurs dominantly in the moat and typical values are 50% of total transport.

The tidal range inside the lagoon may be smaller than in the open ocean because of the elevation difference necessary to drive water through the reef. This will also result in a phase difference with the lagoon tide lagging behind the ocean tide. For the situation where the reef crest is dry the lagoonal tide is controlled by the value of β (the fraction of reef occupied by breaks). It is important to note that a tide gauge deployment inside the lagoon gives data on the lagoonal tide and this may differ significantly from the open ocean tide. There may also be a spatial variation along the lagoon.

If the tidal range in the lagoon is H and the period T, and in the absence of wave driven flow over the reef crest the mean tidal current through the break is

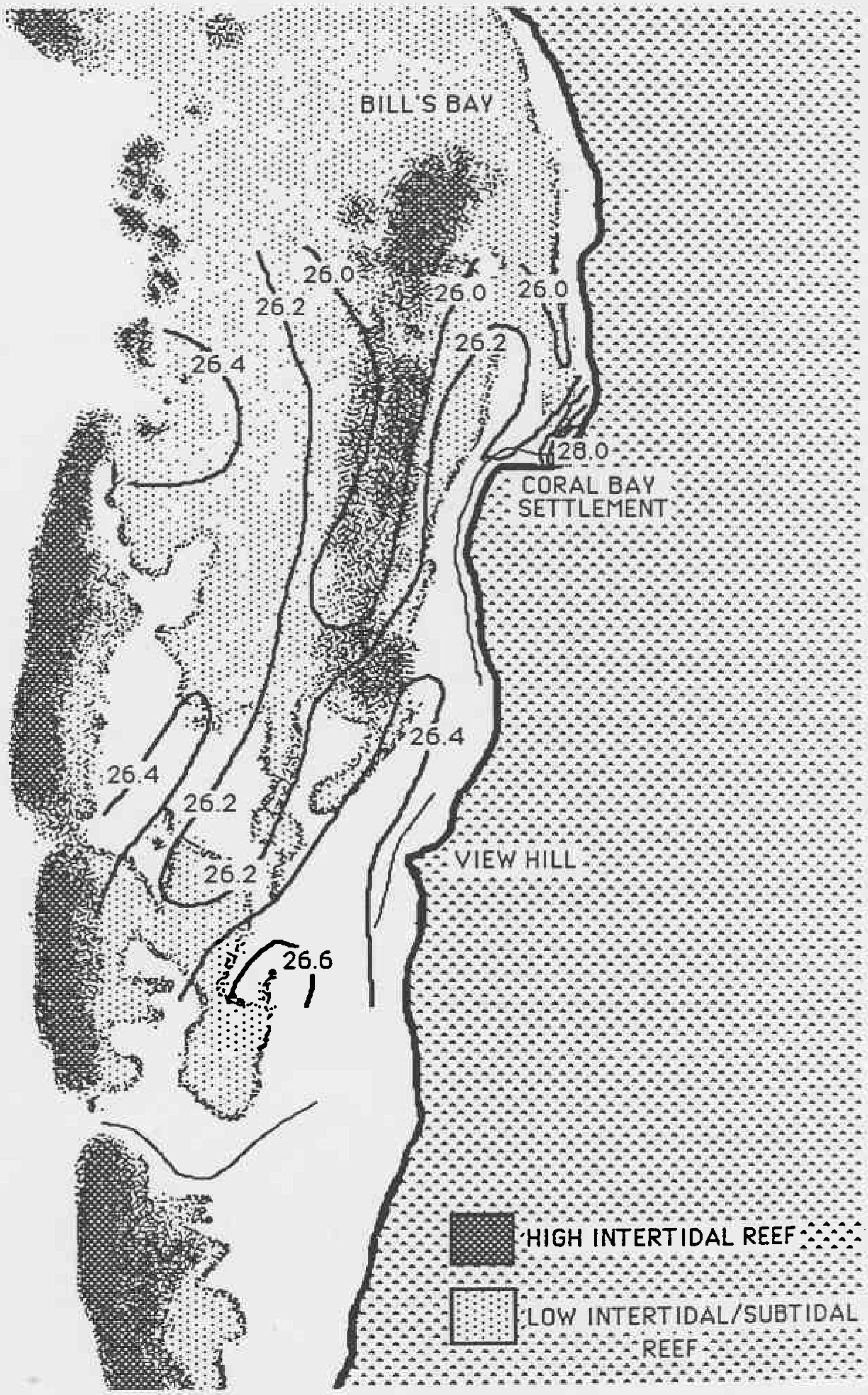
$$u_{tide} = \frac{2\,H\,w}{T\,h_{break}\,\beta}$$

where w is the width of the lagoon and h_{break} the water depth in the reef breaks. Using $w \sim 2 \times 10^3$ m, and $h_{break} = 2$ m, this gives for a semi-diurnal tide with H = 2 m, $u_{tide} = 0.6 \text{ m s}^{-1}$.

Figure 4.5 shows a map of surface temperature inside the Ningaloo Reef lagoon in the vicinity of Coral Bay Settlement. The map was obtained in early morning on a rising tide. Night-time cooling has depressed water temperatures within the lagoon relative to those of the open ocean. There appears from the map to be a temperature signal of a few tenths of a degree celcius which is acting as a tracer of ocean water tidally flooding the lagoon. Certainly the map provides support for the tidal model in which water enters through breaks in the reef and then follows the channels alongshore. It also shows the dominant northward alongshore motion discussed in the previous section. This result emphasises that the northward motion is a result of the orientation of the channels. Full details of this temperature data set are given in Appendix I and the method is discussed in Chapter 5.

At higher values of water level tidal currents can pass over the reef crest as well as through the breaks in the reef. Although pressure and friction are the dominant forces within the lagoon the flow through the channels created by reef breaks, and over the crest, are strongly influenced by inertial forces. A measure of the relative magnitude of such forces compared with friction is given by the Reynolds number for the reef defined as the ratio of inertial to frictional force,

$$R = \frac{h\ u}{2\ C_D\ u_f\ X}$$



BILL'S BAY
26.0
26.0
26.0
26.2
26.2
26.4
28.0
CORAL BAY
SETTLEMENT
26.4
26.4
26.2
26.2
VIEW HILL
26.6
HIGH INTERTIDAL REEF
LOW INTERTIDAL/SUBTIDAL
REEF

Figure 4.5

where h is a water depth, u current speed, u_f the background velocity, and X is the horizontal space scale. Sensible choices of typical values for these parameters in the reef breaks lead to R in excess of unity and probably of order 10. However, over the reef crest R may be lower due to wave breaking which increases u_f.

If the Reynolds R is much greater than unity over the reef crest and reef breaks, the ratio of volume transport through the breaks to that over the crest is

$$\beta(h_{break}/h_{crest})$$

If however friction is dominant on the reef crest the ratio is increased by a factor $R_{crest}^{-1/2}$. Therefore, as h_{crest} increases, there is a smooth transition to a flow pattern in which part of the tidal current flows over the reef. However, tidal flow over the crest may be largely obscured, or masked, by the larger wave-pumped current.

Detailed mapping of tidal currents flowing through gaps between reefs has recently been obtained by the HF radar surface Doppler technique (Heron et al. 1985). This shows that given suitable geometry a system of vortices forms behind the reef. This is a comparatively small-scale aspect of the tidal flow on the Ningaloo Reef tract but may be important in the interaction of tidal flow across the reef and through the breaks.

4.3 Wind Forcing

It has been shown in Section 3.2 that wind stress occurs predominantly from the south but with easterly and westerly components during certain seasons; occasional storm gales come from the north. This wind pattern is similar to that at points further south along the west coast. However, the circulation patterns induced by the wind at Ningaloo will be very much affected by the topography of the lagoon and reef. Discussion within this section is restricted to the effects of the southerly component of the wind; cross-shore winds are not considered here.

Flow driven by a southerly wind will consist basically of an anti-clockwise gyre with water moving northward (downwind) in the shallow lagoon and returning southward (into the wind) beyond the reef. This gyre has its origins in the depth differential between the lagoon and open ocean. Thus wind stress is dominant in the shallow lagoon whilst the resultant pressure gradient is the major force in the open ocean. These remarks apply to the depth-mean currents; surface currents will flow with the wind in both cases.

Additional to this main gyre there will be some flow through the reef line. This can, in principle, occur both over the reef crest and through the breaks in the reef. The relative magnitudes of these contributions depends on a number of factors and this has been discussed in Section 4.1. For the case in which flow only occurs through the breaks in the reef the circulation pattern will have the form shown in Figure 4.6.

The magnitude of the current in the lagoon can be estimated by making the assumption that the water is sufficiently shallow to require a force balance between wind stress and bottom friction thus

$$u_{wind} = \frac{\rho_A \, C_w \, U^2}{\rho \, C_D \, u_f}$$

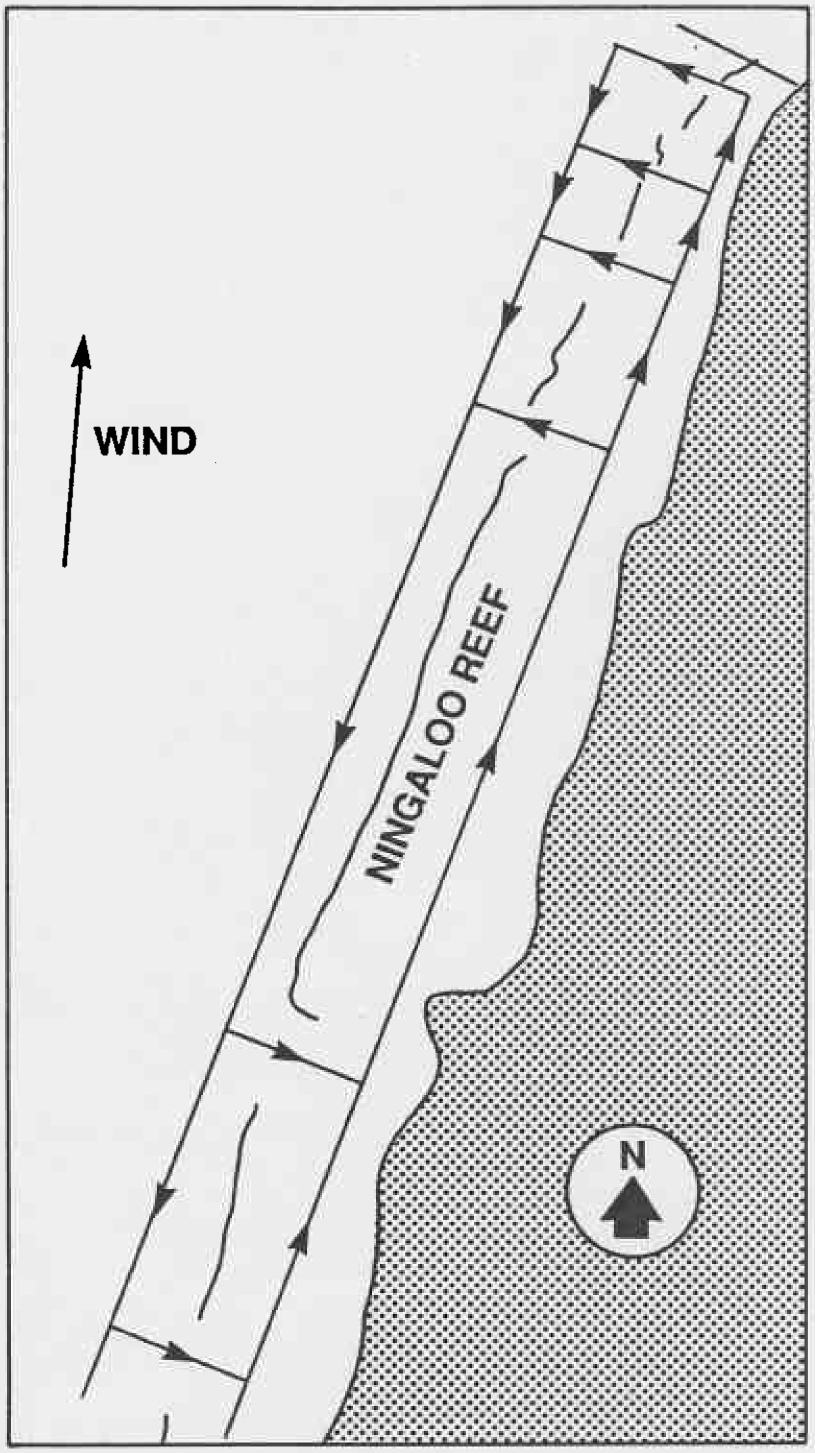


Figure 4.6

where ρ_A is the density of air (~ 1.3 kg m^{-3}), C_W is the surface drag coefficient (~ 10^{-3}), U is the wind speed and ρ is the density of water (~ 10^3 kg m^{-3}) thus

$$u_{wind}\, u_f \simeq 5 \times 10^{-4}\, U^2$$

Taking the annual mean value of U^2 for Carnarvon (U = 5.51 m s^{-1}, Table 3.1) and u_f = 0.1 m s^{-1} gives u_{wind} = 0.15 m s^{-1}.

This value of u_{wind} is smaller than tidal speeds at springs and lower than speeds induced by wave pumping (for moderate or high swell). The wind-driven currents will have a superimposed internal motion within the lagoon. This may consist of southward flow through the deeper channels and northward flow over the shallower reef tops. This type of structure is very uncertain because of the problem of friction over the reef and it is possible that flow will only occur (northward) in the channels.

The above value for u_{wind} may be an overestimate because of set-up forces within the lagoon; these are most likely to be caused by orientational changes in the channels as they exit the lagoon through breaks in the reef.

4.4 Flushing Times

There is considerable interest from all types of marine scientists in the flushing time of any water body (e.g. the dispersal of coral larvae in relation to settlement and recruitment). In reality this flushing time is not a well defined concept and consequently considerable confusion and ambiguity usually arises in calculations of its value. For the Ningaloo

Reef tract an operational definition of flushing time will be adopted: 'What is the average time taken for a particle released within one of the channel regions of the lagoon to exit through the reef line'. This of course gives no information on the associated process in which an ejected particle re-enters the lagoon across another part of the reef. In fact, existing knowledge of mixing processes immediately outside the reef is too limited to make any realistic attempt at calculating these return probabilities.

Three major processes are likely to be important in flushing the lagoon.

Wave Pumping: Flow occurs inward across the reef and mostly into the deep channels from which it is guided out through breaks in the reef. Inspection of the tracks (discussed in Section 4.1) on the aerial photographs suggests that only inflow occurring within about one lagoon width of a break takes the shorter route of a direct arc out through the break (without reaching the inshore channel). The final exit velocity through the break due to wave driven flow has been calculated in Section 4.1 as u_{wave}. Volume transport in the channel will vary approximately linearly with alongshore position, starting perhaps from a stagnation point, and reaching a maximum within a lagoonal width of a break. The corresponding speed depends, of course, on the width and depth of the channel.

Consider the simple alongshore model of Section 4.1 in which u increases linearly with x, i.e.

$$\frac{d}{dx}(A_{channel}\ u) = u_{reef}\ h_{reef}$$

where $A_{channel}$ is the cross-sectional area of the channel. If this is assumed constant together with the velocity and depth over the reef,

$$u = u_{wave} \frac{x}{l}$$

where l is the length of the section of unbroken reef. Hence the travel time t from x to the break is

$$t = \frac{l}{u_{wave}} \log \frac{l}{x}$$

If $u_{wave} = 1 \text{ m s}^{-1}$, $l = 10$ km, $x > 1$ km, $t \lesssim 6$ hours. This value of u_{wave} therefore gives the flushing time due to wave pumping as $\tau_{wave} \sim 6$ hours and this applies to moderate to strong surf conditions. It only applies when the tidal height and wave set-up covers the reef crest. However, given the mixed diurnal/semi-diurnal frequency of the tide the flushing time clearly remains below one day for these surf conditions.

Particular regions of the lagoon will clearly have much longer flushing times for wave pumping. This is due to the sheltering effect of headlands, varying speeds around fringing reefs, and trapping of water within deep holes. The ecology of systems may be as much controlled by these atypical regions as by the majority behaviour; it is necessary to derive a whole spectrum of flushing times and use those appropriate to a given ecological question (Hatcher et al. 1986).

Tidal Exchange: A standard calculation for the tidal flushing of an enclosed water body is the volumetric ratio time,

$$\tau_{volume} \equiv \frac{V}{\Delta V} T$$

where ΔV is the tidal prism (the volume difference between high and low water), V is the mean volume of the water body, and T is the tidal period. For the Ningaloo Reef lagoon ΔV lies roughly in the range V/2 to V/4 so that τ is one or two days.

For shallow water systems the actual tidal fushing time τ_{tide} is considerably in excess of τ_{volume}, because the incoming tidal prism does not mix totally with the water remaining in the system at the end of the last ebb tide. Instead, the interface between the residual water and the tidal prism tends to be more nearly vertical than horizontal. Thus, instead of the tidal prism filling space above the residual water it forms a front and the residual water increases its depth, and reduces its horizontal spread, as the tidal level rises. This front, or interface, moves inward on the flood tide and outward on the ebb.

There are two aspects to the calculation of the tidal flushing time in such systems. The first concerns the amount of mixing across the front during a tidal period. The second involves the fate of the tidal prism whilst it is external to the system. It is usual to assume that a new tidal prism enters the system on each flood tide; any pollutant collected by the prism during cross-frontal mixing on the previous tidal cycle does not re-enter the system. However, empirically, a 'return coefficient' r varying from zero (no re-entry) to unity (total return) can be included to represent this aspect of the flushing process.

Mixing across the front leads to a characteristic time

$$\tau_{front} = \frac{4\ l^2}{\pi^2\ D}$$

where l is the length of the system, and D is a dispersion, or mixing, coefficient; this is the time that it would take for all the water of the system to totally mix with the tidal prism. The main difficulty lies in estimating a value for D. Major contributions to D arise from vertical and along-frontal mixing processes. They depend upon the dynamics of the system, such as wave pumping and wind forcing, and are greatly affected by density differences between the two water masses.

If l = 10 km in the Ningaloo Reef lagoon and D is taken as 100 $m^2\ s^{-1}$ then τ_{front} = 4.7 days. This value of D is probably applicable to conditions of strong wave pumping. The inclusion of a return coefficient differing from zero has virtually no affect on τ_{tide} for such a value of τ_{front} and so

$$\tau_{tide} \sim 5 \text{ days}$$

It may be more appropriate near high tide to replace l by the width of the lagoon which varies from 1 to 2 km. In this case τ_{front} becomes much shorter, i.e. 1 to 5 hours. The tidal flushing time is then limited by τ_{volume} and the value of r, i.e.

$$\tau_{tide} \rightarrow \frac{\tau_{volume}}{1 - r}$$

Hence a typical value of r ~ 0.5 leads to τ_{tide} being of value 2 to 4 days which is not too dissimilar to the earlier value using l = 10 km, which gives confidence in this value of τ_{tide}.

Wind Flushing: Surface wind stress acts to flush the Ningaloo Reef lagoon by driving water within the lagoon in a downwind direction. Such water crosses the reef to form a return current outside the reef (Figure 4.6). The process relies on the depth difference between the lagoon and ocean; wind-stress is dominant in the lagoon and alongshore pressure gradients in the ocean.

It is estimated in Section 4.3 that the mean annual wind-driven current in the lagoon is 0.15 m s^{-1}. This is probably an overestimate but it seems fair to adopt 0.1 m s^{-1} as a working, order-of-magnitude, value. It is less easy to calculate a volume flux because so much is uncertain about flow over the top of fringing reefs within the lagoon. However the definition of flushing time adopted here only involves flow through the channels. A particle released into a channel with speed 0.1 m s^{-1} will take a time of 10^5 s (27.8 hours) to exit the lagoon if it is required to travel 10 km.

This places wind flushing as somewhat faster than tidal flushing. However wind flushing, like wave-pumped flushing, is subject to uncertainty regarding the subsequent route taken by water expelled through the breaks in the reef.

Considering the overall flushing problem for Ningaloo Reef lagoon it seems that wave pumping is the dominant process. For moderate to strong surf conditions, with the reef covered, the flushing time, as defined here, is less than 6 hours. Thus, given that the reef crest is exposed during a part of most tidal cycles (Simpson and Masini 1986), it seems safe to quote half a day as the usual flushing time. However there is a natural variance to the incoming wave field. It would be necessary to look at the detailed annual statistics of the wave rose in order to assess the recurrence period of longer flushing times.

Tidal flushing appears to be 2 to 5 days under springs tides and a factor of two longer at neaps. Wind flushing is estimated at 1 day as an upper limit for mean annual wind stress. Wind flushing is also subject to the natural variance of wind events.

By the standard of most coral reef systems studies to date, the Ningaloo Reef lagoon appears to have a very short flushing time. This is really just a consequence of the shallow nature of the lagoon and its limited width. This may place the reef in a special category in regard to its ecology and environmental stability.

A further matter of importance here is the vertical mixing time. Because the lagoon is so shallow and current speeds are high, the vertical mixing time tends to be very small; a value of less than 1 hour is an estimate. It would therefore be anticipated that the water column should always be well mixed.

The only piece of physical data available to support the present calculations of flushing times within the Ningaloo Reef lagoon is a comparison by Simpson and Masini (1986) of lagoon and ocean temperatures. This was based on mean weekly data collected from fixed instrument deployments in the lagoon and satellite (GOSSTCOMP) surface temperatures. For approximately 5 weeks, distributed over May to December 1986, the agreement between the temperatures was good ($r = 0.98$). This corresponds to a 5°C variation in ocean temperature which is near to the annual range at this latitude (Section 2.2) for the western coast of Australia. This confirms that the flushing time is less than 1 week.

4.5 Temperature Anomalies

Simpson and Masini (1986) have found temperature anomalies in the Ningaloo Reef lagoon which involve the water temperature being depressed, below the seasonal values, for 2 or 3 days. An example is shown in Figure 4.7. The magnitude of the anomalies is about 1 or 2°C, and it appears that the events usually occur near neap tides. Simpson and Masini point out that this is the time of maximum wave pumping because the reef crest is covered during the whole of the tidal cycle. They therefore conclude that the anomalies are probably due to advection into the lagoon (over the reef crest) of a mass of cold water from the open ocean.

The anomalies may have their origin in the entry of cold water masses into the lagoon. Under moderate to high surf conditions the flushing time of the lagoon is only 6 hours by wave pumping (Section 4.4). Thus a sudden drop of water temperature outside the reef would give an identical signal inside the lagoon within a time of about 12 hours. The time for onset of the anomalies in the Simpson and Masini (1986) data certainly seems to be of this order. Thus the onset time would seem to be limited by the transfer process into the lagoon rather than by the time for the ocean temperature to drop. This means the ocean temperature must drop by about 1°C in 12 hours.

There are two obvious scenarios for the change in ocean temperature. The first is that a surface temperature front arrives at the reef causing cold water to advect into the lagoon. However, to give an anomaly of

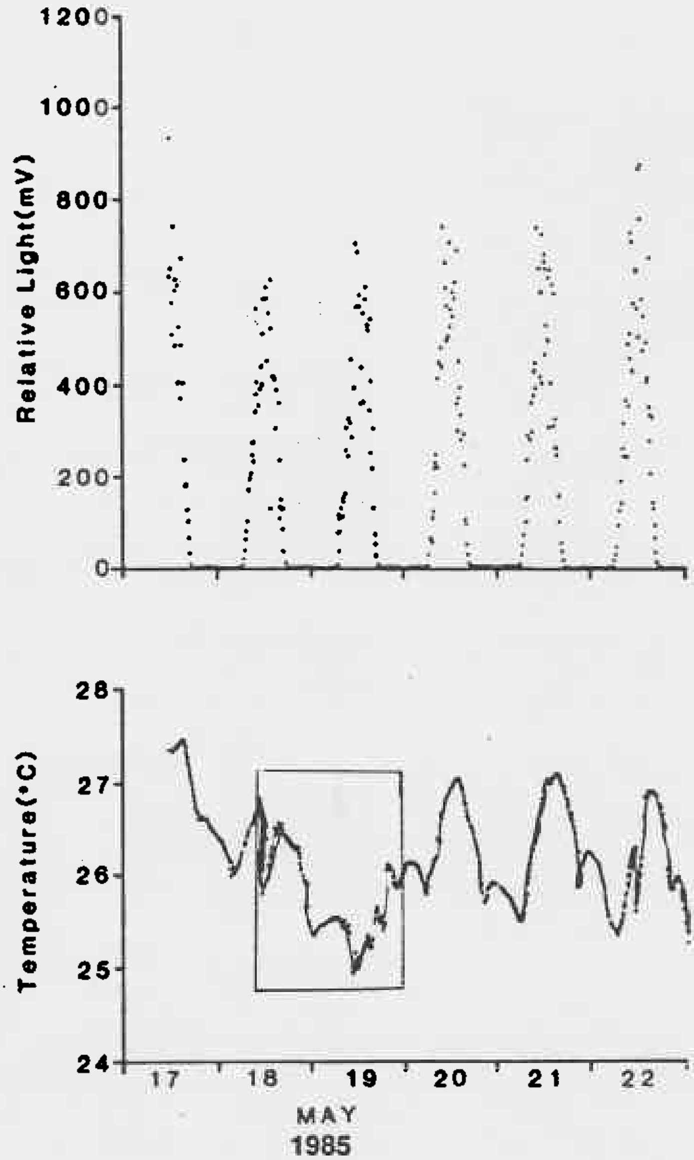

1200
1000
800
600
400
200
0
Relative Light(mV)
28
27
26
25
24
Temperature(°C)
17
18
19
20
21
22
MAY
1985

Figure 4.7

duration 1 or 2 days, the front must be followed by recovering to normal sea surface temperature. If the front moved onto the reef at 0.1 m s^{-1} the temperature gradient would be 1°C in 4.3 km which seems very high. It is feasible that such fronts are produced by the Leeuwin Current near North West Cape, but the anomalies also occur outside of the winter season when the current is not flowing.

The second scenario for the arrival of cold water at the reef is that upwelling has occurred. A drop of 1°C involves a vertical displacement of only 20 m (based on Exmouth Plateau data; Holloway, 1986c). The previous discussion implies that the upwelling occurs in the immediate vicinity of the reef; it is not upwelled and then advected.

Figure 4.8 shows weather maps for Australia for May 13 to 20, 1985 to encompass the period of the anomaly in the Ningaloo Reef lagoon from May 18 to 19. It shows a fairly steady weather pattern over the northwest and there is certainly nothing to suggest a dramatic change in wind or pressure at the time of the anomaly. The wind appears to be easterly or northeasterly. This is not likely to produce the northward flow necessary for classical upwelling although the offshore component might be effective in moving surface waters away from the reef. The difficulty is that this offers no explanation for the sudden occurrence of upwelling.

Local upwelling is known to be caused by vortex, or eddy motion, in the lee of islands (Takahashi et al. 1981; Wolanski et al. 1986). Such eddies evolve very quickly in a time-dependent current such as the tidal currents on the Great Barrier Reef. The upwelling is due to an internal secondary motion within the eddy which forces bottom water to the surface at the centre; downwelling occurs near the edges. It is also responsible

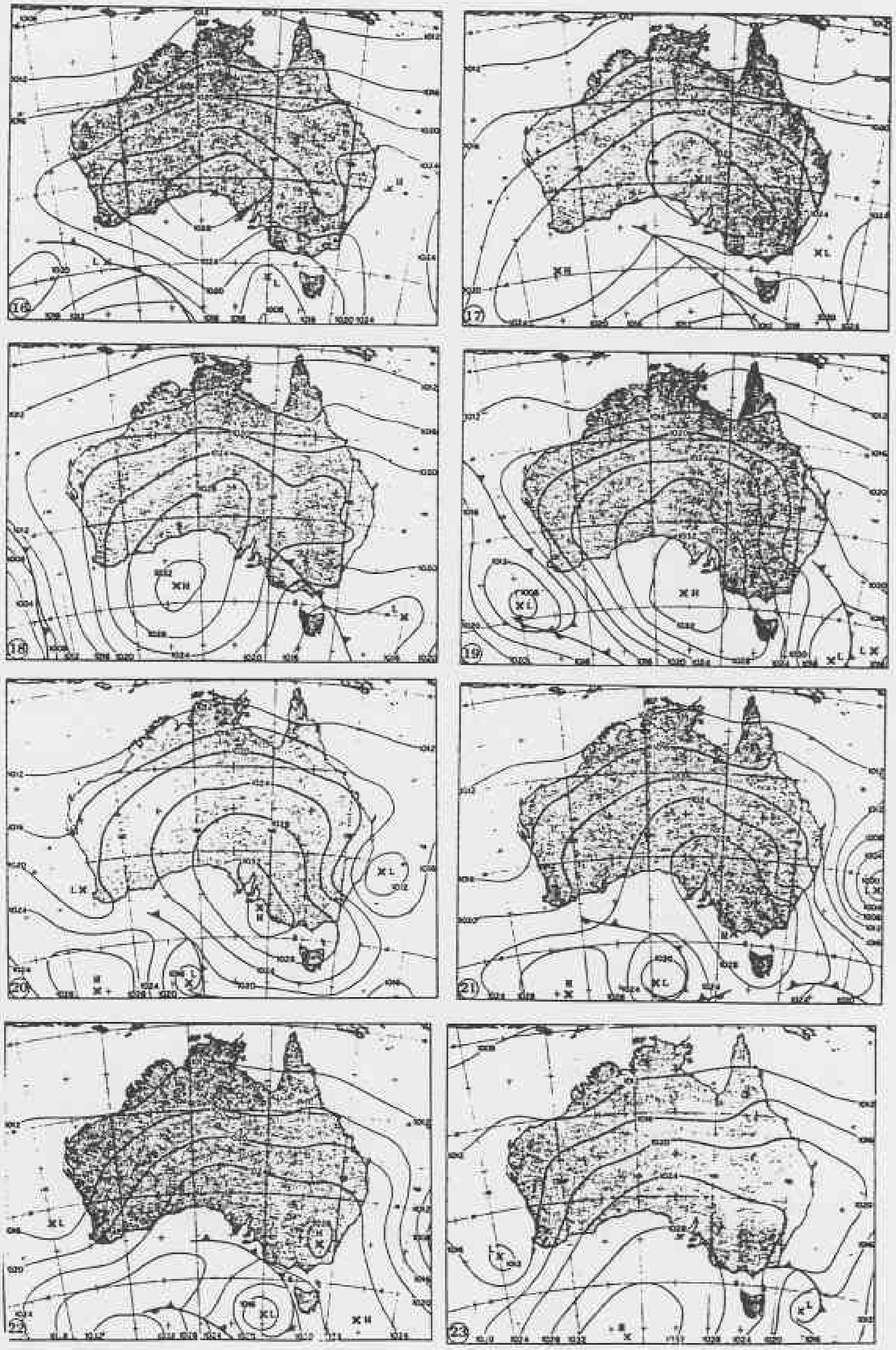


Figure 4.8

for upwelling by the Gulf Stream flowing over seamounts and along the continental shelf edge (Pingree 1978). However the motion would need to be strong enough to overcome density differences through the water column.

Two sources of eddies would be the Leeuwin Current or tidal streaming near the reef. The former is unlikely because anomalies occur outside the usual season for the current. Tidal sources for the eddies are possible but would be recurrent.

The most likely explanation of the transient upwelling which seems to generate the temperature anomalies is the arrival of an internal wave of suitable amplitude. Nearly all internal waves in this region are semi-diurnal so that the full vertical rise of an isotherm takes only 6 hours. This is consistent with the onset time of the anomalies.

Internal waves have zero displacement at the surface. Water entering the lagoon would originate from the top 10 m or so of ocean water. Hence an internal wave which could produce the type of temperature changes necessary for the anomalies would need to be of very large amplitude. It is possible that the waves are forced into a very non-linear state by the shoaling of the shelf or are breaking (Wallace and Wilkinson 1981) or undergoing the type of internal hydraulic jump reported by Holloway (1986b) on the North West Shelf.

One of us (B.G.H.) has observed temperature anomalies at the Abrolhos Islands. An Applied Microsystems pressure and temperature sensor, sampling for 4 minutes every half-hour at a frequency of 4 Hz has been deployed in 4 m of water at the northern end of the west reef of the Easter Group (21° 43' 30"S, 114° 43' 00"E). It shows anomalies of several degrees celcius about once per month between December and April. A very superficial inspection of meteorological data (from Geraldton) does not suggest that these events are linked to unusual, or specific, wind conditions. Furthermore the anomalies do not seem to occur at a special point in the spring-neap cycle. Since the Abrolhos Islands are situated on the shelf break (like the Ningaloo Reef tract) it is tempting to assume that the anomalies have a similar origin.

The onset and recovery times of the anomalies at the Abrolhos Islands are much greater than on the Ningaloo Reef tract. The onset time is typically several days. This could either be due to the time required to advect water into the lagoon, i.e. a time of order of the flushing time, or the time for build-up of the cold water body externally to the lagoon. The Easter Group lagoon differs from the Ningaloo Reef lagoon in being much deeper (about 10 m compared to 2 m for Ningaloo) and having a greater lagoonal area per length of windward reef. These factors increase the flushing time of the bulk of the lagoon. However, the instrument location near to the reef implies that it is sensing a water mass which is flushed more rapidly than the bulk of the lagoon. Thus an onset time of 2 or 3 days would seem to be reasonably attributed to the advection/mixing processes near the reef.

The recovery time of the anomaly in the Easter Group lagoon is 3 or 4 days which is again consistent with the flushing time at the location of the instrument.

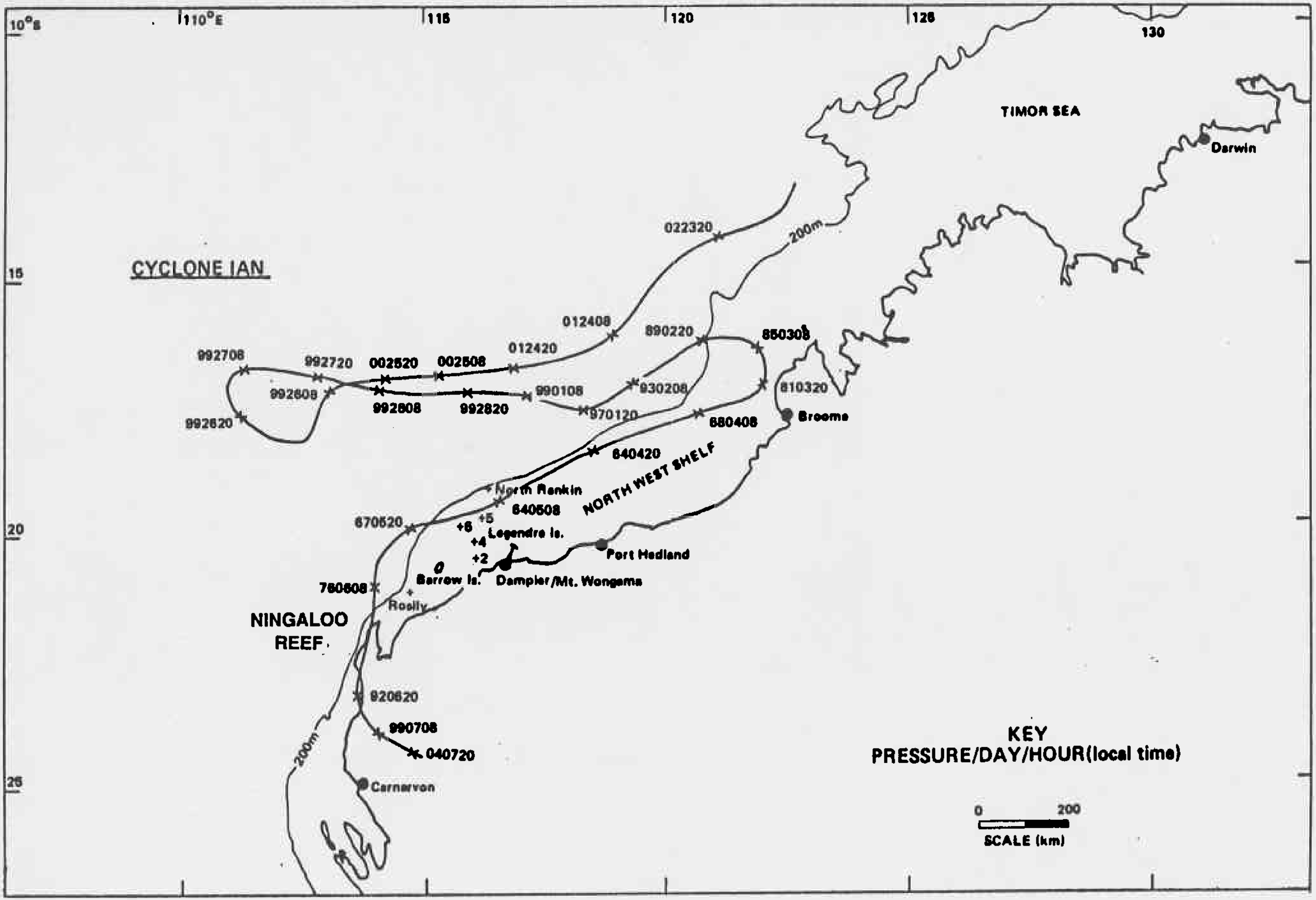
Anomalies of the type discussed here do not seem to have been observed on the Great Barrier Reef. However, Wolanski and Pickard (1983) have noted such changes in deeper water due to internal waves. Upwelling and downwelling occur periodically on the Queensland shelf and accompany long-period waves with a wide range of frequencies (Andrews 1983). These produce onshore surges of shelf-break water which is 1 to 4.5°C cooler than lagoon water (depending on the season). This upwelled bottom water is enriched in nutrients, which play a fundamental role in the functioning of reefal ecosystems.

Thompson and Golding (1981) have suggested that the very strong tidal currents which flow through breaks in the Great Barrier Reef may serve to induce upwelling. This mechanism relies on the reef being situated on the edge of a very steep shelf break. However, tidal processes are recurrent and do not therefore seem to be the origin of the temperature anomalies discussed here.

4.6 Tropical Cyclones

A tropical cyclone develops from a tropical depression formed over the sea. It may exist for weeks over the ocean but once it crosses land it very quickly begins to dissipate and loses much of its intensity and characteristics. Nearly 10% of the global total of tropical cyclones develop in the warm tropical water off northwest Australia. In a typical cyclone season, from November to April, about seven to 10 cyclones form between the Cocos Islands and Darwin; two to three may make landfall on the Western Australian coast.

Cyclones have gale force winds (speed greater than 17.5 m s^{-1}) which typically reach 50 m s^{-1}. These can produce wind waves of the order of 10 m height. Most cyclones on the northwest coast have minimum central pressures of about 950 mb. This produces a sea level change which would be



10°S
110°E
115
120
125
130
15
20
25
CYCLONE IAN
TIMOR SEA
Darwin
200m
022320
012408
012420
890220
850308
810320
Broome
992708
992720
002520
002508
992608
992808
992820
990108
930208
970120
880408
992620
840420
NORTH WEST SHELF
North Rankin
840508
670620
+6
+5
+4
+2
0
Legendre Is.
Port Hedland
Barrow Is.
Dampier/Mt. Wongama
760608
Rosily
NINGALOO
REEF
920620
990708
040720
Carnarvon
KEY
PRESSURE/DAY/HOUR(local time)
0
200
SCALE (km)

Figure 4.9

about 0.5 m in isostatic equilibrium in the open ocean. Major sea level changes are also produced by the wind. The actual dynamics of these processes is a topic of current research in oceanography. At a coastal crossing a cyclone typically produces a sea level surge of up to 1 m. Coastal rainfall is typically some hundreds of millimeters during the few days of the passage of the cyclone.

A survey of Australian tropical cyclones from 1909 to 1980 by Lourensz (1981) showed a mean incidence of 8.9 cyclones every ten years within the 5° latitude/longitude square enclosing the Ningaloo Reef tract. The total decadal incidence of coastal crossings by cyclones was about 0.5 per 100 km of coast at Ningaloo, i.e. about 2 per decade for the complete reef.

Figure 4.9 shows the path of cyclone Ian (March 1 to 6, 1982) taken from Lynch (1982). The numbers on Figure 4.9 give information on the time and day at which the cyclone occupied a particular position and its central pressure p. Thus, e.g. 640420 means $p = 964$ on March 4 at 20.00 hours. This cyclone crossed the reef several times. Ian reached peak intensity on March 4 while offshore from Broome with estimated central pressure of 964 mb and maximum wind of 40 m s^{-1}.

Holloway and Nye (personal communication) have analysed the oceanographic response to cyclone Ian using data collected for the North West Shelf gas project. Currents were measured at five moorings across the shelf from Dampier. On the shelf, barotropic currents of up to 0.8 m s^{-1} were observed with a current of 0.3 m s^{-1} at 3 m above the sea bed in 70 m of water. On the shelf break (North Rankin) currents were much weaker (although the cyclone passed directly over this station) and did not exceed the Leeuwin Current in magnitude. Near surface currents over the shelf were about 0.5 m s^{-1}. A summary of current meter data from the Dampier Archipelago during the same cyclone is given by Mills et al. (1986) and shows peak currents of 0.5 m s^{-1}.

Temperature profiles on the North West Shelf before and after the passage of cyclones have been studied by Holloway and Nye (pers. comm.). For 12 cyclones, for which data exists between 1979 and 1984, all produced longshore currents towards the southwest. This direction favours downwelling (rather than upwelling) although only three of the cyclones did produce downwelling on the shelf break (of which cyclone Ian was one). Only one of the cyclones produced significant vertical mixing over the shelf; for Ian the waters were already well mixed.

It therefore appears that the possible oceanographic effects of cyclones on the Ningaloo Reef occur only through the special current regimes. The direct currents induced on the open shelf are not significant compared with the normal currents experienced in the lagoon; they would also be frictionally limited in shallow water. However, the extreme winds produce abnormally large waves which in turn create excessive wave-pumped currents through the reef. A wind strength of 40 m s^{-1} fetch-limited over a distance of 100 km would produce 8 m high waves. Thus exit currents from the reef would be of order 3 to 5 m s^{-1}. Furthermore, the breaking waves damage the reef and produce rubble which erodes the living coral whilst it remains within reef structure.

It seems that such currents may well serve as a controlling influence on the size of breaks in the reef. In Section 4.1 u_{wave} was shown to be inversely proportional to β the fraction of reef occupied by breaks. It would be of value to survey the reef before and after a cyclone to determine whether physical changes have occurred near the breaks.

The other major possible effect on the reef comes from intense rainfall often associated with tropical cyclones. Terrigenous sediment transported into the lagoon via creek runoff may settle on corals causing 'stress'.

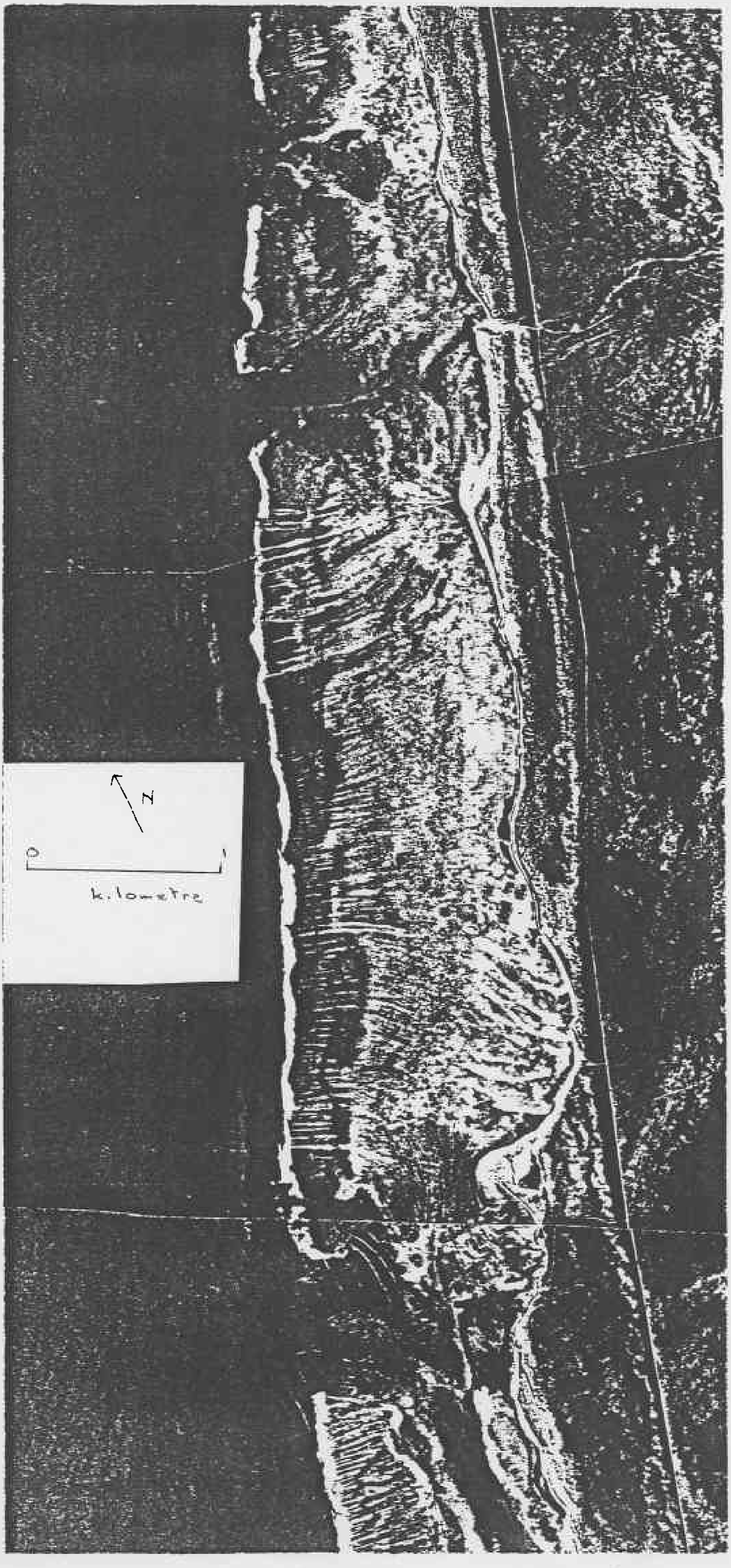


Figure 4.10

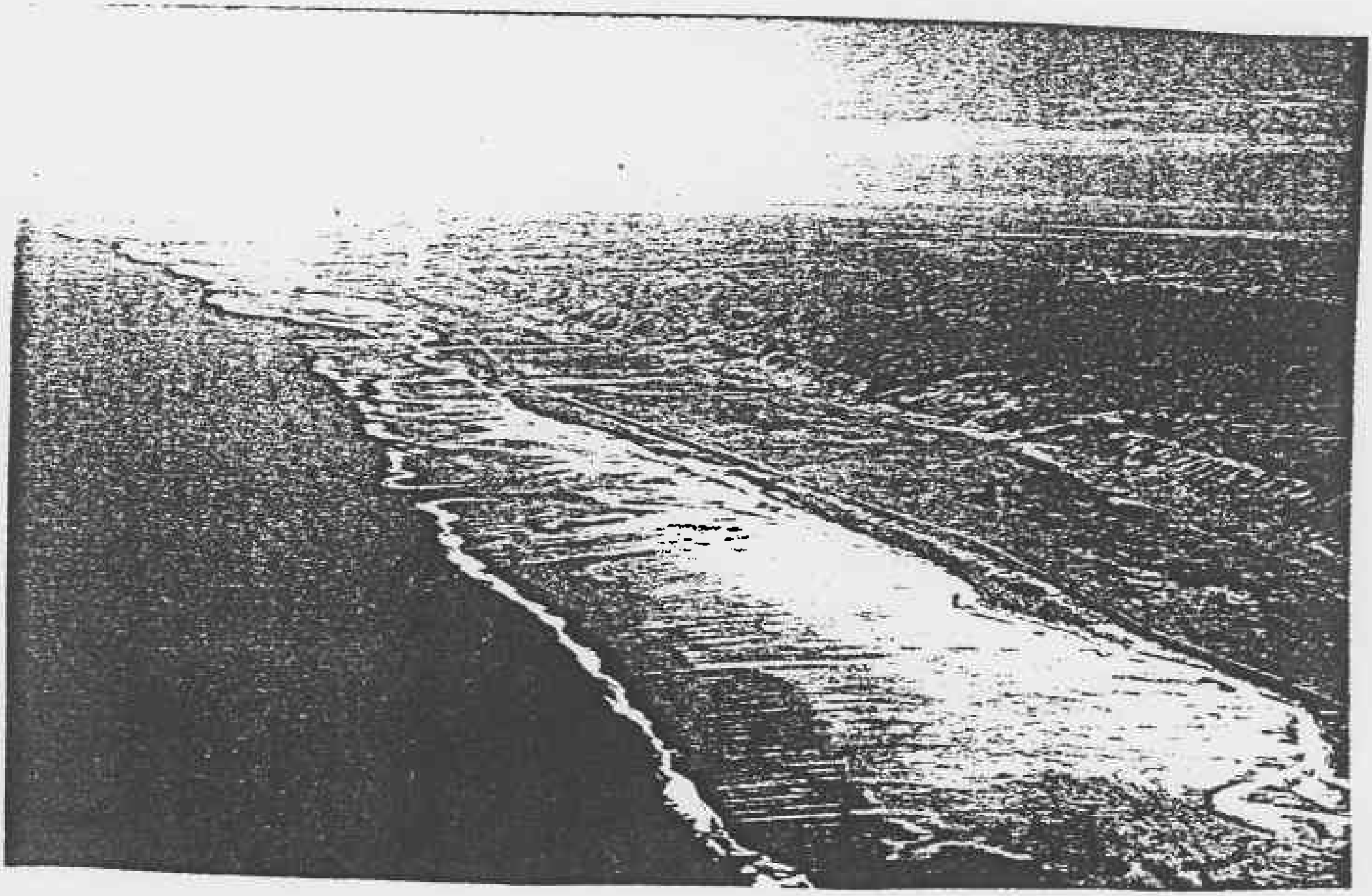


Figure 4.11

CHAPTER 5 CONCLUSIONS AND RECOMMENDATIONS

5.1 Overview of the Oceanography of Ningaloo Reef

The Ningaloo Reef tract is located near the shelf-break on the western coast of Australia. From the physical viewpoint it is unique because the coastline lies only 2 to 6 km behind the reef. This creates a narrow and very shallow (mean depth ~ 1 to 2 m) coastal lagoon behind a major barrier reef, located on the narrowest portion of continental shelf in Australia.

Wave-pumped currents have a major influence on the dynamics of the lagoon. The current trajectories cross the straight reef-line normally and are only deflected into an alongshore direction at distances of a few hundred metres from the shore. Over 50% of this alongshore flow occurs through deep channels. These basically run alongshore and then swing offshore to exit the lagoon through breaks in the reef.

The Ningaloo Reef tract appears to divide naturally into three sectors having different topographies of the reef and lagoon. Only over the northern sector (North West Cape to Point Cloates) is the reef strictly on the shelf-break and this sector is of greatest physical interest.

Over the northern sector some 15% of the length of the reef-line is occupied by breaks. It is estimated here that at high spring-tide the exit current through the breaks generated by breaking waves is 1 to 2 m s^{-1} under moderate to high swell conditions.

The average flushing time of the lagoon due to breaking waves is estimated to be less than a day; the lagoon is therefore well-flushed and this is a consequence of its very small volume per unit length of reef (i.e. cross-sectional area).

The lagoon will be well-mixed vertically at most times. This is a consequence of its limited depth and strong current flow; vertical mixing times are estimated to be less than 1 hour.

Lines are visible in the aerial photography of the Ningaloo Reef flat which conform with the expected current trajectories of water pumped into the lagoon by breaking waves. The lines enter the deep channels as expected and may serve as a unique long-time average of surf-induced water flow in the lagoon.

The outer reef slope and reef crest appears to contain a groove and spur structure with wavelength of order 10 m. This is absent in the occasional segments of north-facing reef which do not lie within the dominant directions of incoming waves.

Tides over the reef are mixed diurnal/semi-diurnal (form factor is about 0.8); North West Cape lies in the transition region between the mixed, but predominantly diurnal tides, of the west coast and the semi-diurnal tides of the North West Shelf. Spring tides are basically semi-diurnal with high water, controlled by the S2 phase, at about midday and midnight. Coincidence of the diurnal and semi-diurnal spring-neap cycles occurs at Point Murat near the end of March.

Whilst the reef-crest is dry (at low water) tidal currents enter through the reef breaks. They then tend to follow through the deep channels; the fraction of volume flux carried by the channels is similar to that for the flux induced by breaking waves. At spring tide maximum current speed through the breaks (in the northern sector) is about 0.6 m s^{-1}.

Tidal flow can occur over the reef when the reef-crest is covered. The form of the variation of flow pattern with water level is controlled by the Reynolds number which is presently unknown.

Tidal flushing times are estimated to be about 4 or 5 days; considerably longer than those due to breaking waves.

The mean wind speed and direction at Carnarvon (based on wind stress) is 5.51 m s^{-1} from 184°. This is probably the most reliable estimate of winds over the southern sector (south of Point Maud) of the Ningaloo Reef tract. Carnarvon shows the daily sea-breeze cycle typical of the west coast of Australia with a strong south-westerly wind during late morning and afternoon.

There is a definite change in wind pattern between the west and northwest coasts. The weather station at Learmonth, on the eastern side of the Exmouth Peninsula, and west coast of Exmouth Gulf shows a mixture of south-westerly and north-easterly sea breezes.

The Cape Range which extends down the middle of Exmouth Peninsula shelters the northern sector of the Ningaloo Reef tract from much of the easterlies. However the shelf in this region is likely to experience alongshore variability of wind direction on a scale of 100 km. Considering the slope of the shelf it appears that alongshore winds are dominant in determining the wind-driven component of water circulation on the shelf.

The southerly wind over the reef creates a structured topographic gyre with northward movement in the lagoon and southward return currents outside the reef; typical lagoonal currents are 0.1 m s^{-1}. Lagoonal flushing-times due to the wind are slightly in excess of one day and faster than tidal flushing.

The Leeuwin Current is a southward movement of tropical warm, low-salinity, water which occurs in autumn or winter along the North West Shelf and west coast of Australia. There is no information available on its structure near North West Cape. It would be anticipated that the current has a significant effect on the Ningaloo Reef tract because the reef is located on the shelf-break where the current is maximal; the current is similarly evident at the Abrolhos Islands.

There are two possible effects of the current. Firtly it may induce a southward water movement in the lagoon. The pressure gradient driving the southward flow is equivalent to a water slope of 3×10^{-7}. In 1 m of water this would create a velocity of 0.01 m s^{-1} if it were balanced by bottom friction with a 'background water speed' of 0.1 m s^{-1}. Thus it is safe to assume that the lagoon is too shallow for the head driving of the Leeuwin Current to generate any appreciable lagoonal flow. However the current should be evident offshore of the reef.

Secondly, the Leeuwin Current may advect warm, lower-salinity, water onto the shelf-break. Because the lagoon is tightly coupled to the open ocean through the fast flushing processes, it will be occupied by this Leeuwin Current water. Existing temperature data for the lagoon confirm the high degree of correlation between lagoonal and ocean mean (weekly) temperatures.

Lagoonal waters show a diurnal heating/cooling cycle of about 2°C. This is typical of water of this depth in estuarine environments on the west coast; in coastal waters it normally only occurs within the surface mixed layer.

The limited amount of temperature data from the Ningaloo Reef lagoon show several temperature anomalies. These consist of a drop in mean water temperature of 1 to 2°C below the seasonal average for 2 or 3 days. It is clear that this is due to cooler water arriving on the shelf and being advected into the lagoon. The times for onset and decay of the anomalies within the lagoon are related to the flushing time of the lagoon. It is evident that the arrival of the cooler water on the shelf is a rapid event occurring within about 12 hours. Similar temperature anomalies have been recorded at the Abrolhos Islands, and the longer onset times there may reflect the greater flushing times of the deeper lagoon.

The most likely cause of the temperature anomalies is upwelling onto the shelf caused by breaking internal waves. Such waves are known to produce vertical displacements of 30 m in water of depth less than 100 m at the North Rankin moorings on the North West Shelf; Ningaloo Reef tract and the Abrolhos Islands are situated a similar distance from the shelf break. The internal waves are semi-diurnal but are not coherent with the barotropic tide and show no spring-neap cycle.

The nearest location to the Ningaloo Reef tract where detailed open ocean studies have been made is in Exmouth Plateau which is a relatively shallow region 250 km offshore from North West Cape. Tidal currents are mainly semi-diurnal. There are also strong inertial period (31.9 hours at Ningaloo) currents and high energy bursts of short-period internal waves.

5.2 Baseline Studies

A most important foundation for any physical, or ecological, marine research program is a set of baseline physical data. This is also vital to the preparation of any environmental resources management plan. Proper protection and management of the coastal environment cannot be realistically attempted without adequate data. There is a minimum data set which physical oceanographers need in order to be able to advise on matters relating to marine biology and coastal engineering.

A basic list follows:

- Tides A tide gauge should be deployed for one year and subject to detailed harmonic analysis.
- Currents A knowledge of mean current speeds over an extended period is important. It is suggested that an electromagnetic or acoustic current meter be deployed in one of the deep channels in the lagoon. This should be subjected to statistical analysis.
- Wind Installation of a wind anemometer near the Ningaloo Reef is vital; a well positioned (easily accessable) land station is adequate. Long-term data should be analysed.
- Waves A wave-rider buoy should be deployed off the reef and subject to standard analysis.
- Bathymetry The bathymetry of the lagoon should be established over one hundred transects normal to the shore.

Satellite data provide mean weekly temperatures for the lagoon but should be properly analysed on a weekly, monthly, seasonal, and annual basis.

5.3 Circulation Studies

The basic processes of water movement within the lagoon have been discussed in this report. This serves as a starting point for more detailed studies.

The first factor to emphasise is the extreme complexity of currents in the lagoon. This is a consequence of the three driving forces: breaking waves, tidal variation, wind stress; which have comparable magnitudes. Furthermore the wave-driven and wind-currents are strongly coupled in magnitude to the tidal elevation.

It is therefore essential that a two-dimensional hydrodynamic numerical model of water flow be commissioned. This should contain the three driving forces and be fully non-linear and include drying of reefs. The model would be used to define periods when field data should be collected. These data enable the model to be verified and calibrated. The model predictions of mean long-term flow could be compared with the lines of the aerial photographs (Section 4.1) which have been attributed to trajectories of wave-driven currents.

The model would be constructed for a specific area of the lagoon which would become a major study area for field work. Considering the topography of the lagoon a finite difference grid consisting of cells 200 x 200 m would seem adequate. Given a maximum depth of 5 m an explicit system would require a time-step of 20 s. For practical purposes this restricts the model to about 1000 cells; if the lagoon is 2 km wide, the model can extend a maximum of 20 km alongshore.

Given the choice of a study area, field work would be necessary to provide detailed maps of bathymetry and reef topography. Following the operation of the model, measurements would be made of water movements. The simplest techniques are dye release over the reef crest and drogue tracking in the deep channels. There is considerable drogue-tracking expertise in the Centre for Water Research and the Ningaloo Reef tract could become part of the field programme conducted by the Australian Survey Office.

In Section 4.2 a set of temperature contours was used to imply the structure of tidal currents within the Ningaloo Reef lagoon.

It is to be stressed that Figure 4.5 is only a first attempt at current mapping through a temperature tracer. However the data set does show the potential of this technique for the Ningaloo Reef lagoon. The method is applicable there for two reasons. Firstly the lagoon is narrow so that it is possible to sample its entire width between reef-breaks in a matter of an hour or so. This allows a synoptic view of temperatures in the lagoon to be obtained over an elapsed time which is small compared to the tidal period. In a larger lagoon this would not be practical using a single boat. Secondly the lagoon is very small, i.e. of depth comparable to the tidal range. This means that tidally advected water entering the lagoon maintains its temperture differential and this is not greatly reduced by extensive mixing with existing lagoonal water.

It is perhaps worth drawing an analogy between this temperature survey from a small boat at Ningaloo and Landsat images of the Great Barrier Reef area; Wolanski et al. (1984). Both use a tracer (temperature in the one case and turbidity in the other) to produce information on currents. The Ningaloo Reef lagoon is so much easier to survey simply because one is concerned with phenomena on scales of a few hundred metres compared with 10's of kilometres on the Great Barrier Reef. The cost of obtaining Landsat data and processing it is very large compared with a simple temperature survey which can be contoured using a standard software package, e.g. the GPGS routines on the DEC-10 at the Western Australian Regional Computing Centre.

The temperature mapping method is capable of giving quantitative information on currents in the lagoon; nothing of this detail has been attempted here, and Figure 4.5 gives only a qualitative impression of currents. The method to be employed is the 'inverse technique' and this attempts to find an advective velocity field, and dispersion matrix, which will allow the advective-diffusion equation to have a solution corresponding to the temperature map.

Data collection from a boat is a perfectly adequate method of mapping currents on spatial scales of hundreds of metres and temporal scales of hours. For higher resolution an airborne infra-red radiometer can be used. An example of this type of survey on the western coast of Australia can be found in Hearn and Pearce (1985). This employed the F27 aircraft which was then operated by CSIRO; it is now a national facility. The infra-red linescan equipment on the aircraft is of a very advanced design giving a spatial resolution of order 10 m (at the flying altitude necessary to scan the whole lagoon and an equal strip of open ocean beyond the reef). The temporal resolution is simply a matter of the flight-time over the length of lagoon being surveyed. Given the high speeds of water movement, i.e. 0.1 to 1 m s^{-1}, instabilities over scales of hundreds of metres may have characteristic times of minutes. Hence an aerial survey would clearly be conducted over a distance of some 10 km (as proposed for the length of study region below) with repeated over flights every few minutes for some hours.

The value of an aerial survey depends on the existence of an adequate temperature signal. Airborne radiometers have a temperature resolution which is typically 0.1°C and so an overall signal of about 1°C is required. The cost of mounting such a survey is very high and a considerable amount of work is involved in data processing and analysis. Specialised image-handling hardware and software is required and this does exist with the Western Australian Department of Lands and Surveys. The cost of a survey and analysis is of order $30,000. It would therefore only be undertaken after considerable knowledge of the system had been acquired through other types of data collection and its analysis.

The NOAA satellite provides a facility for collection of sea surface temperatures via its advanced very high resolution radiometer (AVHRR). This has the much lower spatial resolution of about 1 km, and (although data collection costs are not borne by the investigator) it is not therefore an attractive tool for studies within the Ningaloo Reef lagoon.

5.4 Study Region

Both the numerical model and the requirements of a field program require the definition of a study region. As discussed in Section 5.3, the model would have a maximum area of about 40 km^2. The need to specify boundary conditions means that not all of this area could be used for the actual study region. Experience gained by Hearn and Hunter (1985) in modelling the Marmion Reef (which also has a continuous coastal lagoon) suggests that although the study region may in practice occupy most of the model area it is wise to use only 50% of its area. Thus the study region should have an area of about 20 km^2.

A special feature of the Ningaloo Reef tract is the existence of a shallow, narrow, long coastal lagoon close to the shelf-break in the northern sector. For this reason the study region should be situated in this sector. An ideal region would have a long length of unbroken reef with a major break towards its southern boundary. The lagoon should be as narrow as possible with a well-defined deep nearshore channel.

A suitable study region is located between Yardie Creek and Pilgramunna, centred on Sandy Bay; the northern part of this region is shown in Figures 4.10 and 4.11.

The study region will serve as the focal point of most scientific investigations of the Ningaloo Reef tract. By coordinating the work within one such area (for which a high quality predictive model of water circulation exists) it will be possible to develop a comprehensive understanding of the dominant ecological processes. The inter-relationships identified in the study region will be generally applicable to the whole reef tract and provide a sound scientific basis for management of the Marine Park.

5.5 Open Ocean Studies

It is worthwhile to investigate the effect of shelf dynamics on the lagoon. Particular events, such as breaking internal waves, may be the source of abnormal phenomena like the temperature anomalies discussed in

Section 4.5. Such processes may be important to the external supply of nutrients to the reef.

This type of study is outside the oceanographic resources of West Australian institutions. However, a large study of the tropical Eastern Indian Ocean in 1986/87 will be conducted by RV 'Franklin' the new Australian national oceanographic facility. Data will be collected from North West Cape to Christmas Island as a part of the LUCIE program (Leeuwin Current Interdisciplinary Experiment). This provides a good case for immediately installing a thermistor station in the study region to collect hourly temperature data over a complete year. RV 'Franklin' will also be collecting other data on the shelf adjacent to Ningaloo Reef and south to Shark Bay during the first half of 1987. It must however be stressed that scientific effort would have to be funded in order to analyse this data.

Consideration should be given to the long-term deployment of a current meter outside the barrier reef for monitoring annual variations in strength, duration and timing of the Leeuwin Current, after the cessation of the LUCIE programme. This would provide useful long-term data for studying dispersal and recruitment patterns of the planktonic larval stages of many coral reef organisms.

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APPENDIX I

Notes on the preparation of a map showing general bathymetry and surface seawater isotherms in the vicinity of Coral Bay in the Ningaloo Reef lagoon, on March 31, 1986.

BATHYMETRY

Aerial photographs at a scale of 1:20,000 [WA 2290(c) 12-03-85 Ningaloo Marine Park (Coast run, Cape Farquar-Norwegian Bay)] were traced, to provide a map of the coastline and position features such as the outer barrier reef and its channels, sand banks, areas of coral cover and bare sand. The areas covered by coral were divided into 'high intertidal reef' and 'low intertidal/subtidal reef' on the basis of tonal differences, comparisons with 1:25,000 aerial photographs (WA 2162(c) 15-07-83 Coastline: Coral Bay - Carnarvon) and 'ground truthing'. It should be noted that some portions of the outer reef shown in Figure 4.5 were not in the aerial photographs, and their position should only be taken as an approximation.

TRAVERSES

Six temperature traverses were performed with a 5 m runabout driven at constant speed (approximately 5 knots), employing a thermistor (± 0.02°C) attached to the stern of the boat, 20 cm below the water surface. Instantaneous measurements of seawater temperature were recorded on a data logger (Windrift Instruments, W.A.) at 30 second intervals. All traverses were performed at a bearing of 80° to conspicuous landmarks on the coast. Traverses were performed during a spring flood tide.

DATA ADJUSTMENT

Seawater temperatures in the Ningaloo Reef lagoon are known to exhibit a pronounced diurnal variation of about 2°C, consistant with heating during the day by insolation, and cooling at night (Simpson and Masini 1986). Approximately 90 minutes elapsed between the end of traverse #2 and the start of traverse #3; all other traverses were performed in succession with an elapsed time of less than 10 minutes between them. Simultaneous measurements of seawater temperature at a site within the traversed area (site 1, Figure 1.3) using a data logger (Windrift Instruments, W.A.) mounted in a submersible housing showed an increase in bottom water temperature of 0.52°C during the 90 minutes that elapsed between the second and the last four traverses. As this site is relatively well-mixed vertically, measured temperatures on transects 1 and 2 were adjusted by +0.52°C to construct the surface isotherms.

NOTES

Deepwater (> 4 m) channels are commonly found inshore from breaks or discontinuities in the outer reef and in some areas of the lagoon closer to shore, adjacent to eroding landforms (such as large sandhills, cliffs, etc) or off the end of sand spits. These channels are predominantly found in narrow sections of the lagoon. One such channel and break in the reef is located on the southern side of the traversed area. From the isotherms it appears that a body of warm water (> 26.2°C) intruded from the vicinity of the break in the outer reef and progressed down channels parallel to the shore. The relatively cooler water around the group of intertidal 'bommies' west of the Coral Bay settlement supports the inferrence that 'warmer' oceanic water is intruding into the lagoon where flow is relatively unrestricted, that is via the channels, and takes longer to mix with the cooler lagoonal water where flow is restricted due to the baffling effect of the coral growth form in that area.

Although the lagoonal water was cooler than the oceanic water at the time that the traverses were performed, arguably, a similar plot of temperatures under the same tidal regime and conditions, but later in the afternoon when lagoonal waters commonly reach their maximum, would appear as an intrusion of 'cooler' oceanic water into the warm lagoon, via these same channels.

The maximum temperature of 28.0°C recorded in Coral Bay was only 1 record and was probably the result of water flowing off the extensive shallow sandflat on the leeward side of sandhills and as such can be considered a very localised phenomenon.