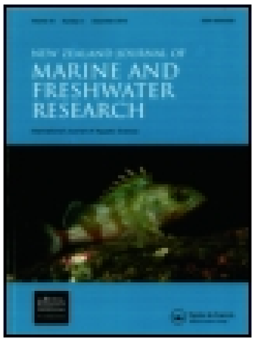

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Dynamics of the Cape Farewell upwelling plume, New Zealand

T. G. L. SHIRTCLIFFE
M. I. MOORE*
Research School of Earth Sciences
Victoria University of Wellington
P.O. Box 600, Wellington, New Zealand

*Present address: New Zealand Oceanographic Institute, DSIR Marine and Freshwater, Department of Scientific and Industrial Research, Private Bag, Kilbirnie, Wellington, New Zealand

A. G. COLE
New Zealand Oceanographic Institute

A. B. VINER
Taupo Research Laboratory
DSIR Marine and Freshwater
Department of Scientific and Industrial Research
P.O. Box 415, Taupo, New Zealand

R. BALDWIN
B. CHAPMAN
New Zealand Oceanographic Institute

Abstract Hydrological observations made in January 1984 in the region near Cape Farewell, New Zealand, are described and previously published observations reviewed. It is shown that upwelling depends on the existence of the intermittent Westland Current, and is intensified by an onshore wind. Such a wind induces a fall in sea level near Cape Farewell, and the resulting favourable sea surface slope accelerates deep water over the bathymetric rise inshore of Kahurangi Shoals. The hydraulic response of the thermocline, coupled with a coastal convergence of the bottom Ekman flow, produce a strong upwelling source near Kahurangi Point.

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Keywords Cape Farewell; Farewell Spit; Kahurangi; upwelling;

INTRODUCTION

Cape Farewell forms the north-west corner of the South Island of New Zealand (Fig. 1). To the north lies a large bay which extends c. 110 km north-east to Cape Egmont and c. 185 km south-east to the Wellington coast of the North Island. Nomenclature is variable, but in this paper the bay thus defined will be referred to as Taranaki Bight. At its south-east corner, Cook Strait separates the North from the South Island and leads to the east coast waters. The region is windy, with frequent strong winds from the west, and there are strong tidal currents through Cook Strait.

The Cape Farewell area is known to include the source of an intermittent plume of water which upwells from a depth of about 100 m; it has been investigated several times. Cruise 1155 of RV *Tangaroa* visited the area in January 1984 with a primary goal of investigating small-scale features and mixing associated with this plume. The present paper reviews previous investigations, describes the hydrological observations which were made, and presents a theoretical description which accommodates the principal features of the observations. The small-scale processes are considered elsewhere. Chief interest in the present analysis lies in clarifying the role of onshore winds in accelerating the longshore flow and intensifying the upwelling inshore of Kahurangi Shoals.

Review of previous work

The first published observation that an extensive region of cold water near Cape Farewell resulted from the upwelling of subsurface water was that of Garner (1954). The data published in that paper were surface temperature measurements made from commercial ships. Garner (1959) published a temperature section measured from HMNZS *Lachlan* in October 1951 which showed a band of cold water

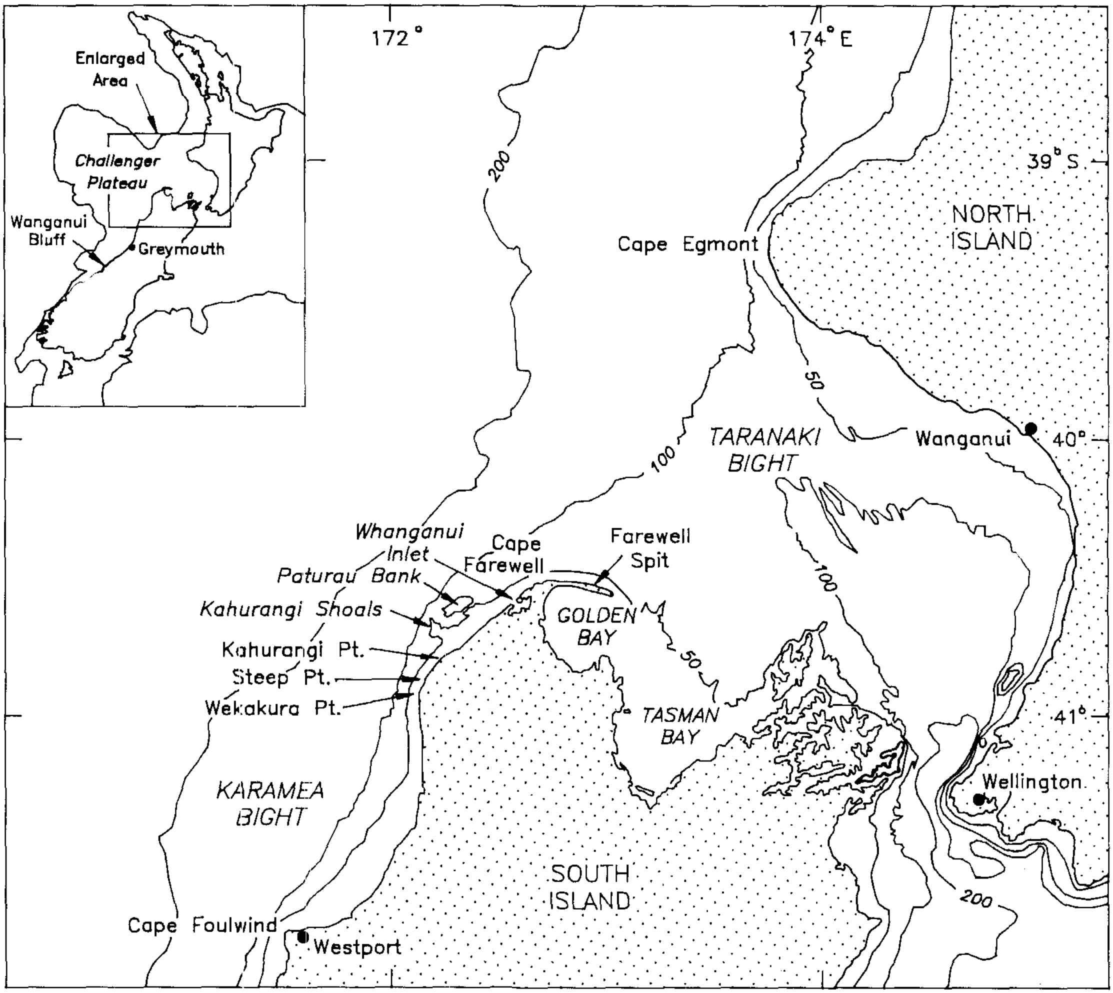

Fig. 1 Location of the observations.

north of Cape Farewell. However, in the same paper, a contrasting near-coast survey by *Lachlan* of sea surface temperature round the South Island in February/March 1952 was presented, and this showed a tongue of warm water stretching southwards along the entire West Coast, with no sign of upwelling.

The first detailed study of the area was conducted in 1969 by Stanton (1971). Upwelling was present, more marked north of Kahurangi Point than south of it. Stanton noted a weak subsurface salinity maximum just outside the shelf edge, and ascribed it to a southward-going compensation current associated with the upwelling. He also commented on the presence of slicks and banding parallel to the coast, associated with internal waves in the highly stratified water. He believed that upwelling was general along the coast, but locally intensified at certain points.

Stanton (1976) surveyed the area again in 1974, in the course of a study of regional geostrophic flow. Conditions had been calm, and the near-coastal flow was southward at all depths, with no upwelling. A tongue of warm saline water projected south-west into the study area north of Cape Farewell. Stanton suggested that the Westland Current, a northwards-moving mean flow identified by Brodie (1960), was a response to the south-westerly winds prevalent along the West Coast. Apart from the absence of this current, Stanton's geostrophic flow agreed with the earlier drogue and drift-card observations. The regional inflow from the west turned south about

42°S, following the bathymetry along the flank of the Challenger Plateau, and an anticyclonic eddy appeared to occupy the shallowest part of the Plateau.

By 1979 it had become clear that the northern West Coast of the South Island was biologically important. Indeed, Bradford & Roberts (1978) had shown that the Karamea Bight – Taranaki Bight – Cook Strait area was the most extensive of only four New Zealand coastal regions in which zooplankton biomass exceeded 300 mg m^{-3}. A cruise of RV *Tangaroa* in June 1979 led to several papers (Bradford 1983; Chang 1983; MacKenzie & Gillespie 1984; Chang & Bradford 1985). Three transects were studied, off Cape Foulwind, Greymouth (42°30'S, 171°10'E) and Wanganui Bluff(43°02'S, 170°25'E). Some upwelling was evident from a depth of 75 m to the surface at the inshore station off Wanganui Bluff, with downwelling below 150 m depth; the other transects showed neither upwelling nor downwelling. The winds were strong from the south-west before the Greymouth section, strong from the north-east during that section, moderate from the westerly quarter during the Westport section, and increased to strong south-westerlies during the Wanganui Bluff section. Taken with the suggestion of Stanton (1976) concerning the importance of the south-westerly wind to the existence of the Westland Current, these results suggest that the tendency to upwelling in this region might also represent a, rather rapid, response to such winds. This would be consistent with the classical picture of upwelling resulting from an offshore surface Ekman transport, the Westland Current being set up to balance the surface stress by bottom friction.

A detailed summer survey was carried out from RV *Tangaroa* in January/February 1980 (Bowman et al. 1983a, 1983b; Bradford et al. 1986). Upwelling was observed at Cape Farewell, and ascribed to bottom friction acting on the Westland Current, reinforced by southerly wind gusts lasting a few days, plus the centrifugal effects of flow past convex coastal bends. Winds were light southerlies before the survey, but steady north-westerlies during it. Classically, such winds should not give rise to upwelling on this coast, but near Kahurangi Shoals water at a temperature of 13.7°C reached the surface from a depth of 90 m. During this survey, flow patterns in Taranaki Bight and Cook Strait were investigated using drifting buoys.

A further study was undertaken in January 1981 from HMNZS *Tui* (Bowman et al. 1983a, 1983b, 1983c; Bradford et al. 1986). With winds moderate from the north-west, upwelling was again present, water reaching the surface near Kahurangi Shoals from a depth of about 50 m. Both in this survey and the previous one, numerous isolated patches of cold water were found in Taranaki Bight, and Bowman et al. (1983c) discussed the possible origin of these "eddies" in terms of a barotropic numerical model. They concluded that upwelling at Kahurangi Shoals was caused mainly by bottom Ekman pumping, that tidal mixing probably contributed, and that centrifugal effects were insignificant. They estimated that the typical wavelength and phase speed of meanders in the plume, 28 km and 0.4 m s^{-1}, respectively, indicated a period close to the inertial period, suggesting that the eddies were formed by wind-induced fluctuations of the intensity of the Westland Current. Pointing out that the spin-down time of such eddies should be less than 1 day, they concluded that their lifetime of 10–15 days showed that rotation was being maintained by the conversion of potential to kinetic energy; but it should be noted that in spite of a maximum tangential velocity of 0.8 m s^{-1} being estimated from the isopycnal slopes in one eddy, there has been no demonstration of closed trajectories in these flows.

The picture that is presented in these two studies is of a persistent Westland Current leading to persistent upwelling at Kahurangi Shoals, with the production of cold-core eddies resulting from modulation of the process by wind. This picture is inconsistent with earlier studies which showed that the Westland Current and the upwelling are both intermittent phenomena, but it seems quite possible that the time scale of this intermittency is rather long.

Further information was obtained during a succeeding cruise of HMNZS *Tui* in February 1981 (Foster & Battaerd 1985). The survey began in light winds. A plume of upwelled water stretched into Taranaki Bight, but the lowest surface temperature of 14°C was found just south of Kahurangi Point. The following day this had gone, and the coldest water was 15°C over Paturau Bank. On this day, a tongue of warm (18°C) water was observed inshore on a NOAA7 satellite image, apparently moving south from Cape Farewell. On the third day of the survey the surface temperatures rose again to a rather uniform 15–16°C in the region between Kahurangi Point and Cape Farewell, but on the following day, the wind having risen to 9 m s^{-1} from the west, the surface temperature had again fallen to 14°C in a plume originating at Kahurangi Shoals.

In 1982, three cruises (in January, February/March, and April/May) were made to the triangular region bounded by the Challenger Plateau, Kahurangi Point, and Wanganui Bluff (Heath & Ridgway 1985). Off Kahurangi Point, below the thermocline, isopleths

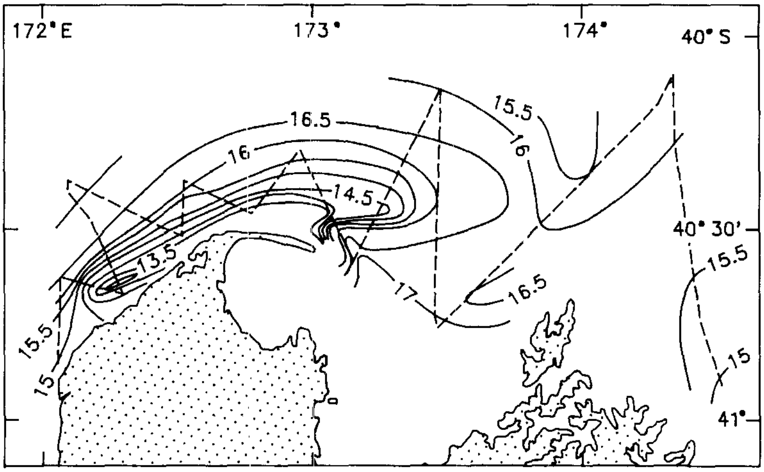

Fig. 2 Contours of sea surface temperature (°C) on 21 January 1984.

sloped upwards towards the coast in January, downwards in February/March, and were horizontal in May. There was no upwelling further south in any of the surveys, and even in January upwelled water near Kahurangi Point failed to reach the surface. The heat content of the water column actually increased between January and May, associated with a deepening of the mixed layer. Measurements of the horizontal turbulent heat flux at a current meter moored on the southern flank of the Challenger Plateau, about 100 km offshore, showed an oscillating onshore flux which fluctuated with time scales as long as 4 weeks, and it was suggested that this might induce changes of the heat content of the shelf water with similar time scales. Further discussion of the current meter measurements was given by Heath & Gilmour (1987b), and it seems clear that the exchange between oceanic and shelf waters provides a long-period excitation which is available to modulate, and perhaps to control, the regional tendency to upwell.

A programme of measurements was undertaken from RV *Tangaroa* in March/April 1983, and physical results have been described by Heath & Gilmour (1987a). Two features of this work are particularly significant. Firstly, a series of detailed surveys of sea surface temperature were carried out, and these show that an upwelling event originated inshore of Kahurangi Shoals about 15 March, and apparently ended by about 20 March. This event, lasting 4–5 days, coincided with a burst of the component of wind directed towards 128°T at Farewell Spit. As it died, a tongue of warm water, the source of which was identified as Tasman Bay, flowed south inshore around Cape Farewell.

Secondly, a current meter was moored at mid-depth in 105 m of water just offshore from Kahurangi Shoals. This revealed fluctuations particularly in the longshore (038°T) flow with maximum variance at periods around 10 days. This variability was strongly correlated with the component of wind towards 128°T as measured at Farewell Spit, and a simple numerical relationship was established which seemed quite successful in predicting the non-tidal longshore flow from a knowledge of this wind component.

Although this correlation between wind and current was a strong one, the discussion of other possible causal relationships based on the concept of integral time scales was unconvincing, given the oscillatory nature of the autocorrelation functions involved. Nevertheless a calculation of the cross-correlation function between the longshore flow and a tidal forcing function showed the latter to be much less important than the 128°T wind.

In all, 13 relevant oceanographic cruises to the region between Jacksons Bay in the south-west and Cook Strait in the north-east had been undertaken before 1984 and reported in the oceanographic literature. Upwelling was observed on only about half the cruises, but when it was observed it was not seen to terminate within the period of the cruise, although its intensity was seen to fluctuate.

OBSERVATIONS IN JANUARY 1984

The present observations are summarised in Fig. 2–7. They were completed between 21 and 28 January 1984, a period coinciding with the transition from spring to neap tides. Fig. 2 shows contours of sea surface temperature as observed on the first day,

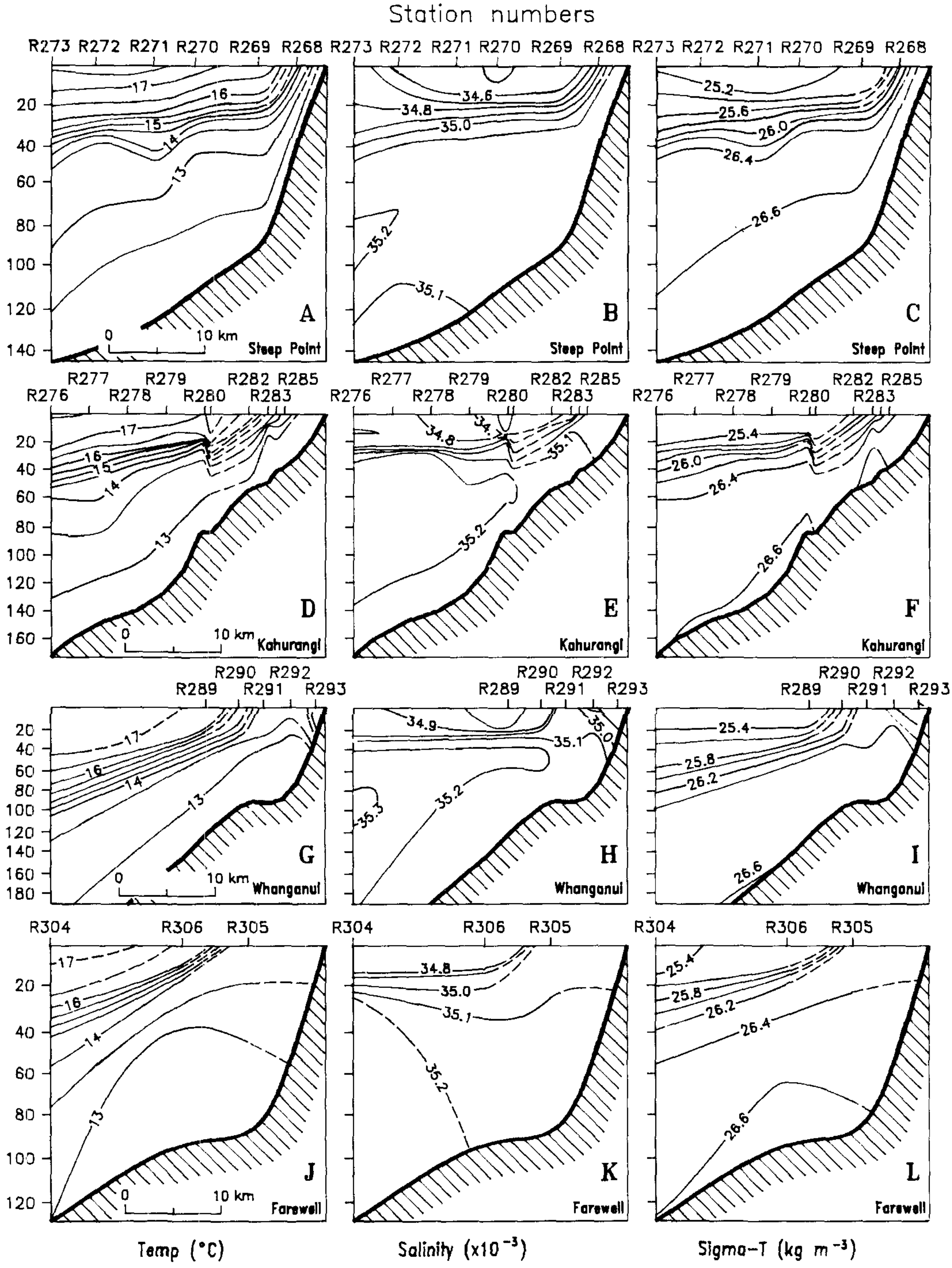

Fig. 3 Temperatures, salinities, and densities observed at Steep Point (A,B,C respectively), Kahurangi (D,E,F), Whanganui (G,H,I), and Cape Farewell (J,K,L). Dashed lines at R281, R282 indicate stations slightly north of Kahurangi line.

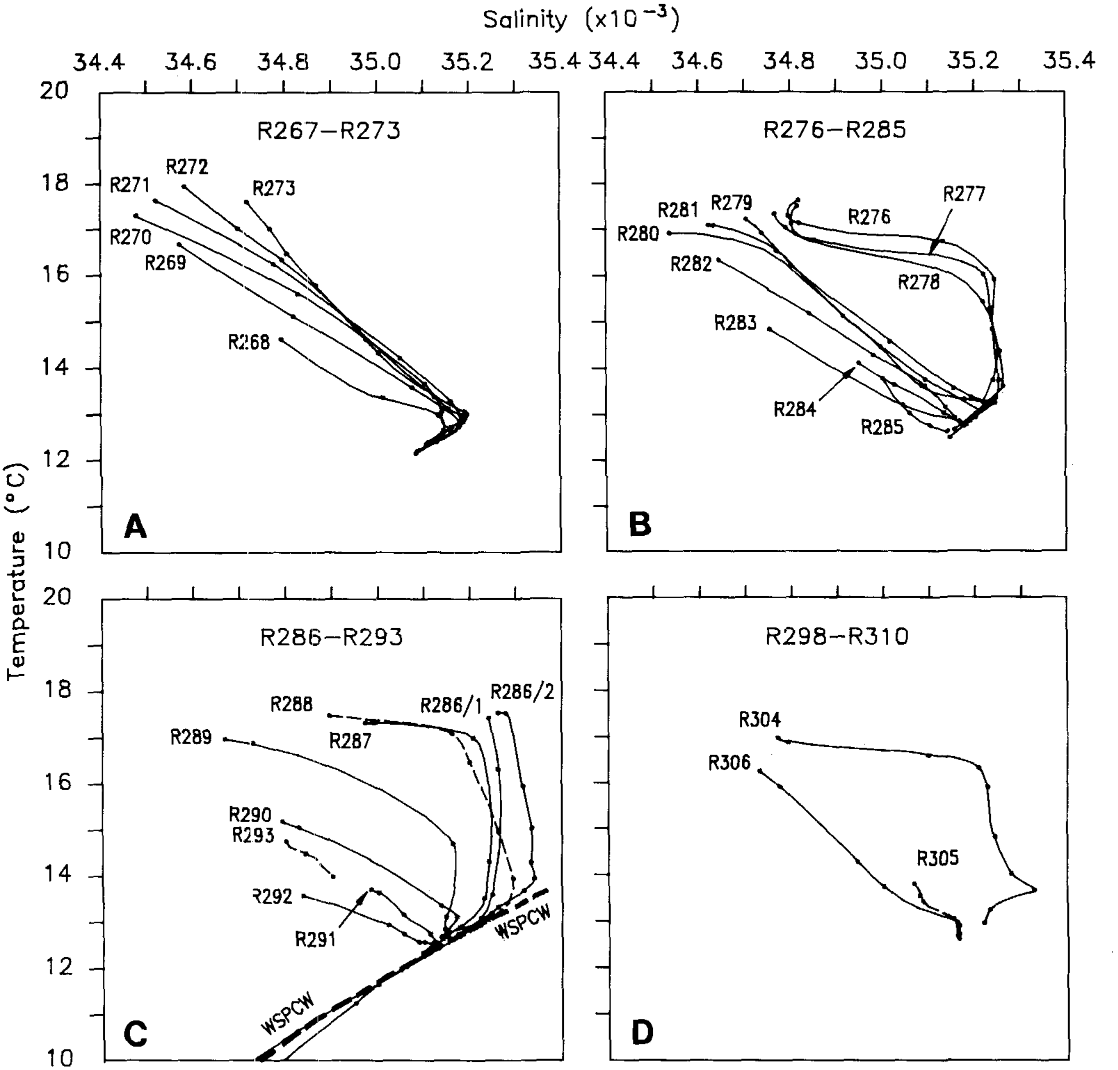

Fig. 4 T–S plots for (A) Steep Point, (B) Kahurangi, (C) Whanganui, and (D) Cape Farewell sections.

outlining the plume from the coldest part of its source inshore of Kahurangi Shoals to its dissipation after turning east into Taranaki Bight. Subsequent observations were generally consistent with these, but with differences in detail.

Between 22 and 27 January, measurements included four lines of stations at which reversing bottles were deployed. These are displayed as temperature, salinity, and density sections in Fig. 3 from south to north identified respectively as Steep Point, Kahurangi, Whanganui, and Farewell sections. The upwelling is evident on all sections and the subsurface salinity maximum reported by Stanton (1971) is also present. Consideration of the temperature sections suggests that the coldest surface water (13°C during the first survey, 12.5°C on 26 January) may have moved inshore from depths in excess of 100 m. However, the salinity sections show a more complex pattern, and it is instructive to examine the data in the T–S plane.

Figure 4 shows the relevant T–S plots. In particular, Fig. 4C includes data from Stations R286 and R287, respectively at positions 39°05'S, 170°45'E and 39°48.4'S, 171°37.4'E. These two stations were

92 n.m. and 33 n.m., respectively, beyond Station R288 on the line of the Whanganui section, too far to allow them to be included in Fig. 3 without a change of scale. Station R286 was located in water 790 m deep on the Challenger Plateau. Two bottle casts were made there 7 h apart, and it is clear that the ship had drifted through a region of inhomogeneous surface water in this period. Nevertheless, the two measurements agree very closely at depths below 200 m, and at greater depths conform closely to West South Pacific Central Water (WSPCW) (Sverdrup et al. 1946). Above 200 m, the water increases in temperature but maintains a fairly constant salinity. All the other water types observed lie on the low-salinity side of the curve of Station R286. It appears that WSPCW is the origin of all the water observed, affected by surface heating and fresh-water dilution. It is not known where this dilution occurs, or what role advection might play. However it is evident in the top 40 m as far as 63 n.m. off shore at R287, though it is only slightly greater 30 n.m. off shore at R288. Closer to shore the degree of dilution increases. Thus the surface water at R292 was presumably derived from deep water at a similar temperature by dilution from a salinity of c. 35.3 to 34.9 $\times 10^{-3}$, requiring the addition of 1% fresh water. In determining the depth of origin of upwelled water, the temperature is a better guide than the salinity.

The T–S diagram for the Kahurangi section, Fig. 4B, shows two distinct types of water. Stations R276, 277, and 278 show much less dilution than those closer to the shore. Although this progression is common to all the sections, there is here a marked change in the short distance between R278 and R279 which distinguishes offshore from coastal water types. The slope of the isopycnals in the density section, Fig. 3F, suggests that water in the subsurface salinity maximum is not flowing north at points between these two stations. However, inshore of that position the flow is northwards and the water type is more characteristic of the coastal waters.

Similar behaviour is shown in the Cape Farewell section; indeed only the Steep Point section fails to show evidence of the offshore water type. It appears that the coastal water extends further offshore there than it does further north.

Some information on non-tidal flow rates in the upwelling area was obtained by tracking surface buoys. One of these, drogued at a depth of 5 m, was released near the inshore end of Kahurangi Shoals, and another, drogued at 10 m, near the offshore end of the Shoals, on 22 January. Over a period of 26.6 h the former moved at an average speed of 0.20 m s^{-1} in a north-eastward direction in water whose surface temperature was 14°C. The latter, in 16.6°C water, moved in a similar, longshore, direction at an average 0.31 m s^{-1} for 25.7 h. This buoy was eventually lost, presumably because its radio beacon failed; it was recovered many months later after beaching on Chatham Island, about 1000 km to the east.

These observations show that the mean longshore flow was 0.11 m s^{-1} greater in the 16.6°C water than in the 14°C water. However, observation of the ship's doppler log showed that this difference was close to 0.5 m s^{-1} 1.5 h after low tide on 26 January; this may be compared to a value of 0.4 m s^{-1} for the velocity difference calculated from Margules' equation for the isopycnal slope in Fig. 3F, so it is probably not primarily tidal. At the time the buoys were tracked, the wind was generally light and variable; when the other measurements were made the wind was predominantly from the west at speeds around 10–15 m s^{-1}. It therefore seems that the longshore flow velocity and offshore shear both increase markedly in response to westerly winds.

The offshore side of the plume was bounded by a front which was marked by slicks lying parallel to the shore, variations of surface texture and colour, and congregations of marine and bird life. There were variations of longshore flow and offshore temperature gradient within the front which indicated the presence of internal waves, and these will be discussed elsewhere.

An attempt was made on 26 January to take the ship southwards along the locus of lowest temperature in the plume. A temperature of 13.1°C was located at 40°36.7'S, 172°15.8'E, between the outer extremity of Kahurangi Shoals and Paturau Bank. Following the minimum temperature southwards, the ship passed inshore of Kahurangi Shoals. The track generally followed an obvious slick, which could be seen to extend nearly to the shore close to Kahurangi Point. It was clear that the source of the coldest water in the plume was along this line.

DISCUSSION

The southern region

The studies reviewed above include some which relate to areas as far south as Jacksons Bay (44°00'S, 168°40'E), even though the plume of upwelled water near Cape Farewell has its origin close to Kahurangi Shoals. Upwelling is a general phenomenon along the whole coast north of Jacksons Bay, and the dynamics of the upwelling centre at Kahurangi Shoals can only be understood in the context of those which apply to the whole region between Jacksons Bay and Kahurangi Point. This southern region is linked to the Cape Farewell plume region by two common factors: the regional wind field and the Westland Current.

Garner (1961) envisaged the Westland Current as originating further south and sweeping north to Cook Strait. Later, Stanton (1976) concluded that it exists only off the northern west coast of the South Island, and that it is driven by the south-westerly winds which prevail along much of that coast, being absent when these winds do not blow, and causing no permanent modification of the mass field.

The response of a coastal sea to a wind pulse has been calculated by Csanady (1982) for a homogeneous (p. 47), a stratified (p. 75), and a two-layer (p. 89) ocean. The non-oscillatory part of the response to a longshore wind consists, close to the shore, of a longshore surface flow, a tilt of isopycnals upwards (or downwards) towards the coast, and little motion below the thermocline; the longshore wind stress is balanced by longshore acceleration of the mixed layer and by the Coriolis force on the surface Ekman flow. When the wind stops blowing, the removal of the surface stress would prevent further longshore acceleration and the flow would become inertial, decaying in a time of order f^{-1}, where f is the Coriolis parameter. Presumably, then, the Westland Current consists of a sequence of such events, with a mean surface flow northwards along the coast from about Jacksons Bay to Taranaki Bight. There, part of the mean flow goes through Cook Strait (the D'Urville Current), while part goes north past Cape Egmont.

The Westland Current is found only north of about Jacksons Bay because south of that point the streamlines which originate far offshore in the eastward Tasman Sea flow come close to the shore in the southward-flowing origins of the Southland Current (Stanton 1976). In turning southward, these streamlines follow the bathymetry which defines the Challenger Plateau. This water is of subtropical origin, typical of WSPCW, and we must presume that the flow is somewhat variable, given the variability of the East Australia Current. It seems possible that the Westland Current might be modulated in turn by such fluctuations, introducing time scales of several weeks into the Westland Current variability. Such time scales have been observed by Heath & Ridgway (1985).

A model of the northern region

Just as the Cape Farewell plume is linked to the southern region dynamics, so it may be affected by the dynamics of the Taranaki Bight – Cook Strait region. Cook Strait is a region of strong tidal flows, since it short-circuits points of nearly opposite tidal phase. It has proved difficult to identify a mean flow over a long period, but non-tidal wind-forced flows in Taranaki Bight are strong (Heath 1978; Kibblewhite 1982).

Bowman et al. (1983a) described the trajectories of six drifting buoys released in the west and centre of Taranaki Bight. Two of these were recovered after 13 and 15 days, respectively, off the eastern approaches to Cook Strait, having travelled through it at average speeds exceeding 0.26 and 0.15 m s^{-1} respectively. The trajectories of three of the other four buoys were also instructive (one was released close to Cape Farewell and came ashore on Farewell Spit). Two which were released 25 and 35 km north of Cape Farewell were recovered to the east, having travelled somewhat to the left of an axial line through Taranaki Bight which points to about 115°T. One of them was recovered in mid-Bight, whereas the other was only 40 km from the eastern extremity of the Bight. Finally, the buoy released west of Cape Egmont was recovered near Wanganui, on the northern extremity of the Bight. The minimum speeds of these buoys fell in the range 0.13–0.27 m s^{-1}. During the period of the buoys' drift, winds were moderate from the north-west. The trajectories therefore demonstrate both a small Ekman flow and stronger downwind flow in the Bight, and a downwind flow through the Strait. Similar trajectories were observed by Heath & Shakespeare (1977) for drift cards released from the Maui A oil platform (39°49'S, 174°07'E).

Taranaki Bight is a shallow basin which, for our purposes, may be regarded as approximately rectangular, its long sides being 185 km long and oriented towards 115/295°T, its short sides being 110 km long. Near its western opening it is c. 100 m deep, but it deepens to 150 m in the south-eastern corner where it joins Cook Strait. Just west of this junction, its vertical cross-sectional area under a line parallel to its short sides is c. 12×10^6 m^2, and its average depth is still close to 100 m. The Rossby radius is c. 300 km, so the surface slope may be assumed independent of position to a sufficient accuracy. Cook Strait is a narrow channel which runs southward from the south-east corner of Taranaki Bight. At its narrowest, it is 23 km wide and at most c. 400 m deep, with a cross-sectional area of 4.8×10^6 m^2 and average depth of 210 m. For our purposes, it may be regarded as having these dimensions for a length of c. 30 km.

We now model the non-tidal flow in this region by assuming that it is driven by the wind stress on the

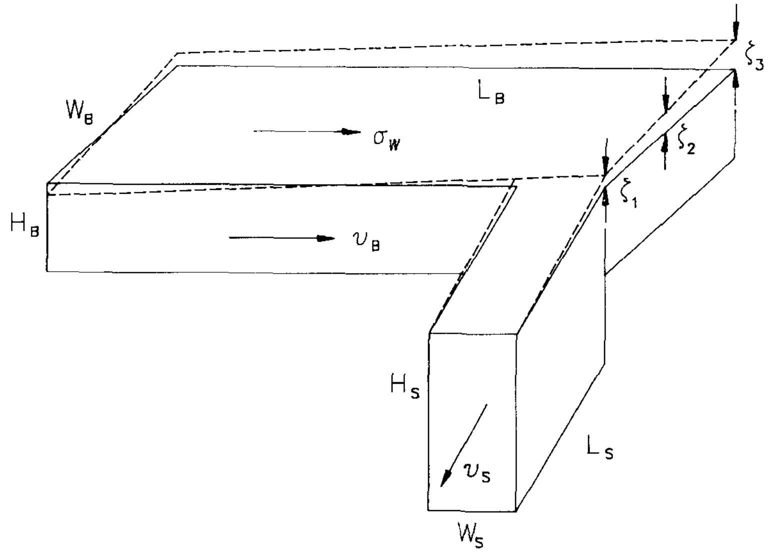

Fig. 5 Model representation of Taranaki Bight and Cook Strait. The tilted sea surface is dashed.

Bight, as was observed qualitatively by Kibblewhite (1982). Consider the effect of a wind blowing along the Bight parallel to its long sides approximately towards the south-east. Initially, an Ekman flow will produce a surface slope, with sea level rising along the northern coast relative to the southern. This in turn will give rise to a downwind slope current in the geostrophic interior of the water column. Steady crosswind total transport must be zero at the northern and southern coasts, so will also be zero between them. Thus once the current has been established, the only forces available to balance the surface wind stress are (a) a pressure gradient associated with a surface slope upwards in the downwind direction, and (b) bottom friction.

In the Bight, we assume that the principal steady force balance is between the wind stress σ_w and the pressure gradient associated with a longitudinal surface slope ζ_2/L_B (Fig. 5). We neglect bottom friction in this region for two reasons: firstly, its inclusion does not alter the conclusions of the model greatly; secondly, Heath (1978) has shown that the water in the Bight is typically stratified, with a reduced flow beneath the thermocline. For the purposes of the model, the flow is assumed uniform and equal to v_B over a depth H_B, and this flow is assumed to be in geostrophic balance with the pressure gradient due to a transverse surface slope $(\zeta_3-\zeta_1)/W_B$. Thus the longitudinal and transverse force balances in the Bight yield the (vertically integrated) equations

$$\sigma_w = \rho g H_B \zeta_2/L_B \qquad (1)$$

and

$$\zeta_2 - \zeta_1 = \frac{-W_B f v_B}{2g} \qquad (2)$$

where g is the acceleration due to gravity.

We neglect the wind stress on the relatively small surface area of the Strait, and assume that the principal steady force balance there is between the pressure gradient due to a longitudinal surface slope ζ_1/L_S and bottom friction. The Strait is strongly stirred by tidal currents in the Narrows region included in the model (Bowman et al. 1983b), so we adopt a linear parameterisation of this friction (Csanady 1976). Then for the Strait, the longitudinal force balance yields

$$r v_S = g H_s \zeta_1/L_S \qquad (3)$$

where r is the bottom friction coefficient (1.6×10^{-3} m s^{-1}), and v_S is the (uniform) velocity in the Strait. Finally, continuity of flow between the Bight and the Strait yields

$$W_B H_B v_B = W_S H_S v_S \qquad (4)$$

Solution of Eqn 1–4 shows that

$$v_S = \frac{\sigma_w L_B H_S}{\rho\left(r L_S H_B - \frac{1}{2} f W_S H_S^2\right)} \qquad (5)$$

from which v_B may be found from Eqn 4, $(\zeta_2-\zeta_1)$ from Eqn 2, and ζ_2 from Eqn 1. If we set $\sigma_w = 0.2$ N m^{-2} (equivalent approximately to a wind of 10 m s^{-1}), and use the dimensions given above for the Bight and the Strait, then we find $v_S = 0.14$ m s^{-1}, $v_B = 0.06$ m s^{-1}, $\zeta_2-\zeta_1 = 0.034$ m and $\zeta_2 = 0.037$ m. In the denominator of (5), the second term is an order of magnitude larger than the first, showing that bottom friction in the Strait is much less important than the longitudinal surface slope in the Bight in determining the velocities v_S and v_B.

Published data are equivocal in providing tests of this model. Support for the estimate of transverse slope $S_t = 2(\zeta_2 - \zeta_1)/W_B$ lies in measurements published by Heath (1978: fig. 7). Components of the wind stress (measured near the outer extremity of Tasman Bay) were compared to measurements of sea level at Wanganui and New Plymouth. Near the beginning of the record, wind stress reached 0.3 N m^{-2} towards the west, veering to south-west; near the end the wind stress reached the same value towards the south-east. Assuming that the difference in level between New Plymouth and Wanganui varies only in response to changes in the transverse slope of the Bight, we can deduce the dependence of S_t on σ_w. The model predicts that $S_t = \alpha\sigma_w$ where

$$\alpha = \frac{-f \ L_B}{\rho \ g \ H_B \ W_B \left(\dfrac{L_S \ H_B}{W_S \ H_S^{\,2}} r - \frac{1}{2} f \right)} \qquad (6).$$

Using the values given above, this gives $\alpha = 3 \times 10^{-6}$ m^2 N^{-1}. The measurements show that, for an increase of 0.6 N m^{-2} in σ_w, the sea level at Wanganui rises c. 0.1 m as a result of the change in S_t, so that $\alpha = 0.2/(0.6\,W_B) = 3.0 \times 10^{-6}$ m^2 N^{-1}. This observed value for α cannot be regarded as certain, but the agreement with the theoretical value is encouraging.

On the other hand, published measurements of flow rate suggest that the model under-estimates the actual flow. It is not possible to derive a simple relationship between wind stress and flow velocity from Heath's (1978) data. The drift-card data of Heath & Shakespeare (1977) showed a median speed less than 0.1 m s^{-1}, but it is not clear how much of this flow was a response to the wind. The drifting buoys of Bowman et al. (1983a) showed values of v_B in the range 0.13–0.27 m s^{-1}, however, at a time of moderate winds for which σ_w = 0.2 N m^{-2} is probably a fair estimate. We conclude that the model under-estimates the wind-induced flow by a factor which probably lies between 1 and 2. The measurements of Heath (1978) showing a reduced flow below the thermocline in the Bight provide part of the answer, since we have assumed a uniform flow at all depths. Heath (1978: fig. 3) suggests that the eastward component of surface-layer velocity is about 50% greater than the same component in the deeper layer, as measured in water 71 m deep. Thus for the range of possible thermocline depths given by Heath, 25–50 m, our estimate of the flow speed v_B must be increased by a factor between 1.1 and 1.3 to give the surface speed. (This would not apply to v_S, since the Strait is strongly stirred by tidal currents). Any further underestimate can perhaps be ascribed to a wind-dependent difference in sea level between the west and east coasts of New Zealand in the presence of an eastward-directed regional wind. Extension of our model to allow for this shows that, in response to a setup ε of the west coast water relative to the east coast water, v_B would increase by

$$\Delta v_B = \frac{g \ \varepsilon}{W_B \left(\dfrac{L_S \ H_B}{W_S \ H_S^2} r - \frac{1}{2} f \right)} \qquad (7).$$

Thus both v_B and v_S would double in response to a setup of 0.036 m.

The prediction of the model which is important to considerations of upwelling originating near Kahurangi Shoals is that, when wind blows from the west or north-west, the sea level at Cape Farewell is expected to fall relative to the levels nearby along that coast.

The origin of the plume

The origin of the upwelling plume has been identified as inshore of Kahurangi Shoals; that is, this is where the deep water is found to reach the surface, before flowing northwards into the Bight. The plume is never observed to flow southwards, and we conclude that it only appears when there is a northward coastal flow present.

The papers which have been reviewed above imply that there are three distinct time scales which are important. The longest of these is about 4–6 weeks which seems to represent the upper limit of the persistence of the Westland Current. Once the plume has formed, the water in it is reported to be substantially mixed with the surrounding water after 10–15 days; this time scale has been put forward by Bowman et al. (1983c) as being biologically important. Finally, there is a time scale of a few days which is determined by meteorological fluctuations. Heath & Gilmour (1987a) have established the importance of this time scale in the fluctuations of the longshore flow near Kahurangi Shoals. Close inspection of the published observations suggest that the appearance at the surface of the coldest water, less than 13°C, can fluctuate on this shortest time scale. However the plume as a whole, when it is observed, tends to outlast the duration of the observational cruise, and therefore its existence seems to reflect the long time scale of variability of the Westland Current.

This range of time scales suggests that the Westland Current longshore flow is the first requirement for upwelling. This is essentially an intermittent upwelling coastal jet driven by longshore wind, but it is worth noting that in practice the wind is likely to have a substantial onshore component. The effect of this would be to diminish the upwelling associated with the jet, but not to prevent the jet from flowing.

So long as the Westland Current is flowing, then, there is likely to be upwelling along its path, in the sense of isopycnal surfaces tilting upwards towards the coast. It appears, however, that this effect is not itself sufficient to bring water to the surface from beneath the thermocline. Thus we seek mechanisms which might intensify the upwelling locally, to produce the Farewell plume. In doing so, we must account for the observed importance of wind directed to c. 128°T, which markedly accelerates the longshore flow at Kahurangi Shoals and is associated with intensified upwelling. The time scale of several days is long enough for this flow to be regarded as a mean flow in relation to the tidal and inertial time scales.

Heath & Gilmour (1987a), in discussing this wind-driven flow, tacitly assumed that wind observed at Farewell spit to be blowing towards 128°T would have been blowing toward 038°T at Kahurangi Shoals, so that it constituted a longshore wind there. However 128°T is actually the onshore direction there. To test whether the wind direction is significantly different at the two locations, records of wind direction made at sea were compared with simultaneous observations made at Farewell Spit. The comparison is shown in Fig. 6, and includes seaborne observations from the Kahurangi Shoals area, the Challenger Plateau area, and the Greymouth – Wanganui Bluff area. There is no clear bias in the last group, and although the first two do suggest a rotation of the wind direction at Farewell Spit in the required sense, it does not exceed 30°. Hence the acceleration of the alongshore flow is induced not by a local longshore wind but rather by an approximately onshore one.

We have shown that a moderate wind stress of 0.2 N m^{-2} directed approximately towards 115°T (that is, in a direction approximately on-shore at Kahurangi Shoals) acting on Taranaki Bight, lowers the sea surface at Cape Farewell by about 0.034 m, while raising it at Cape Egmont by the same amount. Hence

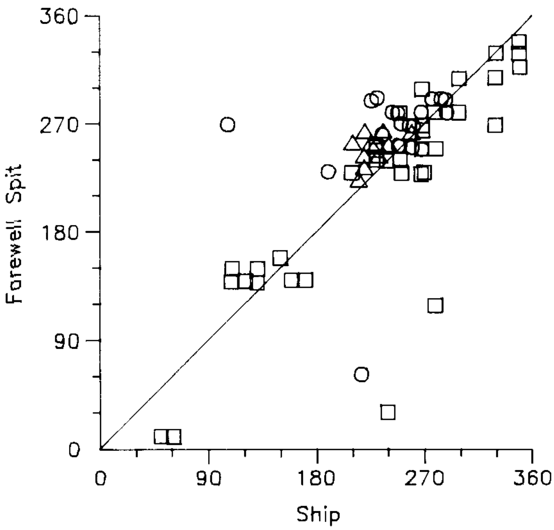

Fig. 6 Comparison of seaborne measurements of wind direction with Farewell Spit measurements. Ship located in: ○ Kahurangi region, △ Challenger Plateau, □ Greymouth-Wanganui Bluff region.

such a wind stress induces a surface slope downwards towards Cape Farewell, from all directions. If we assume that the longshore slope is experienced for roughly the same distance south of Cape Farewell as north of it, then it would extend at least to Steep Point, south of Kahurangi Shoals, with a value c. 3×10^{-7}. If this surface slope induces a flow in which the balancing force is linearised bottom friction, an equation like Eqn 3 applies and shows an increment to the velocity of c. 0.2 m s^{-1} for this wind stress and a water depth of 100 m. The fluctuations recorded by Heath & Gilmour (1987a: fig. 2) have a comparable magnitude, providing support for this mechanism, although the approximations and uncertainties involved make the quantitative agreement somewhat fortuitous.

Given, then, that the longshore flow near Kahurangi Shoals is accelerated by an onshore wind in this way, we need to consider how this acceleration contributes to the upwelling. Bowman et al. (1983c) have considered the distribution of vorticity induced in a homogeneous flow through the region by means of a numerical model. For reasonable flow velocities, they show that the bathymetry and bottom friction induce relative vorticity, the greatest magnitude of which is comparable with the planetary vorticity; and that the resulting convergence of bottom Ekman flow should give rise to upwelling. Furthermore, they show

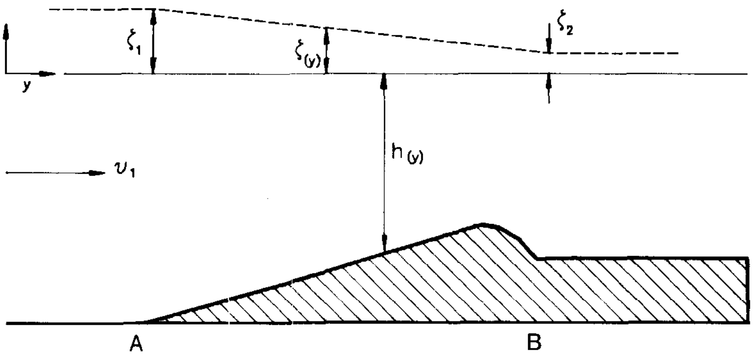

Fig. 7 Longshore cross-section of Kahurangi region. Tilted sea surface is dashed, sea floor hatched.

that the greatest upwelling velocity occurs on a transport streamline which passes close to Kahurangi Point inshore of Kahurangi Shoals, in agreement with observations of the plume's origin. The vertical velocity which they calculate for that streamline averages about 1 mm s^{-1}, and there is evidence that such a velocity can account for the observations. In the measurements reported in Fig. 3, the $S = 35 \times 10^{-3}$ isohaline lies at a depth of 25 m offshore. If the Steep Point and Kahurangi salinity sections are compared, it may be seen that this isohaline intersects the surface about 2.5 km offshore in the former and 8 km offshore in the latter. If one assumes a constant longshore velocity of 0.3 m s^{-1} between the two positions, a vertical velocity of 1 mm s^{-1} would raise the additional volume of deep water above the depth of 25 m in a distance of 8 km. This is much less than the distance between the two sections of about 20 km, although comparable with the distance up stream of the Kahurangi section at which the sea bottom starts to shelve as the flow approaches Kahurangi Shoals. The estimate of 1 mm s^{-1} for the vertical velocity therefore seems reasonable.

In spite of the encouraging agreement between the calculated and observed vertical velocities, though, there are two serious difficulties with the theory used by Bowman et al. (1983c). Firstly, the region of upflow which they predicted began near Kahurangi Point and extended to Cape Farewell, whereas the evidence shows that deep water surfaces near Kahurangi Point, which implies that the upflow starts further up stream. Secondly, their numerical model did not resolve the Ekman layer flow, and this had to be calculated indirectly so that the vertical velocity could be found from its divergence. This calculation rested on the boundary layer theory of Pedlosky (1979: 178) which is only valid when the Rossby number (and certain other parameters of the flow) are small compared to unity. In this case, however, we are dealing with variations of vorticity with a typical horizontal scale of about 3 km, yielding a Rossby number of unity when the flow velocity is 0.3 m s^{-1}. In physical terms, the variations of flow are on too small a horizontal scale for the Coriolis force to adjust the Ekman layer; and the resulting vertical velocity, which derives from the horizontal variations of Ekman flow, will be much less than the value given. We must therefore seek an alternative explanation for the observed vertical velocity.

Consider a longshore cross-section of the flow region inshore of Kahurangi Shoals. The essential elements of the bathymetry and sea-surface topography are illustrated in Fig. 7. The flow is assumed to be in the plane and geostrophic except for a thin bottom Ekman layer. Flow is from left to right, and a horizontal pressure gradient is maintained in the plane by a slope of the sea surface, which departs from a level surface by a small elevation $\zeta(y)$. The water depth below this level surface is $h(y)$. Along a streamline of the geostrophic flow near the bottom, Bernoulli's law gives

$$p_S + \frac{1}{2}\frac{\rho Q^2}{h^2} = \text{constant} \tag{8}$$

where the pressure $p = p_S + \rho g h$, ρ is the density, and $Q = vh$ is the volume transport which is independent of y. Then between the two ends of the region of sloping bottom, we can write for this streamline

$$p_{S1} - p_{S2} + \frac{1}{2}\rho Q^2\left(\frac{1}{h_1^2} - \frac{1}{h_2^2}\right) = 0 \tag{9}$$

where $p_{S1} - p_{S2} = \rho g(\zeta_1 - \zeta_2)$. Hence the change in sea level between these points is

$$\zeta_1 - \zeta_2 = -\frac{Q^2}{2g}\left(\frac{1}{h_1^2} - \frac{1}{h_2^2}\right) = \frac{v_1^2}{2g}\left(\frac{h_1^2}{h_2^2} - 1\right) \quad (10)$$

Taking $v_1 = 0.3$ m s^{-1}, $h_1 = 90$ m, $h_2 = 60$ m, gives $\zeta_1 - \zeta_2 = 0.006$ m. If the sea surface were not higher at A than at B by this amount, the deep water would not surmount the rise in the sea bed, and would instead be diverted offshore to flow around Kahurangi Shoals. It should be noted that the surface slope $(\zeta_1 - \zeta_2)/L$ is needed to supply the extra kinetic and potential energy required to flow over the bathymetry; it is additional to any slope which is needed to balance the force of bottom friction.

To calculate the effect of such a flow on the depth of the thermocline, we may undertake a similar application of Bernoulli's law for two streamlines just above and just below the thermocline which is assumed to be thin and to lie at a depth $t(y)$. The method is similar to that used by Turner (1973: 64) for the case where the upper layer is stationary. It may be shown that, as h changes, the corresponding change of t is given by

$$S = \frac{dt}{dh} = \frac{F_b^2}{F_b^2 + F_u^2 - 1} \quad (11)$$

where $F_b^2 = v_b^2/g'(h-t)$, $F_u^2 = v_u^2/g't$, v_b and v_u are the flow speeds in the lower and upper layers respectively, $g' = g\Delta\rho/\rho$ and $\Delta\rho$ is the difference in density between the two layers. F_b and F_u are the Froude numbers of the flows in the two layers. If we evaluate S at point A in Fig. 7, with $t = 5$ m, $b = 85$ m, $v_u = 0.3$ m s^{-1}, $v_b = 0.2$ m s^{-1}, we find $F_b^2 = 0.047$, $F_u^2 = 1.8$ and $dt/dh = -0.055$. F_u is larger than unity because t is small, so t decreases, and the thermocline rises, as the water gets shallower. The effect is initially modest, but becomes more significant as the water depth h decreases, since v_b increases simultaneously (the transport being constant) and increases approximately as the inverse cube of h. Thus near the minimum depth, where $h \doteqdot 40$ m, $F_b^2 = 0.54$ and $S = -0.4$. The average value of S is -0.12, so the thermocline rises about 5 m between A and B. This would just bring it to the surface inshore of Kahurangi Shoals, for the conditions assumed. The average vertical velocity of the thermocline in this process is $-v_u(dt/dh)(dh/dy)$, however, and this is only about 0.2 mm s^{-1}. Thus an additional effect must be present if the observed vertical velocity of about 1 mm s^{-1} is to be explained.

The hydraulic calculations show that, given a favourable longshore slope of the sea surface, the longshore flow can surmount the elevation of the sea bed inshore of Kahurangi Shoals, with a consequent rise of the thermocline. However it takes no account of the Ekman flow associated with bottom friction. The volume transport in this flow is rv/f which is about 5 m^2 s^{-1}, shoreward, at points about 5 km offshore at the seaward edge of the region of high cross-shore bottom slope. (There must, of course, be a balancing offshore transport in the geostrophic flow above the Ekman layer.) The Ekman transport must become zero at the shoreline itself, as discussed by Csanady (1982: 189), and continuity requires that this reduction be accommodated as a vertical transport, with vertical velocity of order 5/5000 m s^{-1} = 1 mm s^{-1}. Thus the final important effect in bringing deep water to the surface is a coastal convergence in the bottom Ekman layer. Added to the hydraulic effect, this brings to the surface, inshore of Kahurangi Shoals, water which was previously at the bottom a similar distance offshore at a depth of up to about 100 m.

CONCLUSION

We have shown that westerly wind events lasting several days will establish a slope of the sea surface downwards from about Steep Point to Cape Farewell, and that this will accelerate the Westland Current locally. A small intensification of this slope over the shelving region south of Kahurangi Shoals then enables deep water to flow up and over the saddle inshore of the Shoals. At the same time its speed is increased, and the near-shore convergence of the bottom Ekman flow increases correspondingly. This, supported by a tendency for the thermocline to rise under the prevailing hydraulic conditions, produces the intense upwelling which is observed near the Shoals. When the westerly wind declines, the coastal jet can still flow, but upwelling is possible only from shallower depths. The deeper water then flows outside Kahurangi Shoals. The succession of wind events thus strongly modulates the surface temperature in the plume, leading sometimes to coherent cold patches down stream. It is also likely that the upwelling is modulated tidally, but there is no evidence that this is a large effect; nor is there any reason to believe that the spring-neap cycle is important.

It has been noted that warm water sometimes moves southwards round Cape Farewell, inshore of the upwelled plume. We interpret this as resulting from a reversal of the wind-induced surface slopes, which causes the southward current when the westerly wind drops. The role of fresh water, as seen for example in the Whanganui section (Fig. 3H), may also be significant in establishing this current.

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