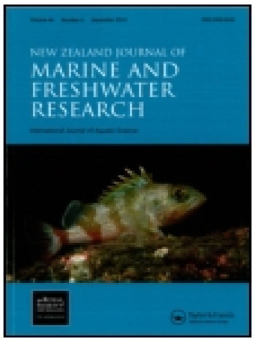

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What drives the mean circulation on the New Zealand west coast continental shelf?

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Abstract A linearised model of Csanady (1976) for the time-averaged mean flow on a continental shelf is applied to the New Zealand west coast continental shelf. The means are time averages over time scales that are long relative to that of local atmospheric forcing. Forcing of the flow is via the sea surface elevation (specified from deep ocean steric levelling with adjustment for boundary effects), mean winds, and thermohaline forcing (specified by observations and determined by the strong fluctuating flow, the time average of which is the weaker mean flow). The main flow components are found generally to be caused by forcing by the sea surface elevation and local winds. North of latitude 42°30′S the mean sea surface slopes down towards the south, with an associated southward-directed flow component; south of 42°30′S it slopes down towards the north with an associated northward-directed flow component. On the north-west coast of the South Island the predominant north-eastward-directed wind produces a north-eastward along-shore flow component. On the west coast of the North Island north of Cape Egmont the predominant onshore wind produces a flow towards the north. Thermohaline forcing leads to a southward along-shore flow decreasing to zero near the sea floor.

Keywords winds; water circulation; continental shelves; thermal properties; sea level; wind-driven currents; mathematical models.

INTRODUCTION

Physical oceanographic studies on the continental shelf of New Zealand's west coast indicate that the flow is highly variable, with close day-to-day correlation with atmospheric forcing (Heath 1978; Sanderson 1979). The mean circulation is not well defined, and in view of its variability can probably be more clearly defined only by an extensive programme of current observations, or through examination of possible driving mechanisms followed by a selective observational programme. Possible flow driving mechanisms are studied in this paper.

New Zealand lies athwart of what, without the presence of the extensive New Zealand submarine platform, would be a generally eastward-directed zonal flow. New Zealand effectively splits this eastward flow, with the resulting strong flows around the northernmost (Stanton 1976a) and southernmost (Heath 1981) extremities contributing to bathymetrically controlled flows on the New Zealand east coast. Associated with the eastward flow far offshore from the New Zealand west coast there is a decrease in sea surface elevation of 0.3 m from north to south over the 12.5° latitudinal range of New Zealand. This slope has the potential to produce strong boundary forcing on the New Zealand west coast.

The New Zealand land mass also influences meteorological conditions, the high axial mountain chain on the South Island inducing strong variable winds on the west coast and a high annual precipitation with a consequent high freshwater discharge. Both the wind and the freshwater discharge are driving mechanisms contributing to the mean flow.

Morphology of the west coast continental shelf

The New Zealand west coast continental shelf narrows towards both north and south from the middle of New Zealand. Here it is about 75 km wide from the shelf break to a line across western Cook Strait between Cape Egmont and Cape Farewell aligned with the main orientation of New Zealand (Fig. 1), with a further 180 km across the indentation of western Cook Strait.

The shelf narrows to only 20 km at the northernmost extremity of New Zealand, 625 km north of Cook Strait, and is practically non-existent south of Big Bay, about 625 km south of Cook Strait (Fig. 1). West of Cook Strait the continental slope extends down to only 750 m depth on to the Challenger Plateau, at the southern end of the Lord Howe Rise, which extends north-westward into the Tasman Sea. North and south from the Challenger Plateau the west coast continental slope extends to a depth of at least 2000 m.

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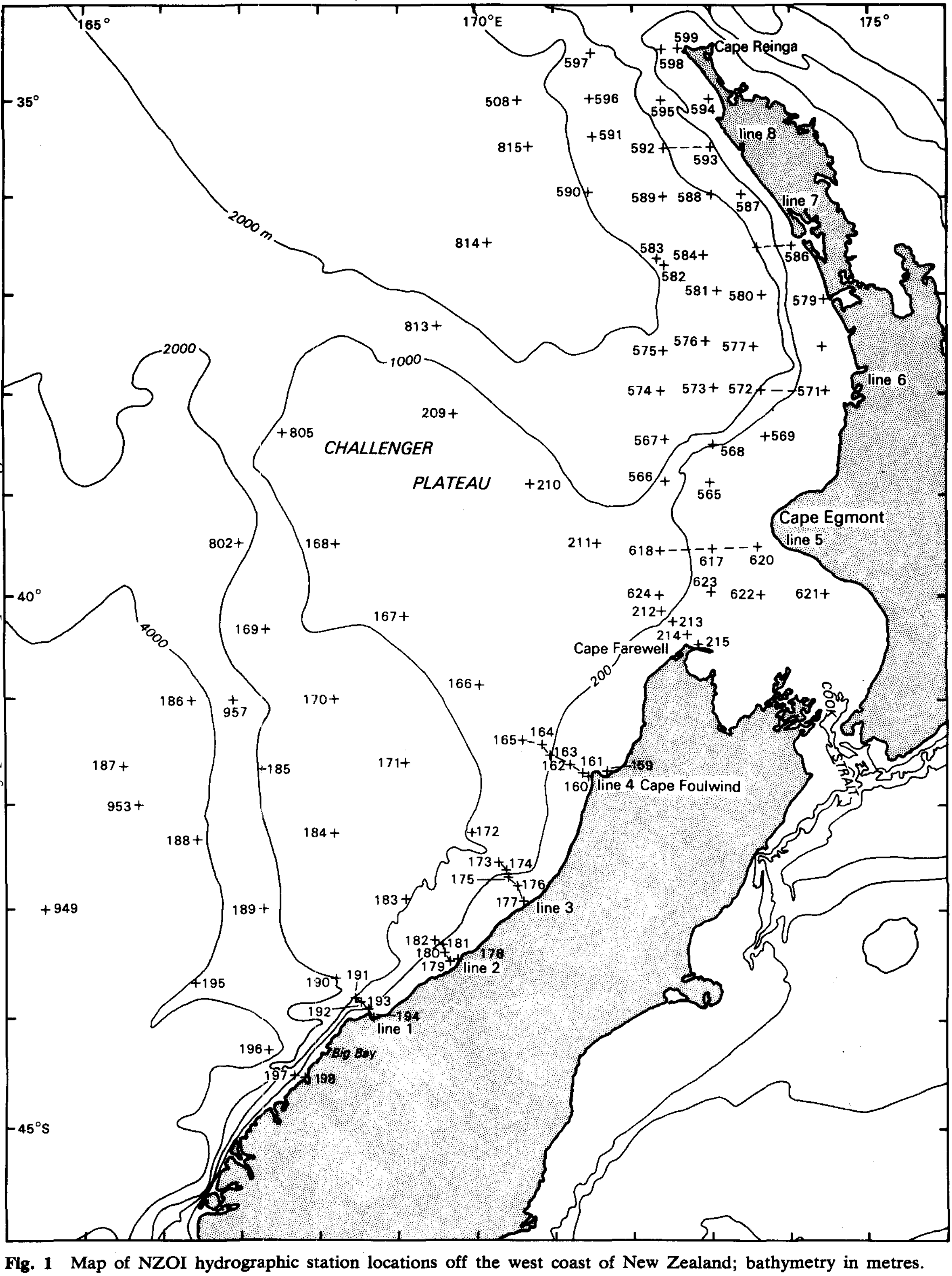

Fig. 1 Map of NZOI hydrographic station locations off the west coast of New Zealand; bathymetry in metres.

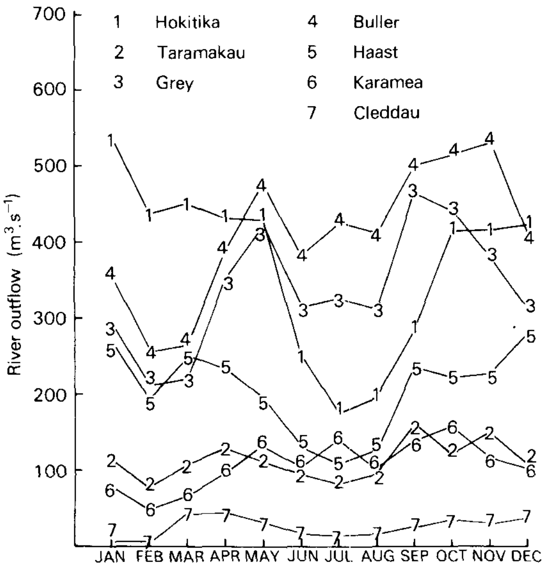

Fig. 2 Monthly mean freshwater discharges from some major west coast rivers, South Island, New Zealand.

Knowledge of the west coast mean flow

Historically, the mean flow on the west coast continental shelf was defined from mariners' reports and drift-card evidence (e.g., Brodie 1960). More recently water properties (Stanton 1973, 1976b; Ridgway 1980) and current drogues and meters (Heath 1973, 1978; Sanderson 1979) have been used to define the circulation. The picture that emerges from these studies is of a very weak mean flow on the continental shelf, northward on the west coast of the South Island (the Westland Current) towards Cape Egmont, with at least part of this flow contributing to the D'Urville Current, which flows eastward into Cook Strait. The direction of flow on the continental shelf north of Cape Egmont is not at all clear. On the basis of drift-card evidence, Brodie (1960) indicates a converging of the currents towards about latitude 37°S, with southward-directed currents north of 37°S (the West Auckland Current) and northward-directed currents south of 37°S (an extension of the Westland Current). However, current-meter observations at 38°12′S, under calm atmospheric conditions, indicate a mean southward flow (Heath 1978).

Current-meter observations (D. M. Garner, University of Auckland, pers. comm.; Heath 1978) and radio-tracked drogue studies (Sanderson 1979) indicate that the day-to-day flow is dominated by a response to atmospheric forcing. This response consists of both a direct wind-induced near-surface flow and a barotropic flow resulting from changes in the sea-surface slope induced by blockage of the direct wind-induced Ekman Transport by the coastal boundaries.

To understand further the circulation on the west coast, it is necessary to determine the influence of the 3 most probable driving mechanisms—boundary, atmospheric, and thermohaline forcing. The difficulty lies in separating the individual influences of the forcing, for there are non-linear interactions between the individual flows, primarily through frictional dissipation. Csanady (1976) has shown that if there is a variable flow present which is considerably stronger than the mean flow (in the present context, the strong atmospherically induced flow), then the non-linear frictional terms in the momentum equations for the mean flow can be approximated by terms which are linear with respect to the mean flow and have coefficients determined by the stronger variable flow, and terms depending only on the stronger variable flow and therefore prescribable. It is Csanady's (1976) model which is used here.

MEAN CIRCULATION MODEL

Csanady (1976) separated the various water movements according to their temporal scales. The first-order flow is that with long periods relative to the tidal cycle, and in the momentum equation for this flow the motions with short periods relative to the tides appear as Reynolds stresses of the first-order flow. Time averages of the first-order flow give the required mean. In the momentum equations for the means, the average Reynolds stress (F_0) reduces to (1) a term of the product of the gradient of the mean flow (e.g., $\partial u/\partial z$) and a stress coefficient (A_x, whose magnitude is determined by the first-order flow) and (2) a correlation term ($\langle\rangle$) linking the fluctuating parts of the stress coefficient (A^1) and the gradient of the first-order flow ($\partial u^1/\partial z$). The former term is linear with respect to the mean flow and the latter is determined entirely by the first-order flow, i.e.,

$$F_0 = A_0 \frac{\partial u}{\partial z} + \left\langle A^1 \frac{\partial u^1}{\partial z} \right\rangle$$

where $u_1 = u^1 + u$ is the first-order flow, $\langle u_1 \rangle = u$, $\langle u^1 \rangle = 0$, and $A_1 = A^1 + A_0$, $\langle A_1 \rangle = A_0$, $\langle A^1 \rangle = 0$.

The momentum equations for the mean flow are

$$-fv = -g \frac{\partial}{\partial x} (\zeta + D) + \frac{A_0}{\rho} \frac{\partial^2 u}{\partial z^2}$$

and

$$fu = -g \frac{\partial}{\partial y} (\zeta + D) + \frac{A_0}{\rho} \frac{\partial^2 v}{\partial z^2}$$

in a co-ordinate system with the x axis pointing onshore, the y axis positive along-shore toward the north-east, and the z axis positive upwards; u, v the velocity components positive in the x and y directions respectively; ζ the elevation of the sea surface; g the acceleration of gravity; D the dynamic height; and A the momentum exchange coefficient. These equations, subject to the boundary conditions at the surface ($z = 0$) and bottom ($z = -h$)

$$A = \frac{A_0}{\rho} \quad , \quad F = \frac{F_0}{\rho}$$

$$A \frac{\partial u}{\partial z} = F_{wx} \qquad A \frac{\partial v}{\partial z} = F_{wy} \qquad \text{at } z = 0$$

$$A \frac{\partial u}{\partial z} = ru \qquad A \frac{\partial v}{\partial z} = rv \qquad \text{at } z = -h$$

where F_{wx}, F_{wy} are the surface wind-stress components and r is a linear bottom-drag coefficient, and at the open boundaries the transport conditions

$$\int_{-h}^{0} u \, dz = 0 \qquad \int_{-h}^{0} v \, dz = -V$$

have a solution

$$w = -(ig/f)(S + z/L) + a \exp(1 \mp i)kz + b \exp[-(1 \mp i)kz]$$

(where 2 alternative operators are shown, the upper signs apply to the Southern Hemisphere, the lower to the Northern Hemisphere, for which Csanady (1976) originally formulated the model)

where

$$w = u + iv; \qquad i = \sqrt{(-1)}$$

$$S = -\frac{\partial \zeta}{\partial x} - i \frac{\partial \zeta}{\partial y};$$

$$F = F_{wx} + iF_{wy}; \qquad k = (|f|/2A)^{\frac{1}{2}}$$

with the dynamic height gradient approximated by

$$-\frac{dD}{dx} = \frac{z}{L}$$

The boundary and transport conditions for the non-dimensional variables

$$\alpha = fa/g \qquad \beta_1 = fb/g \qquad \lambda = kh$$
$$\phi = F/gh \qquad \delta = h/L \qquad \tau = fV/gh$$
$$R = r/fh \qquad q = \exp(1 \mp i)\lambda$$

become

$$\alpha - \beta_1 = \lambda\phi (1 \pm i) \frac{f}{|f|} + \left(\frac{\mp 1 + i}{2\lambda}\right) \delta$$

$$\alpha q^{-1} (\mp 1 - i + 2i\lambda|R|) - \beta_1 q (\mp 1 - i - 2i\lambda|R|) + 2\lambda|R|S = 2\lambda|R|\delta + \frac{\delta}{\lambda}$$

$$\alpha (1 - q^{-1}) - \beta_1 (1 - q) + (\mp 1 - i) \lambda S = (\mp 1 - i) \frac{\lambda}{2} \delta + \lambda\tau (\mp 1 - i)$$

from which, with the non-dimensionalised wind stress ϕ, dynamic height δ, and the along-shore surface slope (S_y) specified, the velocities can be evaluated. For the situation where the water depth h is sufficiently greater than the Ekman depth that $\exp \lambda \gg 1$, we have

$$S[\,|R| - \lambda|R| \pm i(1 + \lambda|R|)\,] = -[\lambda|R| \mp i(1 + \lambda|R|)]\left[\tau + \frac{\delta}{2} \pm \phi \frac{f}{|f|}\right] + \delta|R|\left(1 - \frac{1 \pm i}{2\lambda}\right) \qquad \ldots(1)$$

from which the offshore surface slope S_x and the along-shore transport can be evaluated. To this approximation the non-dimensionalised surface (W_s), bottom (W_b), and mid-depth ($W_{h/2}$) velocities are given by

$$W = \frac{f}{g} w$$

$$W_s = -iS + 2\beta_1 + \lambda\phi \frac{f}{|f|} (1 - i) + \left(\frac{\mp 1 + i}{2\lambda}\right) \delta \qquad \ldots(2)$$

$$W_b = \left(\phi - \tau + S - \frac{\delta}{2}\right) / R \qquad \ldots(3)$$

$$W_{h/2} = -i\left(S - \frac{\delta}{2}\right) \qquad \ldots(4)$$

The forcing parameters ϕ_x, ϕ_y, S_x, and δ enter linearly in these equations, and thus allow each component to be evaluated separately. In order to determine the relative contributions to the mean circulation it is necessary to determine the forcing parameters appropriate to the west coast situation.

FORCING FUNCTIONS

Wind stress

On the west coast of the North Island north of Cape Egmont (Fig. 1) the predominant wind is onshore (i.e., directed eastwards) with a slight along-shore component towards the north (Tomlinson 1976).

On the north-west coast of the South Island the predominant wind is from the south-west (Tomlinson 1976), with wind speed increasing to seaward (Neale & Thompson 1978). This wind curves into Cook Strait from the west to give rise to the dominant south-eastward-directed wind in western Cook Strait. The other pronounced wind far offshore from the north-west coast of the South Island is from the north-west, but veers clockwise on approaching the coast to come from the north-east on the coast (Neale & Thompson 1978). The wind veer is predominantly anti-clockwise around the south-western coast of the South Island.

From these winds alone we would expect to find: (1) a weak flow along-shore to the north on the west coast of the North Island north of Cape Egmont (Ekman Transport towards the north); (2) a stronger flow along-shore towards the north on the north-west coast of the South Island and eastward into Cook Strait (the wind stress balanced by an along-shore bottom stress); (3) a northward-directed flow along the south-western coast of the South Island, progressively weakening further south to become a simple Ekman Transport (where the predominant wind is onshore) and then strengthening along-shore towards the south (the along-shore wind stress balanced by an along-shore bottom stress).

Freshwater inflow

The west coast of the South Island has in excess of 2.4 m of rain annually, with the high country south of Hokitika having over 6.4 m (Tomlinson 1976). This rainfall feeds several major rivers which drain on to the west coast continental shelf.

Plots of the monthly mean freshwater discharges from the 3 largest rivers draining on to the west coast of the South Island (Fig. 2) indicate that, in general, the discharges are greatest in early winter (April–June) and spring (September–November) and least in late summer (February) and mid winter (July–August).

The influence of this freshwater inflow extends well offshore, and substantially reduces the salinity. For example, Stanton (1976b) found that the freshwater influence extended 56 km offshore in March 1969, and in April 1974 there was 1% fresh water on the west coast continental shelf out to the 200 m isobath. The amount of fresh water on the west coast shelf can probably vary by several orders of magnitude (Stanton 1976b).

On the west coast of the North Island north of Cape Egmont the freshwater inflow is substantially less than that on the west coast of the South Island; the only large river, the Waikato, has an average inflow of 420 $m^3 s^{-1}$ (Adams 1979). Comparison of the salinity distributions on the west coast of the North Island in February–March 1974 (Ridgway 1980, fig. 4) and on the west coast of the South Island in April 1974 (Stanton 1976b) indicates that the influence of freshwater inflow is substantially larger off the South Island. This influence produces an along-shore flow towards the south which decreases with depth.

Sea level differences

Well offshore in deep water the slope of the sea surface can be determined using steric levelling. The height of the sea surface is given by the dynamic height (or geopotential) anomaly relative to a deep reference surface which is assumed to be level. This slope offshore, however, cannot simply be projected inshore on to the continental shelf to be used alone as the required forcing mechanism, for boundary effects may change the slope. These effects include: guiding of the flow from essentially onshore in the deep ocean to along-shore in coastal areas; the associated variation in sea surface elevation needed to balance the change in Coriolis acceleration with latitude; and the partial balance of the sea surface elevation by an onshore momentum flux. Land levelling between tide gauges, can, in theory, be used to determine the mean sea surface slope at the coast but, in practice, on a coast such as the west coast of New Zealand where the mean flow is known to be weak, the actual slope is at the limit of experimental error in the available observations. Plots of sea surface elevation relative to different reference levels down to 1000 m and at different distances offshore out to about 300 km, approximately parallel to the west coast of New Zealand, are shown in Fig. 3. The plots were constructed from data collected in the austral summer (Garner 1967, 1970). Beyond 300 km offshore the sea surface slope relative to different reference levels below about 500 m does not change significantly (see Heath 1980, fig. 6). At about 300 km offshore the sea surface slope relative to 500 m (12.8×10^{-8}) is greater than that relative to 1000 m (6.2×10^{-8}), which is about the same as that relative to 100 m (6.6×10^{-8}). The main difference in these slopes occurs south of the Challenger Plateau (Fig. 3), and is indicative of strong shear in the geostrophic flow in this region

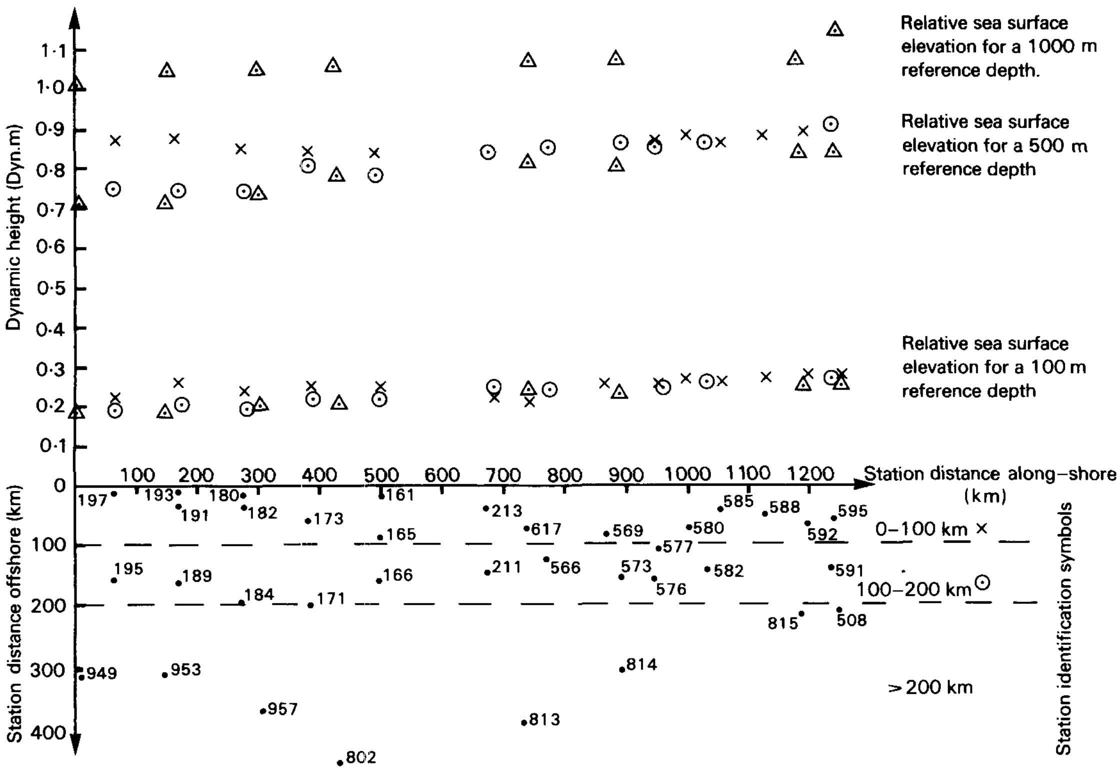

Fig. 3 Along-shore dynamic heights of the sea surface relative to different reference levels at different distances off the west coast of New Zealand; station positions shown in Fig. 1.

(Garner 1967; Heath 1972). Either there is a minimum flow near 500 m with the flow above and below this depth having a south-eastward component, or the flow is towards the north-west at about 500 m with the surface and deep flows directed towards the south-east. Examination of the salinity distribution on the surface of minimum salinity, associated with Antarctic Intermediate Water, would suggest that, at least offshore, the latter situation is most likely.

The relative sea surface slope south of the Challenger Plateau flattens out as the coast is approached (Fig. 1 and 3); this is indicative of the deflection of the flow towards the south, as found by Stanton (1976b). Ridgway (1981, fig. 13) also shows a southward trend in the weak flow in the deep ocean off the west coast of the North Island. Relative to 500 m the overall sea surface slope decreases from 12.8×10^{-8} at about 300 km offshore to 5.1×10^{-8} at about 60 km offshore. The respective sea surface slopes relative to 100 m are 6.6×10^{-8} and 2.7×10^{-8}. South of the Challenger Plateau, close inshore, the sea surface slope appears to be upwards to the south; for example, the slope of the sea surface relative to an isobaric surface at 500 m for stations in approximately 600 m depth on the continental slope (Stations 173, 182, and 191 in Fig. 1 and 3) is -1.5×10^{-7}. Much of the change in density leading to the increase in the relative sea surface elevation towards the west coast of the South Island results from the lowering of the salinity by freshwater inflow. This is illustrated by the change in the dynamic height of the sea surface associated with the decrease in salinity observed near the coast, with the temperature held constant (Table 2); the changes in dynamic height are about equally divided between salinity and temperature variations.

In brief, steric levelling gives values of 6.2×10^{-8} and -1.5×10^{-7} for the slope of the sea surface along the continental slope off the west coast of New Zealand north and south respectively of about 42°30′S. These slopes should be compared with those arising from the need to balance the latitudinal change in the Coriolis acceleration and that balancing the onshore momentum flux.

From the steady-state, along-shore momentum equation for the first-order and mean flows with flow components of short period relative to the tidal cycle incorporated in Reynolds stresses $A\, \partial v/\partial z$, τ_{yx}, τ_{yy} (of which only the first is considered in Csanady's (1976) model)

Table 1 Variations in the dynamic height of the sea surface relative to 100 m associated with observed along-shore variations in temperature and salinity on the west coast of the South Island, April 1974 (data from Stanton 1976b). O, offshore; C, coastal. Dynamic heights: A, with observed temperature and salinity offshore; B, with observed temperature and salinity coastal; C, with offshore observed temperature and coastal observed salinity.

Stn no. (Fig. 1)	Position (°'S)	Position (°'E)	Dynamic height A	Dynamic height B	Dynamic height C	Salinity difference, O cf. C (σT)
195	43°38.7′	166°24.0′(O)	0.207		0.234	0.37
197	44°31.5′	167°37.8′(C)		0.273		
189	42°59.0′	167°16.2′(O)	0.215		0.275	0.83
193	43°55.2′	168°35.0′(C)		0.306		
184	42°16.4′	168°09.0′(O)	0.207		0.239	0.43
180	43°21.6′	169°35.4′(C)		0.265		
166	40°54.6′	170°01.6′(O)	0.231		0.241	0.14
161	41°41.7′	171°20.5′(C)		0.260		

$$fu = -g\frac{\partial}{\partial y}(\zeta + D) + A\frac{\partial^2 u_1}{\partial z^2} + \frac{\partial \tau_{yy}}{\partial y} + \frac{\partial \tau_{xy}}{\partial x}$$

integrating vertically with $\partial D/\partial y = 0$ (there is little along-shore change in density) and $\partial\tau_{yy}/\partial y = 0$ we have

$$-\int_{-H}^{0} g\frac{\partial \zeta}{\partial y}\,dz + F_{yr} - F_{yb} + \int_{-H}^{0}\frac{\partial \tau_{xy}}{\partial x}\,dz = 0$$

where F_{yr}, F_{yb} are the surface wind stress and bottom stress components respectively in the along-shore direction. Integrating again across the continental shelf from the shore ($x = 0$) to a distance $-L$ offshore gives

$$-g\left(\frac{\partial \bar{\zeta}}{\partial y}\phi + \int_{-L}^{0}\int_{-H}^{0}\frac{\partial \zeta^1}{\partial y}\,dz\,dx\right) = -(\overline{F_{yr}} - \overline{F_{yb}})L + H\tau_{yx}(L)$$

where ϕ is the cross-sectional area across the shelf to $x = -L$, $\tau_{yx}(L)$ is the depth-averaged value of τ_{xy} at $x = -L$; $\overline{F_{yr}}$, $\overline{F_{yb}}$ are the average values across the shelf, and the along-shore gradient of the sea surface elevation has been split into a mean across the shelf $\partial\bar{\zeta}/\partial y$ and a component giving the variation across the shelf $\partial\zeta^1/\partial y$, where

$$\int_{-L}^{0}\frac{\partial \zeta^1}{\partial y}\,dz\,dx = 0$$

An estimate of $\tau_{yx}(L)$ on the open coast is given by the current-meter records at depths of 11 m and 79 m in 106 m of water collected over a 19-day period 37 km offshore at 38°12′S, 174°12′E (Fig. 1). Velocity components from these records have been expressed in terms of a time-averaged mean and a fluctuating contribution ($u = u_0 + u^1$, $v = v_0 + v^1$) with the axes chosen such that the axis coincides with the direction of the mean flow. The stress τ_{xy} is then given by $\tau_{xy}/\rho = -uv$, which gives values of 32 $\times 10^{-4}$ and 83 $\times 10^{-4}\,m^2\,s^{-2}$ for the current meters at 11 m and 79 m respectively. With the cross-sectional area ϕ taken as $\frac{1}{2} L \times H = 1.96 \times 10^6\,m^2$, the mean momentum flux would balance a sea-surface slope of 3.2×10^{-8} downwards towards the south. The along-shore bottom speed needed to balance the same sea-surface slope via the bottom stress $\tau_{yb} = \rho\, v^2 C_D$, with ρ the water density and C_D ($= 2 \times 10^{-3}$) the drag coefficient, is 0.09 m s^{-1} towards the south.

An estimate of the change in along-shore sea surface elevation across the continental shelf due to the latitudinal variation in the Coriolis acceleration is given by integrating the assumed along-shore geostrophic flow balance

$$-fv = -g\frac{\partial \zeta}{\partial x}$$

across the shelf and then differentiating along shore, i.e.,

$$\beta\bar{v}P + f\frac{\partial \bar{v}P}{\partial y} = g\frac{\partial \Delta h}{\partial y}$$

where $\beta = \partial f/\partial y$, Δh is the change in surface elevation across the shelf from $x = -P$ to $x = 0$

and $\bar{v}$ is the along-shore surface speed averaged across the shelf.

Given a uniform along-shore flow on the continental shelf ($\partial \bar{v}P/\partial y = 0$), at a mean latitude of 41° for the west coast of New Zealand ($\beta = 1.7 \times 10^{-11}\ s^{-1}\ m^{-1}$), with $P = 7.5 \times 10^4$ m, the change in slope for a 0.1 m s^{-1} flow is 1.35×10^{-8}, with the near-shore sea surface elevation increasing towards the south for a southward-directed flow and increasing towards the north for a northward-directed flow.

To obtain more meaningful estimates of the contributions to the along-shore sea-surface slope, the 2 boundary effects can be combined. The 2 areas of differing sea surface slope will be dealt with separately.

North of 42°30′S. If we add half the expected change in meridional slope across the continental shelf ($vP/2g$) β to the along-shore slope in sea surface over deep water north of 42°30′S (6.2×10^{-8}), we obtain the mean along-shore slope on the continental shelf

$$\left[-H\tau_{yx}(L)\frac{P}{L} + (\bar{F}_{yT} - \bar{F}_{yb})P \right] \Big/ g\phi$$

where the along-shore mean speed for zero surface stress ($\bar{F}_{yT} = 0$) and quadratic bottom stress is given by

$$v = \left[\frac{\beta P}{2g} \pm \sqrt{\left\{ \left(\frac{\beta P}{2g}\right)^2 \pm \frac{4P\alpha C_D}{g\phi}(6.2\times10^{-8}) + \frac{H\tau_{yx}(L)}{g\phi}\,\frac{P}{L} \right\}} \right] \Big/ \pm \frac{2PC_D\alpha^2}{g\phi}$$

where the depth-integrated Reynolds stress $H\tau_{yx}$ is increased in the ratio of the width of the continental shelf (P) to the distance offshore of the current meter site (L), and α is the ratio of the near-bottom to surface speeds. The positive signs inside the square root and in the denominator are used for v negative and the negative signs for v positive. With the values given above (and with $P = 7.5 \times 10^4$ m, $H = 2.5 \times 10^2$ m, $C_D = 2 \times 10^{-3}$), the mean flow must be towards the south with a speed of 0.15 m s^{-1} for an assumed uniform flow from top to bottom and 0.25 m s^{-1} for a flow which decreases to 50% of the surface speed just above the bottom. The main balance is between the sea-surface slope and the bottom stress. For example, under the assumption of uniform flow, 21% of the sea-surface slope is accounted for by the Reynolds stress flux, and is reduced from 6.2×10^{-8} in the open ocean to 4.2×10^{-8} at the coast by the effect of the latitudinal variation of the Coriolis acceleration.

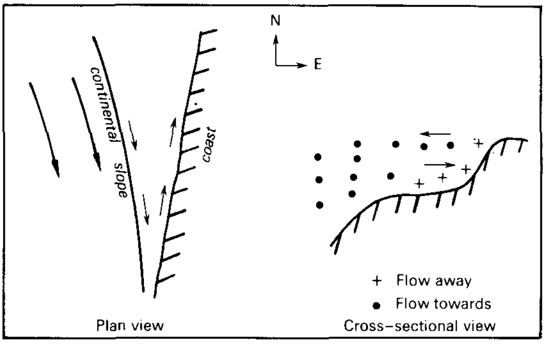

Fig. 4 Pictorial representation of the South Island mean circulation on the continental shelf and slope south of Cape Foulwind.

Table 2 Components of flow (m s^{-1} $\times 10^{-2}$ = cm s^{-1}) positive onshore (u) and along-shore (v) towards the north for sea surface slope, thermohaline, and wind forcing. Off the west coast of the South Island the wind is taken as 10 m s^{-1} along-shore towards the north, and off the west coast of the North island as 5 m s^{-1} on shore. h, indeterminate but small.

		Mid-shelf (50 m)		Outer shelf (100 m)	
		u	v	u	v
South Island south of 42°30′S					
Surface	slope	−1.5	11.0	−1.5	22.7
	thermo	−1.6	−5.9	−1.6	−15.8
	wind	−9	31	−9	31
Mid-depth	slope	−1.5	11.0	−1.5	22.7
	thermo	0	−2.6	0	−7.5
	wind	0	22	0	22
Bottom	slope	+1.6	4.6	3.9	9.2
	thermo	+0.77	0	0.8	0
	wind	4	9	4	9
South Island north of 42°30′S					
Surface	slope	3	−2	0.3	−5
	thermo	−1.6	−5.9	−1.6	−15.8
	wind	−9	31	−9	31
Mid-depth	slope	3	−2	0.3	−5
	thermo	0	−2.6	0	−7.5
	wind	0	22	0	22
Bottom	slope	−0.3	−1	−0.8	−2
	thermo	0.8	0	0.8	0
	wind	4	9	4	9
North Island north of Cape Egmont					
Surface	slope	0.3	−2	0.3	−5
	thermo	h	h	h	h
	wind	3	3	3	3
Mid-depth	slope	0.3	−2	0.3	−5
	thermo	h	h	h	h
	wind	h	h	h	h
Bottom	slope	−0.3	−1	−0.8	−2
	thermo	h	h	h	h
	wind	h	h	h	h

On the west coast continental shelf the speeds associated with the first-order flow and tides are substantially larger than the mean flow. In this situation it may be more meaningful to use a linear-bottom frictional law (see, e.g., Csanady 1976). For a resistance coefficient $r = 1.6 \times 10^{-3}$ m s^{-1} ($F_{yb} = r\rho v$; Scott & Csanady 1976), the estimates of the mean flows are 0.01 and 0.03 m s^{-1} along-shore towards the south for the respective situations of a uniform flow with depth and a flow which reduces to 50% of its surface value just above the bottom.

The significant point is that the flow is towards the south, with the main balance between bottom friction and the along-shore slope of the sea surface. This direction agrees with the only available current-meter observations on the open west coast of the North Island. Over a 19-day sampling interval, with the current meter at a depth of 79 m—i.e., below the depth affected directly by the wind—at 40°12′S, 174°12′E (on the mooring from which the Reynolds Stresses were calculated), the mean velocity was 0.03 m s^{-1} at 253°T. For a period of 4 days within this 19-day record, which appeared to be little affected by meteorological forcing, the mean velocity was 0.10 m s^{-1} at 233°T.

South of 42°30′S. South of the Challenger Plateau the zonal geostrophic flow swings southward, giving rise to a strong flow on the west coast of the South Island south of Big Bay (Fig. 1). The sea-surface slope relative to 500 m changes from $+1.72 \times 10^{-7}$ (Stations 949, 953, 957, 802, and 813) at approximately 300 km offshore to -1.5×10^{-7} along the 600 m isobath close inshore. This change in sea-surface slope (-3.2×10^{-7}) is largely accounted for by the increase in flow towards the south, ($f/g \times \partial vP/\partial y$). With v increasing from 0.13 to 0.23 m s^{-1} over 2° of latitude, as indicated by Stanton (1976, fig. 3), taking P as 0.5°, the associated change in slope is -2.5×10^{-7}. This gives some confidence in the above sea-surface slope values being absolute.

On the continental shelf south of 42°30′S the sea surface appears to slope downward towards the north with a slope of about -1.5×10^{-7}. On the continental shelf a significant percentage of the vertical shear is due to the inshore dilution of oceanic waters (see p. 219). The shear is such as to increase the current flow towards the north with depth; the nett circulation in the absence of wind is envisaged as indicated in Fig. 4. The flow on the continental slope and outer continental shelf is that sweeping south from the southern flank of the Challenger Plateau. Inshore of this flow the sea surface slopes downward along-shore towards the north (Fig. 3). This slope produces an offshore flow, except close to the coast, where an along-shore flow towards the north will be produced. For a linear frictional stress law ($r = 1.6 \times 10^{-3}$ m s^{-1} in 50 m of water) the sea surface slope of -1.5×10^{-7} would be balanced by a near-bottom flow of 0.05 m s^{-1}. As a result of the increase in density in an offshore direction the northwards coastal flow intensifies with depth.

APPLICATION OF THE MODEL

I will now quantify the components resulting from each of the forcing-functions separately.

Wind forcing

On the west coast of the North Island north of Cape Egmont the most persistent wind is directed mainly onshore. For a typical wind speed of 5 m s^{-1} (~10 knots) the wind stress F_x^1 [$= 1.4 \times 10^{-3}$ (wind speed)^2] is 3.5 newton m^{-2} (0.35 dynes cm^{-2}), using a drag coefficient of 1.4×10^{-3} (see, e.g., Pond & Pickard 1978). From Equation (1) we see that when h is sufficiently large the sea surface slope is very small, and in this circumstance the flow consists of a surface Ekman spiral with the surface flow components u, v ($u + iv = F_x/fh_e\,(1-i)$, with $h_e = 1/k$ the Ekman depth) both of 0.025 m s^{-1}. The along-shore transport given by $F_x^1/f \times$ (shelf width) is 1.6×10^6 m^3 s^{-1} for a shelf width of 5×10^4 m.

On the west coast of the South Island the most persistent wind is directed mainly along-shore towards the north with a typical speed of 10 m s^{-1} (~20 knots) (Neale & Thompson 1978). From Equation (1), solving for the real and imaginary parts (with $\delta = S_y = 0$), we have

$$\zeta_x = \phi_y \frac{f}{|f|} \lambda \left(\frac{2\lambda|R| + 2 + 1/\lambda|R|}{1 + \lambda|R|} \right)$$

which, on substitution in Equations (2–4), gives the surface, mid-water, and bottom velocity components (Table 2).

This wind-derived flow has a strong along-shore component directed towards the north, with an offshore component near the surface and an onshore component near the sea floor.

Boundary forcing

The sea surface slope south of about 42°30′S on the west coast of the South Island has been estimated as $\partial\zeta/\partial y = -1.5 \times 10^{-7}$, whereas on the west coast of the North Island it is estimated as $\partial\zeta/\partial y = 6.2 \times 10^{-8}$. Observations of currents are not available to allow estimation of the Reynolds Stress off the South Island, but on the basis of North Island observations this stress is likely to change the sea surface slope by only 20%, and therefore will be neglected south of 42°30′S. I will also ignore the

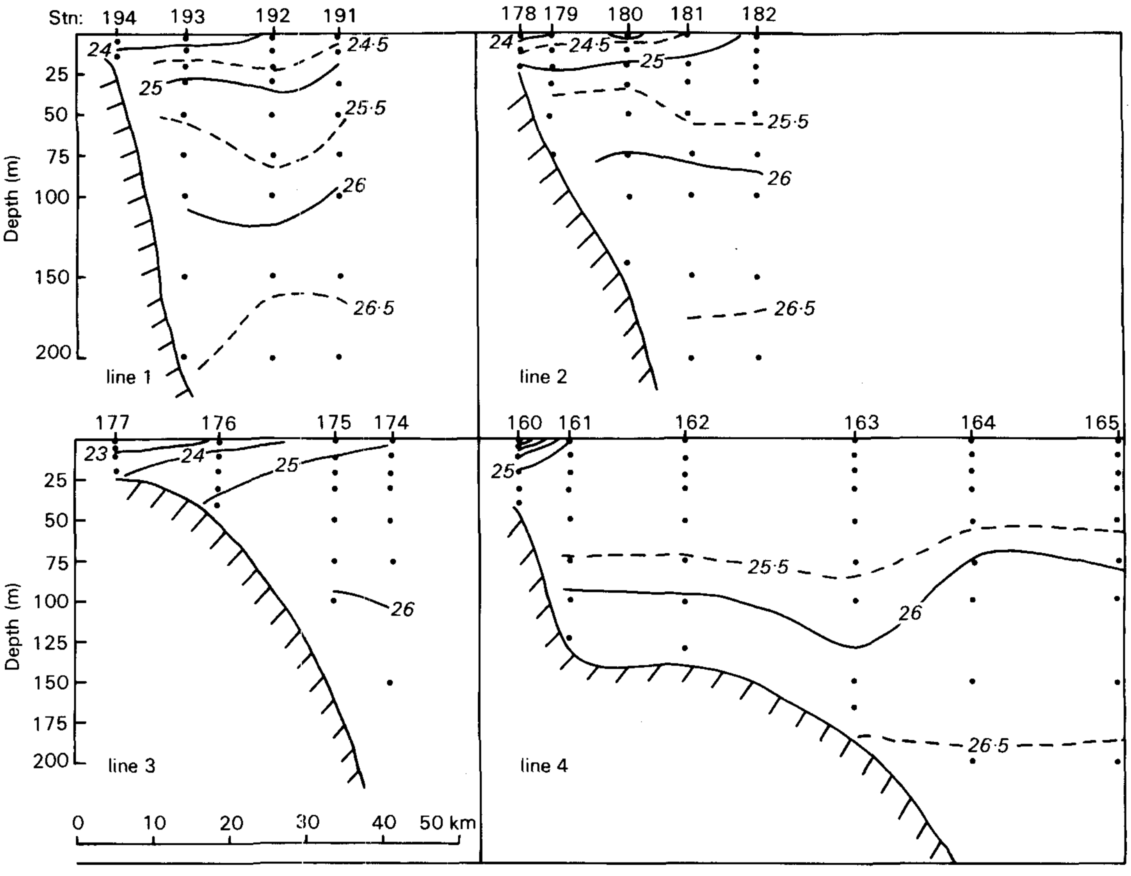

Fig. 5 Density plots across the continental shelf of the west coast of the South Island from data collected by Stanton (1976b); station positions shown in Fig. 1.

variation of the along-shore sea surface across the shelf which is needed to balance the latitudinal variation in the Coriolis acceleration.

From Equation (1), solving for the real and imaginary parts (with $\delta = \phi = 0$), we have

$$\zeta_x = \pm\zeta_y\left(\frac{2\lambda|R| + 2\lambda^2R^2 + 1 - \lambda R^2}{|R|(1+\lambda|R|)}\right)$$

which, on substitution in Equations (2–4), gives surface, mid-water, and bottom velocity components. These are listed in Table 2 for water depths of 50 m and 100 m, for $r = 1.6 \times 10^{-3}$ m s^{-1} and Ekman depth $(k^{-1}) = 16$ m.

North of about 42°30′S ($\partial\zeta/\partial y$ +ve), the along-shore component of flow is directed towards the south, with the surface and mid-water across-shelf components directed onshore and the bottom across-shelf flow directed offshore. South of about 42°30′S ($\partial\zeta/\partial y$ −ve) the along-shore component of this flow is directed towards the north, decreasing to 50% of its surface value near the bottom, with the surface and mid-water across-shelf components of the flow directed offshore and the bottom across-shelf flow directed onshore.

Freshwater forcing

From Equation (1), solving for the real and imaginary parts (with $\zeta y = \phi = 0$), we have

$$S_x = \delta\left(1 - \frac{1}{2\lambda} - \frac{|R|}{(1+\lambda|R|)^2}\right)$$

Plots of density (σ_T) across the west coast of the South Island continental shelf (Fig. 5), from data collected in April 1974, indicate that the slope of the isopycnals shows a marked spatial variability. In view of the large temporal variability in the freshwater inflow, there will be a large temporal

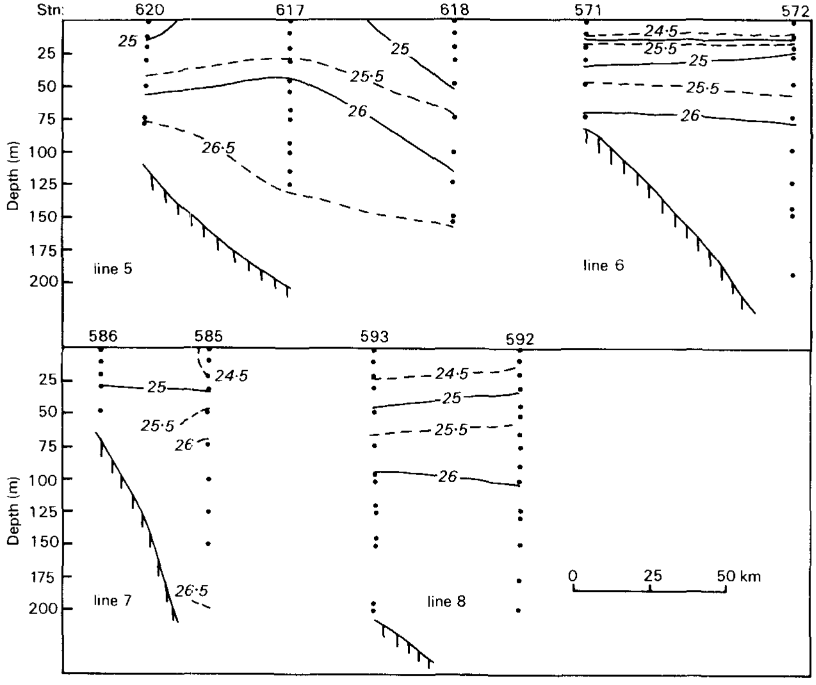

Fig. 6 Density (σ_T) plots across the continental shelf of the west coast of the North Island from data collected by Ridgway (1980); station positions shown in Fig. 1.

variability in the density distribution, some indication of which is given by Stanton (1979b) from sections of hydrographic data collected at different times near Cape Foulwind.

Calculations of the thermohaline velocities have been made (Table 2) for a value of $L = 5 \times 10^9$ m, which is taken from the observations at a depth of 25 m between Stations 174 and 177 (Fig. 5). The associated along-shore and onshore bottom flows are substantially less than those associated with the wind or sea-surface slope. However, the onshore flow near the surface is comparable in speed to that associated with the along-shore slope of the sea surface.

Off the west coast of the North Island the horizontal variations in density are small, and no distinct density pattern emerges (Fig. 6), a reflection of the smallness of the freshwater inflow relative to that on the west coast of the South Island.

DISCUSSION

From the model's results it appears that the circulation on the New Zealand west coast continental shelf is driven mainly by boundary forcing and by the local winds. South of about 42°30′S both the boundary forcing component of the flow and that due to the predominant north-eastward-directed winds reinforce each other, whereas north of about 42°30′S, off the South Island, these flow components are opposed.

Current-drogue observations on the continental shelf near Cape Foulwind (Fig. 1) under calm weather conditions indicated a mean southward drift (Heath 1973). This observation agrees with the results from the above model, the flow being that due to sea-surface slope forcing. Current-meter observations on the North Island west coast continental shelf north of Cape Egmont indicate a southward flow under calm conditions (Heath 1978) due to sea-surface slope forcing, which also agrees with the model.

Clearly, extensive observations of currents on the west coast shelf are needed to enable the mean flow to be better defined. The model indicates that as well as correlating such current observations directly with atmospheric conditions, an extensive analysis of sea-surface elevations would provide insight into both variations in forcing from the deep ocean and the local atmospherically induced flow.

Over the outer shelf north of about 42°30′S the near-bottom flow is directed southwards. This flow and the general oceanic southward flow would lead to the subsurface salinity maximum observed at a depth of about 100 m. Stanton (1971) attributed this salinity maximum to an undercurrent associated with wind-derived upwelling, but the present analysis confirms Stanton's later (1976b) study, in that the salinity maximum is a permanent feature. The prominence of the salinity maximum is enhanced by the presence of the near-surface salinity dilution.

The onshore-offshore flow also changes character near 42°30′S. South of 42°30′S the near-bottom flow has an onshore component; this is consistent with the rise in the subsurface isohalines near the coast evident in Stanton's data (Stanton 1976, fig. 7). North of 42°30′S the model predicts that the near-bottom across-shelf flow associated with the along-shore slope of the sea surface opposes the other across-shelf components. The depth at which the across-shelf bottom velocity component is zero is (from Equations (1) and (3)) given by

$$h = \frac{-1}{k} - \frac{F}{gS_y} + \frac{1}{2k^2LS_y}.$$

South of 42°30′S the across-shelf component is everywhere onshore. North of 42°30′S, the depth of zero across-shelf flow is 99 m with the wind forcing excluded, and with a 10 knot along-shore wind towards the north is 290 m. The across-shelf bottom flow component probably influences the dispersion of sediment. Carter's (1980) distributions of sediment textural types on the New Zealand west coast show mud predominating inshore of the 100 m isobath north of 42°S on the west coast of the South Island and a lack of mud on the west coast of the North Island. The reason for these distributions may be the zero across-shelf flow at about 100 m on the west coast of the South Island north of 42°30′S and the offshore bottom flow across the entire west coast North Island continental shelf. As indicated by Carter's (1980) analysis, however, there may be other reasons, such as the present versus relict source of supply and the influence of past stands of sea level.

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