

The Hydraulic Regime and its Potential to Transport Sediment on the Canterbury Continental Shelf

by

L. CARTER and R. H. HERZER



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FOREWORD

Early studies of continental shelf sediments were concerned primarily with the distribution of various textural and mineralogical parameters. More recently there has been increased interest in the transport processes affecting sediment distribution patterns, not only from a scientific viewpoint, but also from an applied viewpoint as marine geologists and engineers attempt to rationalise problems related to dispersal of pollutants, and erosion and deposition in the vicinity of man-made submarine structures.

Ideally, sediment transport is best approached through long term, *in situ* measurement of currents and their effects on the seabed. The instrumentation required for monitoring these effects is still in early stages of development and the logistics associated with instrument installation, maintenance, and retrieval are complex and expensive. In the absence of such measurements, the present study of transport on the Canterbury continental shelf has relied on a theoretical approach using, as a basis, available hydrological and meteorological data. The paper provides an insight into the processes affecting transport, the frequency and dominance of these processes, and their overall influence on the sediments. It is essentially a "state of the art" document pending the day when effective instrumental monitoring of shelf sedimentation can be made.

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The Hydraulic Regime and its Potential to Transport Sediment on the Canterbury Continental Shelf

by

L. Carter¹ and R. H. Herzer²

ABSTRACT

Sediments on the Canterbury continental shelf are subject to a complexity of water motions as inferred from a theoretical evaluation of the hydraulic regime.

In calm weather, typical of summer, the regime is dominated by tides, mean flow (Southland Current), and deep sea swell, which together periodically transport bedload on the inner continental shelf (<30 m deep) towards the north-east. Transport at greater depths is uncommon except on the tide-dominated shelf off Banks Peninsula and perhaps at the internal wave-influenced shelf edge.

In winter, gale and storm-induced motions strongly reinforce existing currents. The addition of storm waves and wind drift currents causes almost continual transport on the inner shelf and periodic transport at middle shelf depths, but their influence diminishes over the outer shelf, where the regime is dominated by storm-induced barotropic flow, mean flow, tides and, possibly, internal waves. Together these motions can probably shift sand.

The prevalence of southerly gales and storms produces a net north-eastward (along shore) transport of sand combined with an onshore component over the inner shelf and a slight offshore component elsewhere. These directions are substantiated by drift card and seabed drifter data.

INTRODUCTION

Even in recent years geologists have continued to invoke the coastal current system described by Brodie (1960) to explain sediment dispersal patterns on various sectors of the New Zealand continental shelf, e.g., Van der Linden (1969), Summerhayes (1969) and Schofield (1976). However, Brodie's work has little direct application to transport studies for it deals primarily with *the direction of the mean flow in surface waters*; speeds and directions of currents near the seabed are not discussed. Furthermore, more recent work has shown that the speeds of the mean flow are generally well below the threshold of sediment movement, and the direction of flow at the surface is not always the same as that at depth, e.g., Carter and Heath (1975), Stanton (1971). What then are the forces behind sediment transport? This question can best be answered through an examination of all currents and their potential to transport bedload.

Ideally, the individual and collective roles played by currents are best appreciated through long term, direct observation of sediment/water interaction. The technology involved in making such observations is still in

its infancy, so workers rely on indirect approaches based on interpretation of either hydrologic or sedimentologic parameters or both. In the case of the present study on the Canterbury continental shelf, a theoretical and practical evaluation of the hydraulic regime is used; the sedimentology is presented elsewhere by Herzer (1977).

In a regional survey of the New Zealand shelf, Carter and Heath (1975) briefly commented on the transport potential of waves, tides, and mean flow over the Canterbury-Otago shelf. Since that survey, new data, combined with an application of existing physical oceanographic theory, have permitted a more detailed analysis of water and sediment movements off Canterbury. The first part of this paper describes individual water motions with emphasis on their velocities, frequency of occurrence, and locality. The second part is concerned with the overall effect of these motions on sediments as predicted by theory and inferred from drifter experiments. These enable an evaluation of transport with respect to its causes, direction, frequency, and spatial variability.

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PHYSICAL SETTING

The Canterbury shelf, off the east coast of the South Island, is 305 km long and 40–85 km wide (Fig. 1). The shelf break lies between the 150 and 180 m isobaths. Apart from the deep incision made by Pegasus Canyon, the shelf surface has little relief; broad featureless areas are broken by small subdued banks and basins and, locally, by well developed ridges and swales.

The shelf is mantled mainly by modern, relict and

palimpsest terrigenous sediments (Herzer 1977). Modern sands are generally confined to the nearshore zone; locally they may extend on to the middle shelf. Modern muds occupy most of Pegasus Bay and a middle shelf position off southern Canterbury. Relict gravels and sands and their reworked equivalents are prominent over most of the shelf in the region of Banks Peninsula.

HYDRAULIC REGIME

OCEANIC CIRCULATION (MEAN FLOW)

The pattern of oceanic circulation in the region was described by Brodie (1960) from drift card data and has since been refined by Heath (1972c, 1973) using the geostrophic method and current drogue measurements. The Southland Current travels north-eastwards along the continental shelf and slope off the east coast of the South Island. Near Kaikoura it bifurcates, one component continuing north, parallel to the coast, and the other heading east to combine with the East Cape Current and flow towards the Chatham Islands. Surface geostrophic speed for the Southland Current was calculated by Heath (1972c) to be 7–8 cm s⁻¹. Drift card data compiled by Herzer in the course of the present study suggest a minimum speed at the surface of 13 cm s⁻¹ (see p. 24).

Drift cards released in the Southland Current off North Canterbury travelled north (Brodie 1960). In contrast, cards released within Pegasus Bay generally moved west on to the beaches, thereby giving credence to the notion that an anticlockwise gyre exists in the bay. Observations by Dawson (1954), however, suggested surface currents were largely influenced by local winds. Apparently the circulation within the bay is complex and variable. Two *Landsat* photographs display different flow patterns: Fig 2a shows a northward flow indicated by plumes of suspended sediment; Fig. 2b shows a confused pattern suggestive of an eddy with a northward flow off Banks Peninsula and north of the bay, and a diffuse southward flow along the south-western shore.

TIDES

The tides have major semi-diurnal components with a measured range at Lyttelton of 1.9 m springs and 1.6 m neaps and at Akaroa of 1.8 m springs and 1.6 m neaps (Hydrographic Office 1953b). The tidal wave advances in an anticlockwise direction around New Zealand and proceeds north-east along the east coast of the South Island (Bye and Heath 1975). The tide on

the continental shelf therefore floods to the north-east and ebbs to the south-west (Fig. 3). Tidal flow in Pegasus Bay is weak (Station B, Hydrographic Office 1953a; Fig. 4) whereas off Banks Peninsula the flow is strong, reaching a maximum speed of 31 cm s⁻¹ at the surface (Station A, Hydrographic Office 1953b; Fig. 4). The speed and marked dipolar character of the tides in this area are due to the local projection of the peninsula and a regional constriction of tides between the South Island and the Chatham Rise.

Parachute drogue measurements made by Herzer (1977) confirm the high speeds of tidal currents in the vicinity of Banks Peninsula (Fig. 4). They also indicate that the polarity of the tidal stream is less pronounced away from the peninsula. Drogues were set at 30 m and 60 m (in a total depth of 75 m) and tracked for 5 hours. The resultant maximum computed velocity was 41 cm s⁻¹ towards 180°T at both depths. That such high tidal speeds extend right across the shelf can be inferred from Heath (1973) who measured currents of predominantly tidal character with maximum speed of 44 cm s⁻¹ at 300 m depth (in 460 m of water) just beyond the shelf edge in Mernoo Saddle (Fig. 1).

WAVES

Data

Wave parameters were obtained from several unpublished and published sources. These parameters were measured using a variety of techniques and consequently data quality is inconsistent. The most reliable measurements were made on offshore oil rigs and covered a total of 112 days during which wind speed and direction, wind wave height, and swell height, period, and direction were monitored at least twice daily by a combination of instrumental and visual methods (Figs 4–6). Less comprehensive, but nonetheless useful data, were collected by the Timaru Harbour Board (Hydraulics Research Station 1970) who recorded swell height, period, and direction once a day for six months (Figs 4, 7, 8). Other data came from

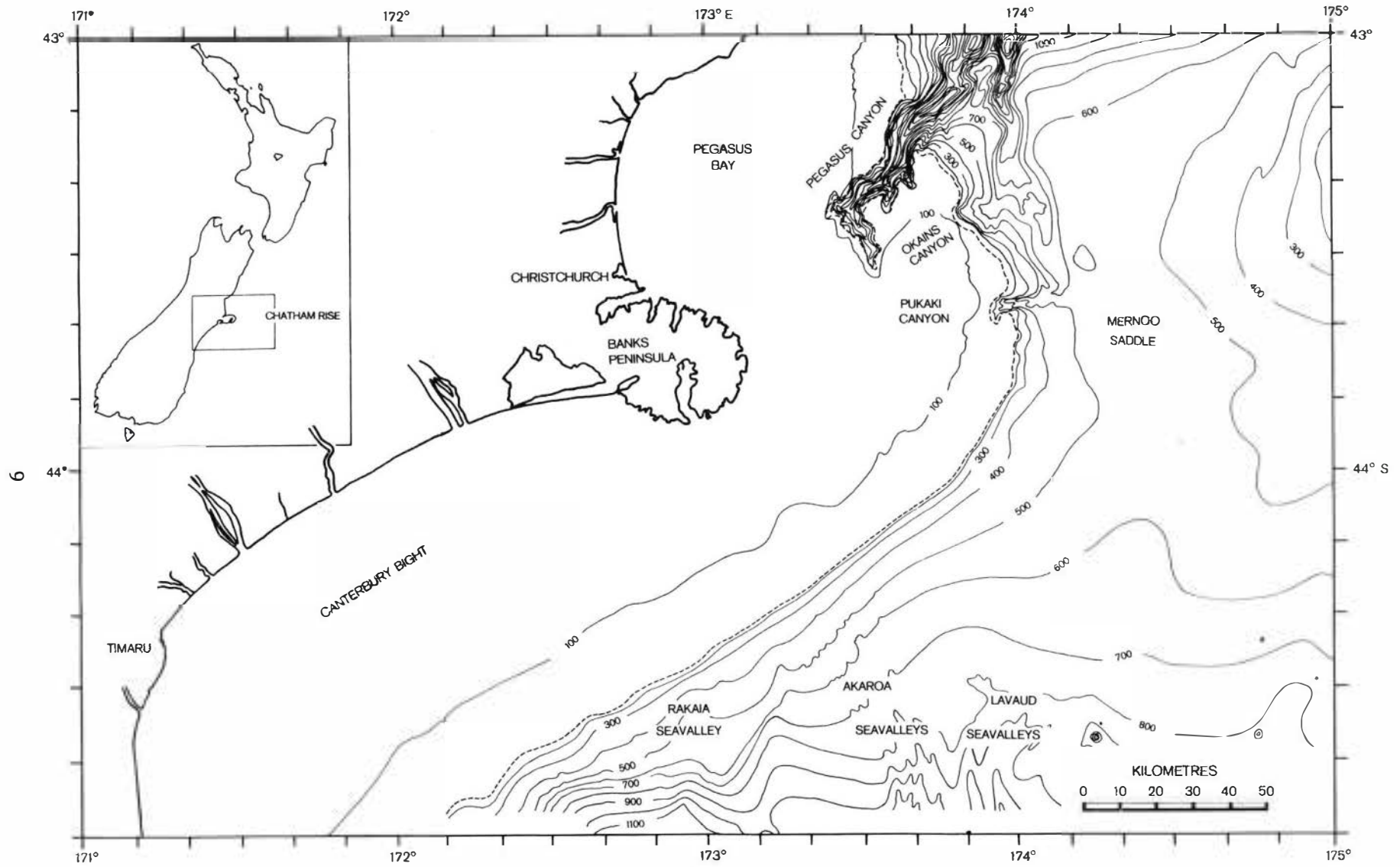


FIG. 1. Locality and bathymetric map *modified from Herzer (1977)*. Contours are in metres; dashed line is the shelf edge.



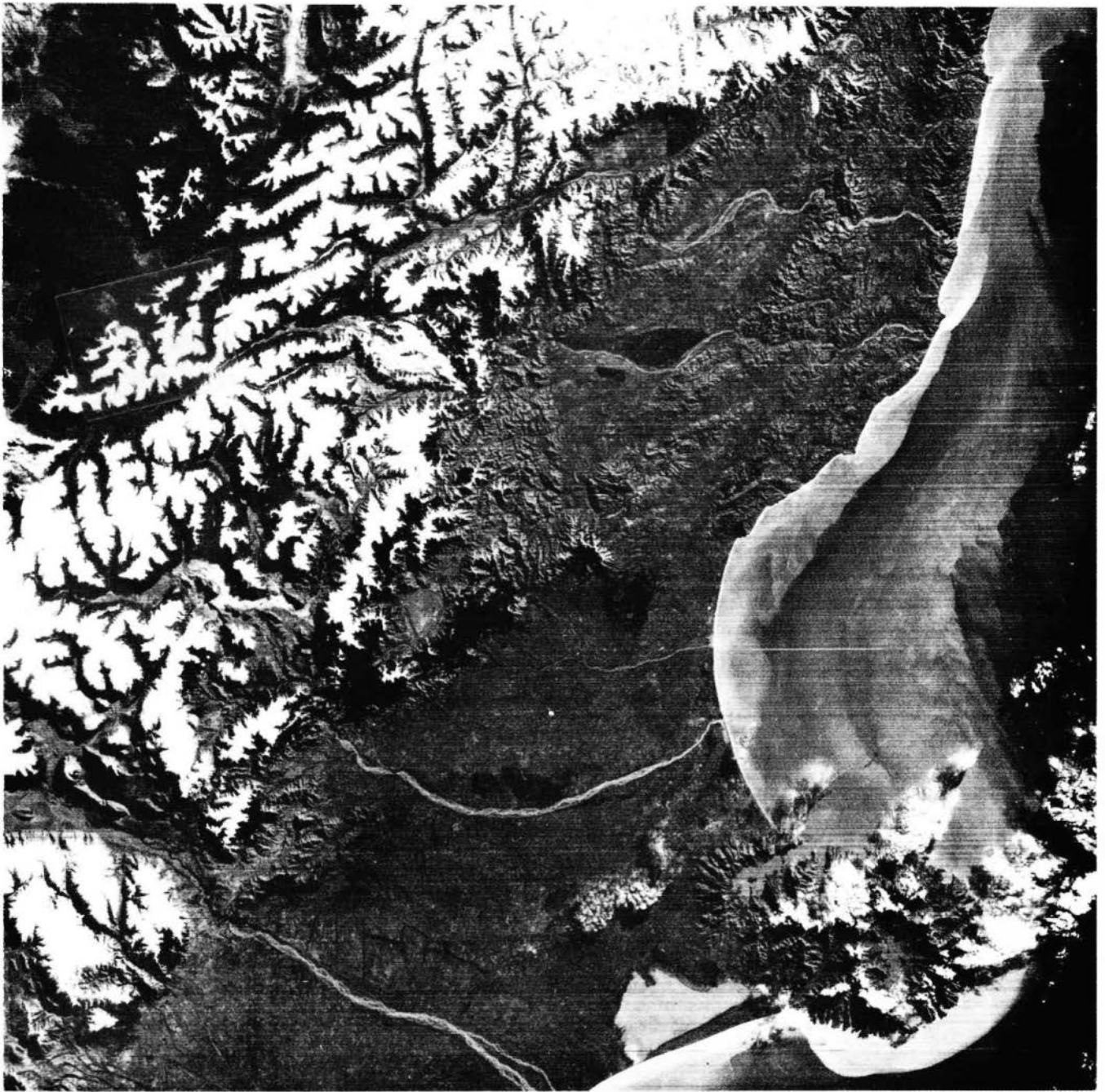


FIG. 2a. *Landsat II* image of Pegasus Bay and Canterbury Bight taken on 2 August 1975, Band 4; reference ERTS E-2192-21265-4. Photograph by courtesy of NASA.

shoreline observations of waves and swell by Burgess (1968) in Pegasus Bay, Dingwall (1974) off Banks Peninsula, Elliott (1958) on the Otago coast, Hodgson (1966) off Otago Pensinsula, and Kirk (1967) near Timaru.

Deep Sea Swell

Waves on the Canterbury shelf, as on most of the shelf off the South Island east coast, are a mixture of

deep sea swell derived from distant storm centres in the Southern Ocean (Snodgrass *et al.* 1966) and locally generated wind waves. Approximately 75% of all swell arrives from southerly (south-east to south-west) and northerly (north to north-east) directions with the remainder coming from the east (Fig. 9). The Timaru records show a strong easterly component, but in view of the proximity of the recording station to the shore, this may be due to refraction of south-easterly swell.

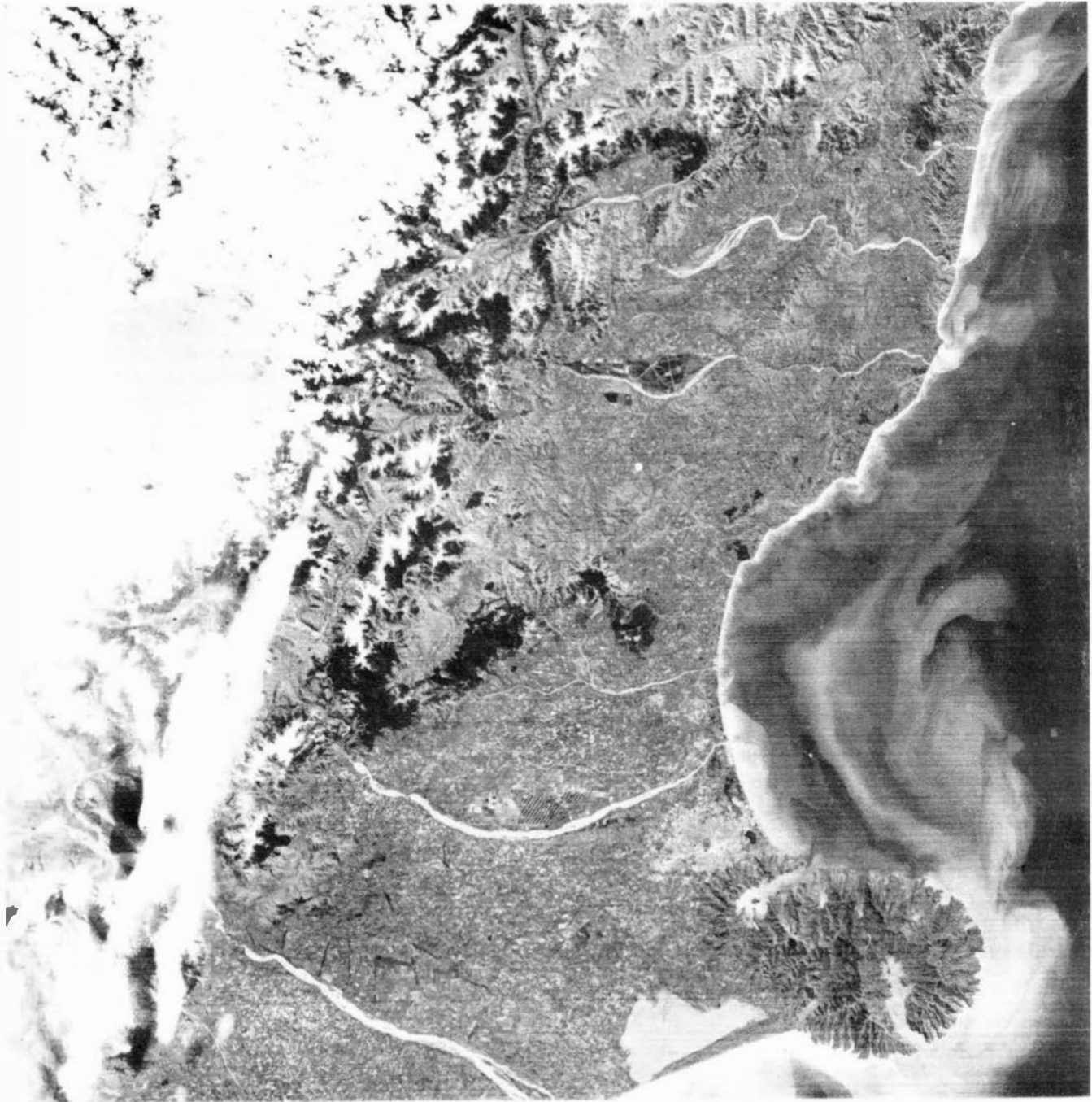


FIG. 2b. *Landsat II* image of Pegasus Bay taken on 31 October 1975, Band 4; reference ERTS E-2282-21254-4. Photograph by courtesy of NASA.

Typical southerly swell has long periods (10-11 s) and dominates the storm-driven wave climate of winter. Northerly swell usually has shorter periods (6-8 s) and only achieves prominence in calmer, summer months when it is as common as its southerly counterpart. Correlation of swell height with direction and seasonal changes is poor. Both northerly and southerly swell usually reach 1-2 m in winter or summer but, as a generalisation, highest waves are more common in the

winter and are mainly associated with southerly swell.

The open-coast wave climate is modified by Banks Peninsula. Southerly swell is refracted around the peninsula, thereby producing a distinct easterly component in Pegasus Bay (Dingwall 1974). This refraction is accompanied by a decrease in swell height (Burgess 1968).

The most noticeable effect of local winds on swell is to modify swell height, which tends to increase and

decrease with following and opposing winds respectively. For example, from 6 to 15 July 1975 an exceptionally high southerly swell off Banks Peninsula dropped a total of 5.5 m as wind speed decreased and wind direction gradually changed from south to north-east and north-west; with resumption of southerly winds, swell height increased (Fig. 6). Correlation between wind changes and fluctuations of swell period and direction are not so obvious. Winds with speeds less than about 20 km h⁻¹ have little noticeable effect, whereas gale and storm force winds may induce a realignment of the swell with wind direction and increase the period.

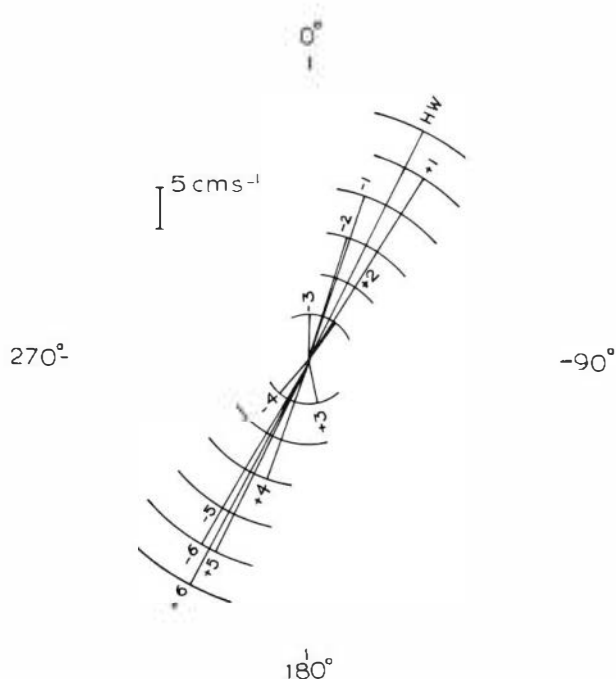


FIG. 3. Tidal current rose for reference station A (Hydrographic Office 1953b). Positive and negative values denote hours after and before high water respectively.

Wave Surge Speeds

When deep ocean swell extends on to the continental shelf it begins to "feel bottom" in water depths equivalent to about half a wavelength (Bascom 1964). As the wave progresses into shallow water μ , the horizontal component of orbital velocity, eventually exceeds the threshold of fine sand movement which, under the accelerating flow of surging waves, is about 10 cm s⁻¹ (Komar and Miller 1973). μ is determined from

$$\mu = \frac{\pi H}{T} \frac{1}{\sinh \frac{2\pi h}{L}}$$

where H = wave height (m), T = wave period (s), h = water depth (m) and $L = \frac{gT^2}{2\pi}$ which is then corrected for water depth using the tables of Wiegel (1948). μ , computed from oil rig and harbour board wave records, is expressed as a function of two arbitrary depths; 30 m, corresponding to the seaward limit of the inner shelf and 75 m, representing typical middle shelf depths (Figs 5–8).

The effect of wave surges on sediments at the aforementioned depths is poorly understood. Theory predicts a net mass transport of water (and by association, sediment) in the direction of wave propagation (Longuet-Higgins 1953; U.S. Army 1973). Yet, although these theoretical considerations have been substantiated in the laboratory (e.g., Russell and Osorio 1958), they are largely unconfirmed for the continental shelf (Bretschneider 1972). Even if transport is achieved, the associated speeds are very low, e.g., the theoretical speed of mass transport near the seabed for a 2 m high, 10s wave is ≤ 1 cm s⁻¹ at 30 m depth. In view of this we have taken a simplistic view; namely, that waves stir rather than transport bedload.

Stirring of sediment by swell is common on the inner shelf, becoming less frequent as deeper waters of the middle shelf are approached. The maximum depth of swell influence was estimated at 116 m. The maximum calculated surge speed for 75 m depth was 20 cm s⁻¹.

Stirring is most frequent in winter and early spring when southerly swell prevails. For example, off Banks Peninsula, 70% of sand-stirring swell came from the south-east to south-west (Fig. 9). By comparison, stirring is less frequent and less prolonged in late spring and summer.

The offshore wave records cover spring, summer, and winter. The autumn wave climate appears to resemble that of winter, as suggested by the shoreline observations of Burgess (1968) and Dingwall (1974).

Local Wind Waves

Superimposed on the swell are waves generated by local winds. The only data available apply to wave heights recorded from offshore oil rigs. (Figs 5, 6). Correlation of height with seasonal variations and wind direction is poor, with highest waves (2.5–3 m) occurring in all seasons and under both northerly and southerly winds. Because the rig records span only a short time, a more representative picture of the wind wave climate was obtained from hindcasts based on long term wind records.

The period and height of waves generated under winds of different speeds, durations, and fetches were estimated from prediction curves of Darbyshire and Draper (1963). Wind data came from wind persistence tables (New Zealand Meteorological Service 1976) which listed speed, direction, duration, and frequency

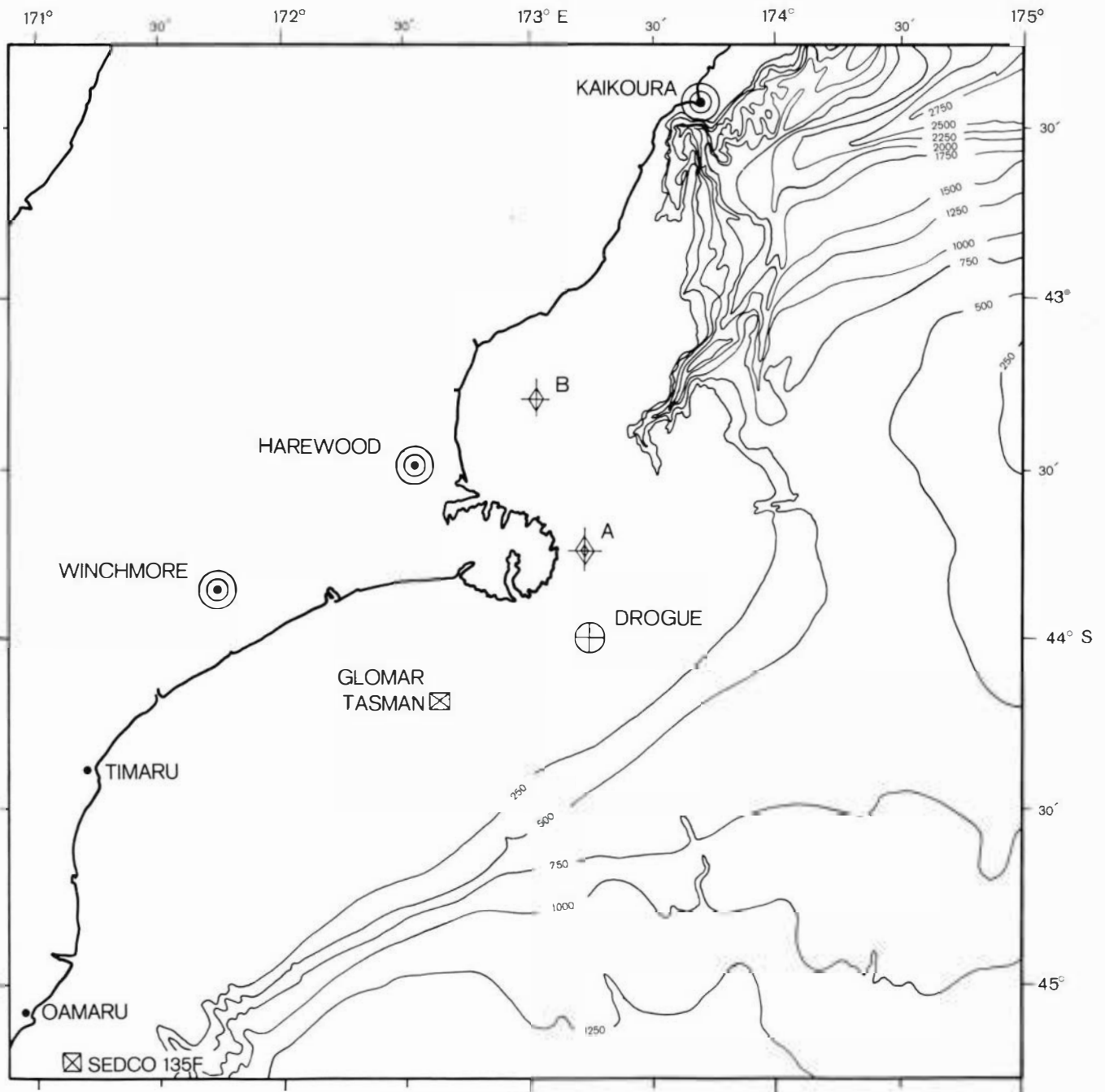


FIG. 4. Locations of offshore drilling rigs; Timaru wave recording site (Hydraulics Research Station 1970); tidal reference stations A, B (Hydrographic Office 1953a, b,); drogue station; and land-based meteorological stations.

statistics covering a period of 15 years. No specific data were available for fetch distances, and reliance was placed on synoptic weather maps, e.g., Garnier (1958).

It should be stressed that hindcasts are approximations because:

(i) they rely on data from land-based stations (Harewood, Winchmore, Fig. 4) which are generally subject to winds weaker than those at sea (Darbyshire

and Draper 1963; A. Neale, New Zealand Meteorological Service, pers. comm. 1976), and

(ii) predictions assume wave generation on an initially flat sea whereas on the Canterbury shelf waves are superimposed on an omnipresent swell. If anything the hindcast waves have underestimated dimensions, and consequently their calculated near-bottom speeds are conservative.

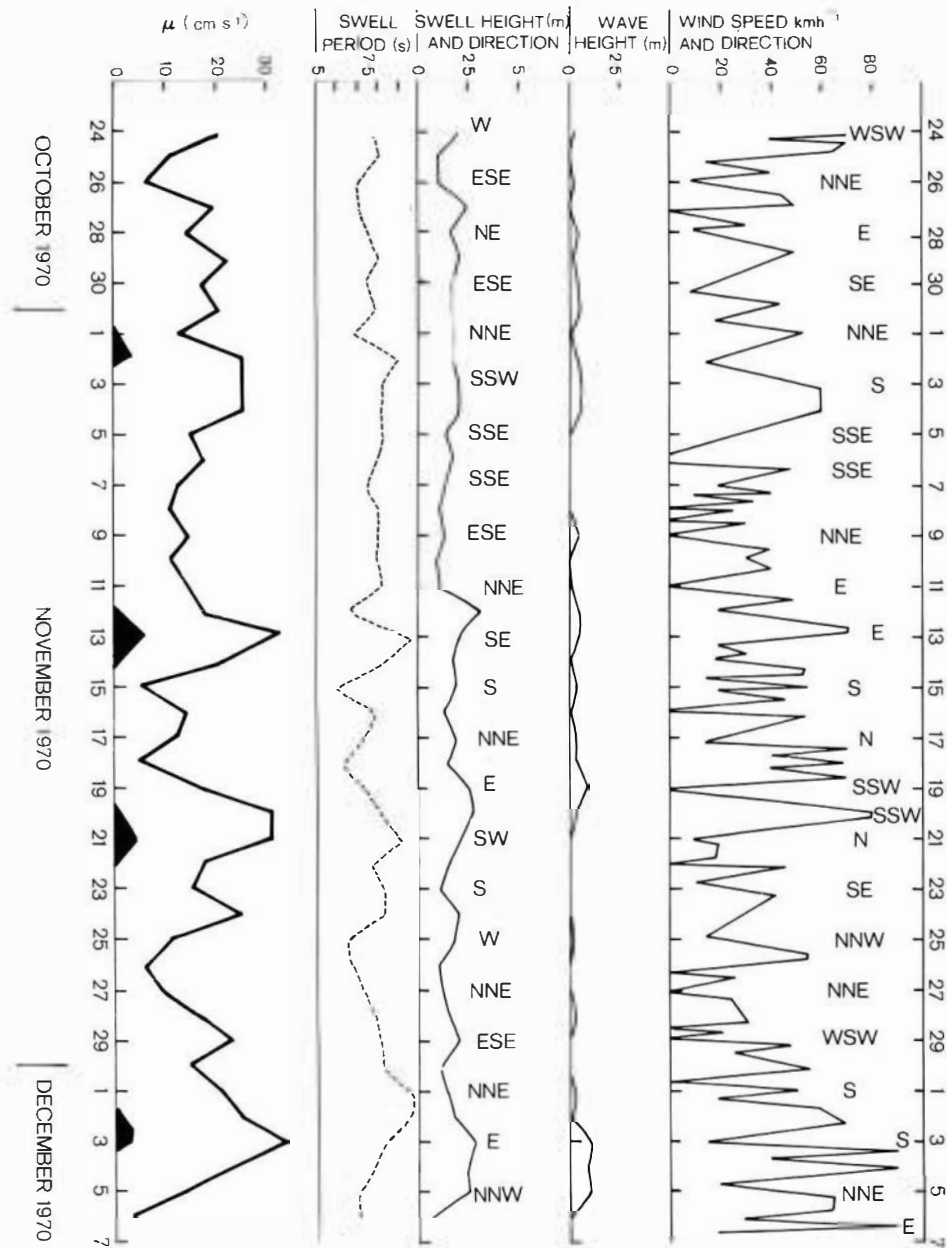


FIG. 5. Wind and wave data from *Sedco 135F* oil rig situated south-east of Oamaru at 45° 14.3'S 171° 07.5'E in 39 m water depth. Wind speed and direction monitored every 10 minutes and wave parameters every 4 hours. A waverider buoy was deployed to monitor swell height and period. When the buoy was inoperative the parameters were determined visually. μ is the horizontal component of orbital velocity near the seabed and is calculated for 30 m and 75 m (shaded area) depth.

The temporal and depth distribution of μ for the hindcast waves generally confirms the picture deduced from the oil rig records (Fig. 10). Sands on the inner shelf and shallow sectors of the middle shelf are periodically stirred by waves coming predominantly from the south, except in the summer months when waves of northerly origin are dominant. On the middle shelf the main force in spring and summer is northerly

waves which stir sands at depths less than 50 m. Autumn and winter waves are mainly from the south and move sand at all depths to 106 m.

STORM-INDUCED CURRENTS

Storms and other meteorological disturbances generate, in addition to the aforescribed waves and

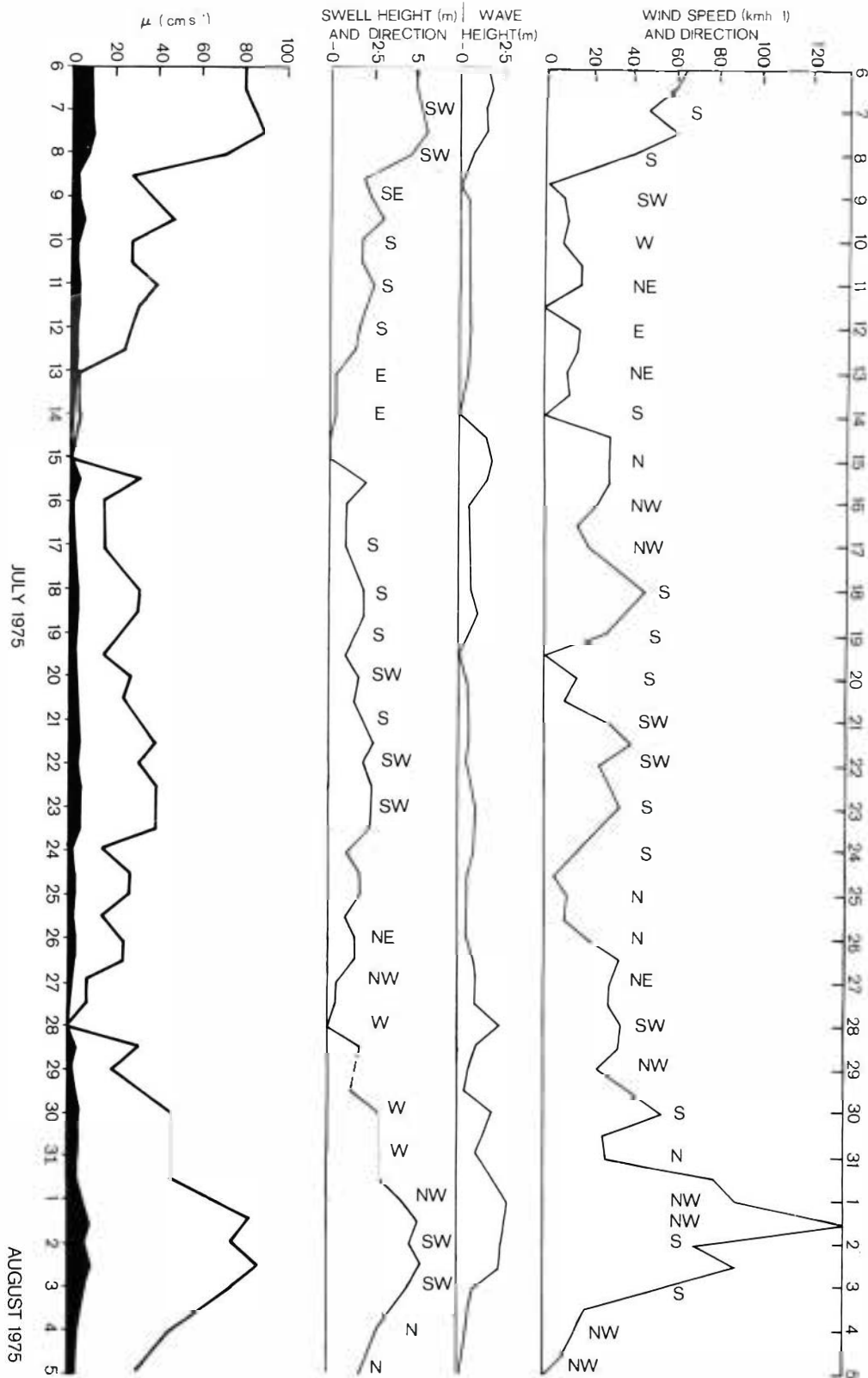


FIG. 6. Wind and wave data from *Glomar Tasman* drilling ship situated south of Banks Peninsula at 44° 11'S, 172° 38'E in 74 m water depth. Wind and wave measurements made once every 12 hours. No data are available for wave period, but to allow calculation of μ (the horizontal component of orbital velocity) a period of 9.4s is assumed, this value corresponding to the mean wave period recorded in July–August 1968 off Timaru. μ is calculated for 30 m and 75 m (shaded area) depth.

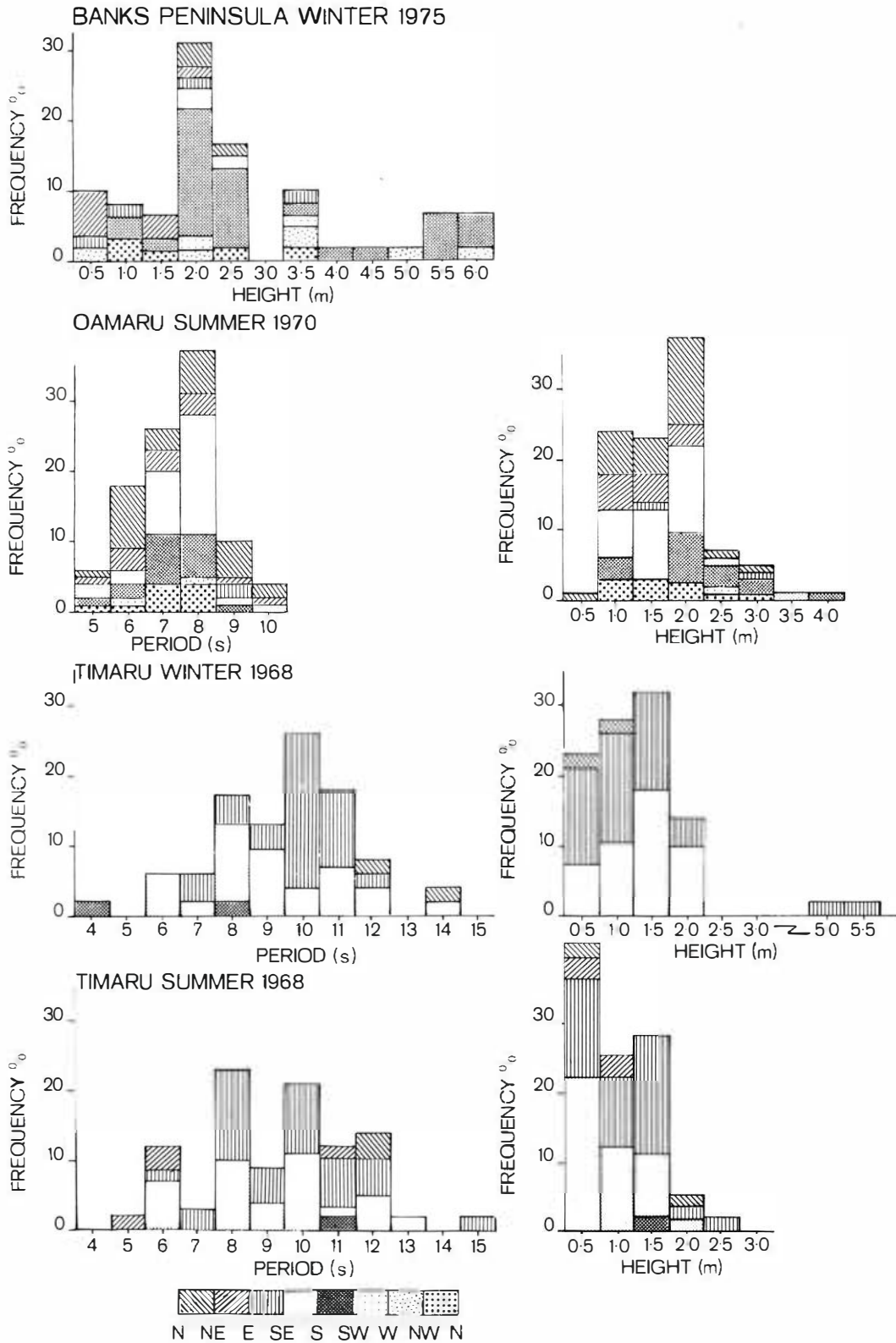


FIG. 9. Frequencies of swell direction, height and period broadly categorised into winter and summer climates. Data from *Glomar Tasman* (south of Banks Peninsula, July to August 1975), *Sedco 135F* (off Oamaru, October to December 1970) and the Timaru Harbour Board (July to December 1968).

swell, a variety of currents which at times may completely dominate the shelf hydraulic regime, e.g., see Sternberg and McManus (1972). Unfortunately, there has been no direct measurement of storm-induced currents on the Canterbury shelf. This has forced us to rely on theory.

Under ideal conditions wind drag on the ocean surface produces a drift current which flows approximately 45° to the left of the wind stress under the influence of the Coriolis force in the Southern Hemisphere. With increasing depth the current progressively deviates further to the left, thereby producing a spiralling velocity field or Ekman spiral in which the overall mass transport is directed 90° to the wind stress. Thus, on the Canterbury shelf, wind-drift transport is directed onshore during southerly winds and offshore under northerly winds.

The depth of frictional influence of the wind-drift layer is defined by the relationship

$$D = \Pi \sqrt{\frac{A}{\rho \omega \sin \phi}}$$

where A = wind-influenced, eddy viscosity coefficient (Neumann and Pierson 1966, fig. 8.19), ρ = seawater density, ω = earth's angular velocity and ϕ = geographical latitude. On the Canterbury shelf, where $\rho \approx 1.026 \times 10^3 \text{ kg m}^{-3}$ (Heath 1972 b) and $\phi = 44^\circ$, D is 144 m for gale-force winds (speed = 17.2 m s^{-1} ; $A = 110 \text{ kg m}^{-1} \text{ s}^{-1}$) and 218 m for storm-force winds (speed = 24.5 m s^{-1} ; $A = 250 \text{ kg m}^{-1} \text{ s}^{-1}$). Thus wind drift theoretically affects most shelf depths.

The speed of a wind-drift current at the surface may be expressed by

$$V_0 = \frac{\tau}{\sqrt{2A \rho \omega \sin \phi}}$$

τ , the horizontal wind stress, is given by $\tau = 2.6 \times 10^{-3} \rho^1 w^2$, where ρ^1 = air density and w = wind speed 10 m above sea level. For gale and storm-force winds V_0 is 29 cm s^{-1} and 41 cm s^{-1} respectively. These speeds decrease exponentially with depth as specified by

$$V = V_0 e^{-\frac{\pi Z}{D}}$$

in which V is the current speed at depth Z . Consequently, wind drift can be expected to have greatest influence on sediment transport over the inner shelf, becoming progressively less important over deeper sections of the shelf (Table 1).

Although wind drift at the surface is deflected 45° to the left of the wind stress, in shallow waters of the inner shelf the influence of the Coriolis force diminishes resulting in a decrease in the angle of deflection. For example, where the ratio of water depth to depth of frictional influence is 0.25, the angle of deflection is

21.5° (Neumann and Pierson 1966). The implication is that wind-drift and associated sediment transport are progressively directed along shore with decreasing water depth.

Other currents are set up in response to wind drift. When southerly winds prevail, the resultant onshore mass transport causes the sea surface to slope up against the coast and form an onshore pressure gradient. As a consequence, a north-trending barotropic current develops, and this is superimposed on a downwelling system formed by the onshore movement of surface waters in the wind drift layer and the compensatory offshore movement of bottom waters. The reverse situation holds for northerly winds. Surface waters are blown offshore and are compensated by an onshore movement and eventual upwelling of bottom waters. Offshore mass transport also depresses the sea surface near the coast. An offshore pressure gradient is formed together with a south-trending barotropic current. Examples of upwelling on the Canterbury coast have been observed by Heath (1972a, b).

In summer, when northerly winds prevail, the upwelling current system may be modified by formation of a sharp pycnocline or density stratification; the winter pycnocline is diffuse judging by temperature and salinity profiles of Heath (1972a, b) and Bradford (1972). Upwelling is confined to waters above the pycnocline; the south-flowing barotropic current occupies normal depths, whereas near the seabed a north-flowing baroclinic current is eventually set up (Yoshida 1967; Smith 1968; Smith and Hopkins 1972).

Compared to currents in the wind-drift layer, the underlying currents are slower. Bottom currents associated with upwelling have horizontal speeds of the order of 1 to 10 cm s^{-1} (Smith 1968) or about one tenth of the speed of the surface current (Yoshida 1967), i.e., 3 and 4 cm s^{-1} for gale and storm-induced motions respectively. Downwelling currents presumably have similar speeds. The speed of the barotropic current may be estimated from

$$\mu = \frac{g\beta}{2 \omega \sin \phi}$$

where g = acceleration of gravity and β = slope of the sea surface derived from

$$\beta = \frac{\lambda \tau}{g\rho d}$$

λ is a coefficient ≈ 1 for the situation near the shelf edge. Typical speeds for gale and storm conditions are $\sim 6 \text{ cm s}^{-1}$ and $\sim 10 \text{ cm s}^{-1}$ respectively.

INTERNAL WAVES

Internal waves are commonly generated along the ocean thermocline. As these waves meet the shelf they

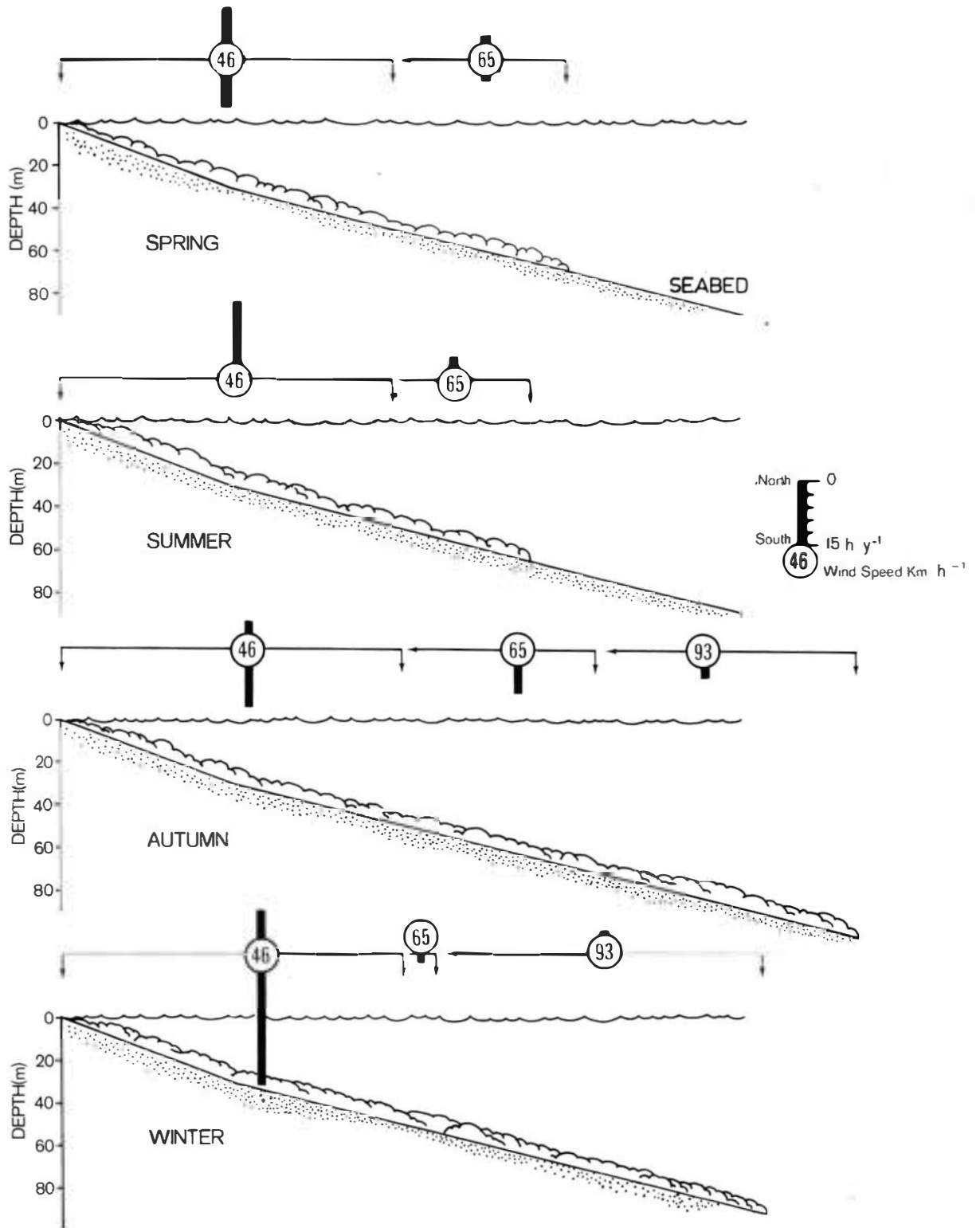


FIG. 10. Effects on bottom sediments of local wind waves hindcast from wind persistence records covering 15 years (N.Z. Meteorological Service 1976). Winds are categorised according to their speed, dominant direction and duration; sediment-stirring waves generated by such winds are depicted as clouds of suspended sediment within a given depth range. For example, in autumn winds of 65 km h⁻¹ blow mainly from the south for a total of 0.25 days annually, during which waves capable of moving sand at 70 m depth and shallower, are generated.

TABLE 1. Summary of water motions on the Canterbury shelf giving their direction (in the case of waves, propagation) and speeds; the latter as a function of various water depths. Values are derived from the following sources: (a) surface drifters, this study; (b) Heath 1972c; (c) seabed drifters, this study; (d) Hydrographic Office 1953b; (e) Heath 1973; (f) N.Z. Meteorological Service 1976; (g) wave records, *Sedco 135F*; (h) Hydraulics Research Station 1970; (i) wave records, *Glomar Tasman*.

Hydrologic Component	Locality of Measurement	Meteorological Forces	Seasonal Variations	Dominant Direction of Travel	Maximum Current Speed (cm s ⁻¹)	Depth of Occurrence (m)
Mean flow (Southland Current)	Canterbury			NE	(a) 13 (b) 7-8 (c) 2.5	Surface Surface Seabed
Tides	East of Banks Peninsula Mernoo Saddle			NE-SW	(d) 31 (e) 44	Surface 300
Wind waves	Hindcasts based on Christchurch Airport wind records	Gale to storm force winds	Spring Summer Autumn Winter	NE SW NE NE	(f) 190 34 76 6 85 7 160 23	30 75 30 75 30 75 30 75
Swell	Oamaru Timaru Banks Peninsula		Spring-Summer Winter-Spring Winter	NW-NE NW NE	(g) 34 6 (h) 100 18 (i) 84 10	30 75 30 75 30 75
Wind drift	Calculated using Christchurch Airport wind records	Gale to storm force winds	Most common in late autumn, winter, early spring	NE and NW	(f) 29 14 5 41 27 15	Surface 30 75 Surface 30 75 200
Wind-induced barotropic flow	Calculated using Christchurch Airport wind records	Gale to storm force winds	Most common in late autumn, winter, early spring	NE	(f) 6-10	200
Undercurrents associated with upwelling or downwelling	Calculated using Christchurch Airport wind records	Gale to storm force winds	Downwelling mainly autumn-winter; Upwelling mainly spring-summer	South (upwelling) North (downwelling)	(f) 3-4	30-200

NOTE: Upwelling may be confined to thermocline and above.

may suspend and transport sediment (Cartwright 1959; Stride and Tucker 1960; LaFond 1961). Although no direct observations of internal waves interacting with bottom sediment have been made, experiments of Southard and Cacchione (1972) suggest the breaking of these waves against the shelf edge may be an important mechanism in producing a net transport of sediment off the shelf and down the continental slope.

Internal waves of varying magnitude have been detected in echograms of the deep scattering layer near Mernoo Saddle (Fig. 11). As the ship was steaming during the echosounding the size and period of the waves could not be measured with accuracy and were

not submitted to spectral analysis. Simple inspection reveals the presence of a series of small, short-period waves superimposed on a longer wave that may be tidal in character. The smallest waves have periods of about 30-60 s and therefore will easily feel the edge of the continental shelf. An alternative explanation, that the oscillations are not internal waves but vertical migrations of organisms in the deep scattering layer, is unlikely. First, such migrations tend to be longer and less frequent than the echogram oscillations and second, migrations usually occur at dusk or dawn (Barham 1966) - the echogram was made between 1420 and 1520 h.

HYDRAULIC REGIME AND SEDIMENT TRANSPORT

The combined effects of the various shelf currents on bedload are examined in the context of the transport potential of the currents, directions of transport, and frequency of transport.

TRANSPORT POTENTIAL

The potential of shelf currents to transport bedload may be gauged from their speeds, which are discussed here in relation to the important influences of water depth and weather (Table 1). Fine sand moves in a linear flow of 35 cm s^{-1} , 1 m above the seabed (Sternberg and McManus 1972); under waves the threshold speed is about 10 cm s^{-1} (Komar and Miller 1973). The calm weather current regime on the Canterbury shelf consists of tides, the Southland Current, and swell. Inclusion of the last may seem unusual, but swell generated at distant storm centres is omnipresent, irrespective of the weather. For example, between 19 and 23 August 1968, off Timaru, swell height increased from 1 to 6 m even though wind speeds remained between 0 and 10 km h^{-1} (Fig. 7). On the open shelf the speeds of the tides and Southland Current are too low, even when combined, to instigate transport. The exception is off Banks Peninsula, where constriction of the combined flow increases its speed beyond the threshold of sand movement – the speed being highest when the Southland Current flows with the appropriate tidal current.

Outside the zone of constricted flow, transport relies heavily on calm weather swell, the main role of which is to stir bedload on the inner shelf and occasionally on the middle shelf. Once in suspension it is assumed that the sediment is readily transported by the tides and Southland Current.

The passage of meteorological disturbances enhances the transport competency and depth of influence of shelf currents through reinforcement by storm-induced currents. Sediments on the inner shelf are more or less continuously stirred by waves consisting of swell reinforced by local wind waves. The suspended sediment is then susceptible to transport by tides, the Southland Current and, in particular, wind-drift currents. The last of these, when driven by storm-force winds (*see* p. 19) are the fastest linear currents on this section of the shelf, having sufficient speed to transport sand by themselves.

On the middle shelf, transport under the combined influence of wave surges and linear currents becomes less frequent with increasing depth, there being no significant wave surges deeper than 116 m. With the exception of storm-driven wind drift, which is effective down to 30 m, individual linear currents are too slow to induce transport. However, when travelling in the same direction, these currents may collectively be an

important transporting agent. For example, the Southland Current and the north-east-flowing mid-flood to mid-ebb tidal current, reinforced by north-trending wind-drift and associated barotropic flow, may transport bedload at most middle shelf depths during storm conditions.

On the outer shelf, current velocities cannot be determined at present because of a paucity of data. Wave records from the oil rigs indicate the swell is too small to generate sediment-stirring surges on the outer shelf. This situation may reflect the short duration of the records. Swell of sufficient magnitude to generate such surges occurs on the outer shelves of Great Britain (Hadley 1964; Draper 1967), Gulf of Mexico (Curry 1960) and Western Canada (Carter 1973). But because this type of large swell is infrequent, its presence or absence here can only be substantiated by long term wave or meteorological records.

There is evidence of internal waves in the region (Fig. 11), and these may be an important instigator of transport near the shelf edge (Southard and Cacchione, 1972). Linear currents are weak except in the tide-constricted zone between Banks Peninsula and Chatham Rise. The influence of wind drift is negligible on the outer shelf, the main currents being (in order of probable importance) storm-induced barotropic flow, tides, and Southland Current, which together may produce a flow with a speed approaching threshold values.

TRANSPORT FREQUENCY AND DURATION

Calm weather, including periods when wind speeds are less than 5 m s^{-1} (10 knots), occurs for approximately 71% of a year in the Canterbury area (New Zealand Meteorological Service 1976). In these conditions the frequency of transport is governed by the frequency of sediment-stirring swell, as the linear currents are generally weak. Calm-weather swell capable of stirring inner shelf sediment most commonly occurs in winter and early spring. For example, it occurred 13 times in the period from July to October off Timaru, but only 3 times from November to December (Figs 7–8).

Calm weather transport is likely to be more frequent off Banks Peninsula, where the constricted tidal flow, assisted by the Southland Current, probably shifts sediment daily.

A realistic appraisal of transport under gale and storm conditions is hampered by a lack of pertinent data. Meteorological observations covering 1960 to 1972 report the occurrence of gales (defined by wind speeds $> 62 \text{ km h}^{-1}$) as 2.7 d y^{-1} for Christchurch Airport and 10.2 d y^{-1} for Akaroa on Banks Peninsula (N.Z. Meteorological Service 1961–1973). The latter is

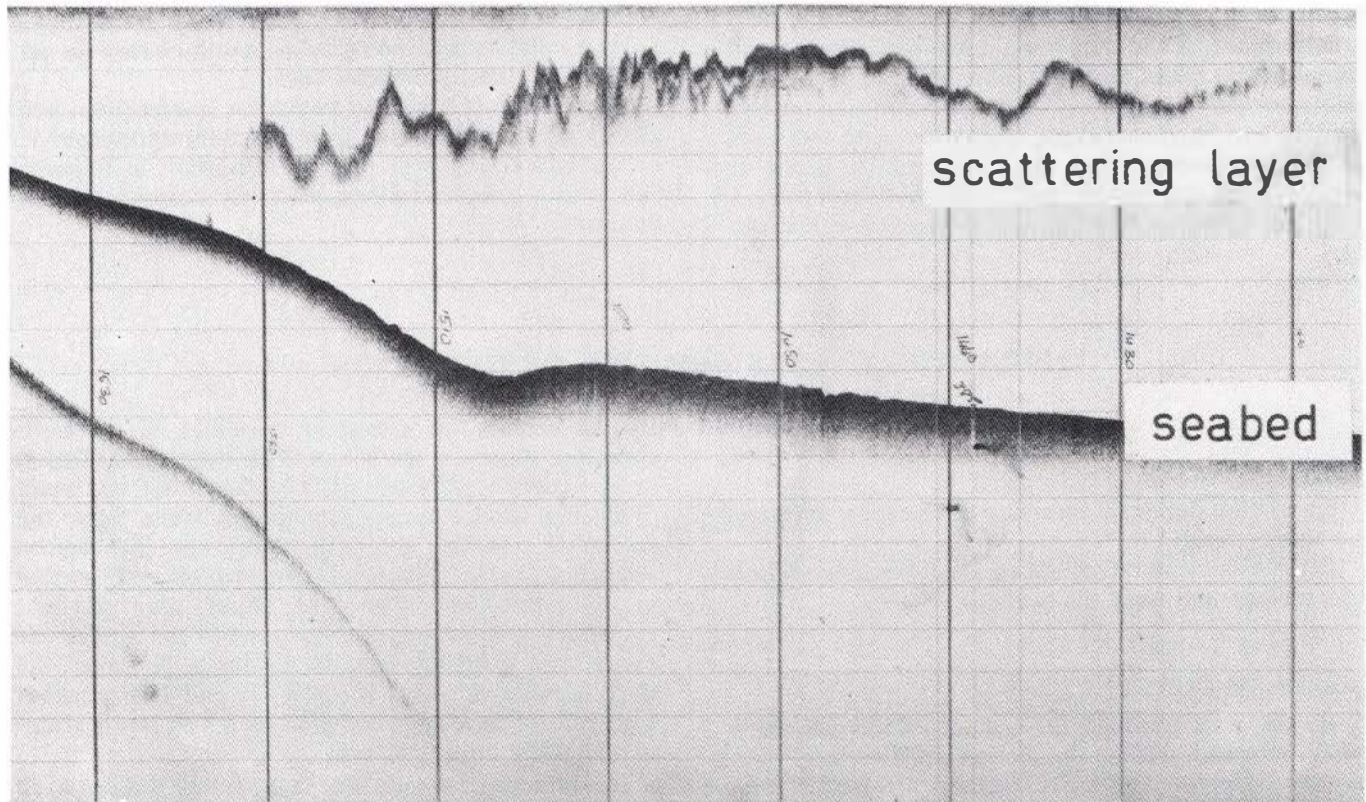


FIG. 11. Waves on the deep scattering layer observed in the Mernoo Saddle south-west of Mernoo Bank on 28 September 1973, NZOI cruise 1011. Horizontal scale lines are 36.6 m (120 ft) apart.

probably more representative of the shelf as Akaroa is closer to the sea than Christchurch. Although specified in days per year, the values are not a measure of gale duration but are the occurrence of gale-force winds recorded at 0900h each day. The actual duration of gale-force winds as indicated by wind persistence figures (available for Christchurch Airport only; N.Z. Meteorological Service 1976) is only 9 h y^{-1} . The implication is that the 10.2 gales are short-lived events in which winds only periodically and briefly reach gale force, e.g., wind records made offshore of Oamaru (Fig. 5).

Storms (defined by wind speeds $>89 \text{ km h}^{-1}$) are apparently rare events, being approximately one fourteenth as frequent as gales, according to wind gust data for 1942 to 1970 (N.Z. Meteorological Service 1973). The Canterbury shelf can therefore expect a storm about every 1.5 years. The *Glomar Tasman* drilling ship, situated south-east of Banks Peninsula, recorded such an event between 31 July and 3 August 1975 when north-west winds reached 140 km h^{-1} (Fig. 6). Winds were above gale force for almost three days and storm force for over six hours. Storms of this type and magnitude are infrequent; a study of the meteorological circumstances giving rise to the storm indicate they may occur once every 200 to 400 years (S. Reid, N.Z. Meteorological Service, pers. comm.).

As a generalisation, most gales and storms occur between late autumn and early spring. They tend to be more severe and of longer duration than summer disturbances and thereby have greater influence on sediment transport.

DIRECTIONS OF TRANSPORT

In calm weather, sediments stirred by swell are transported north-eastwards by the Southland Current combined with the north-east-flowing tidal current of the mid-flood to mid-ebb phase. When the tide reverses and is contrary to the mean flow, transport is unlikely to occur. Off Banks Peninsula the tides are sufficiently powerful to transport bedload to the north-east and south-west (Fig. 3). However, superimposition of the Southland Current produces a net movement to the north-east.

Storms are mainly from the south and produce an overall north-eastwards transport of sediment. Deviations from this trend occur on the inner and middle shelf, where there is an additional onshore component induced by wind drift, and on the outer shelf where sediment may shift offshore in response to

- (i) bottom currents associated with downwelling,
- (ii) interception of transport routes by submarine canyons, and,

(iii) breaking internal waves.

In summer, if the prevailing north-east winds reach gale or storm force, they may generate significant wind drift currents which, aided by the appropriate phase of the tide and by wave surges, shift inner shelf sediment to the south-west. This dispersal trend is probably negligible on the middle and outer shelf because

(i) the influence of wind drift decreases with depth

and may also be restricted by the summer pycnocline, (ii) a north-east-trending baroclinic flow may be set up beneath the pycnocline, and

(iii) north-east gales and storms are less frequent and generally less intense than their southerly counterparts.

Southward transport may be important in Pegasus Bay, which is sheltered from southerly storms by Banks Peninsula.

PRACTICAL MEASUREMENT OF CURRENTS

A programme of drift card and seabed drifter releases was carried out on the Canterbury shelf between 1973 and 1975 in order to

(i) test the dominant direction of transport predicted by theory, and

(ii) obtain estimates of the speed of the mean flow at the surface and near the seabed.

SURFACE CURRENTS

In May and October 1973, a total 2136 drift cards were released during the course of two geological cruises (Herzer 1977). A further 450 cards were released in July 1975 from the oil rig, *Glomar Tasman*, south of Banks Peninsula. Vectors between release and recovery points are shown in Fig. 12.

The pattern of drift that emerges corresponds closely to that recorded by Brodie (1960). As a rule, cards released on the continental shelf were ultimately washed ashore whereas most of those beyond or near the shelf edge were lost to the coastal system, several reaching Chatham Island. With few exceptions the cards were carried north-east; of those released in Pegasus Bay, most moved west on to the beach.

As significant minimum velocities were obtained for the northbound cards, viz., 18.3 cm s^{-1} (Table 2), the influence of the wind on the drift card tracks was assessed. Wind vectors, derived from hourly wind data at three New Zealand Meteorological Service weather stations in southern Marlborough, central and south Canterbury (Fig. 4), were examined for the period when the fastest ten cards of the present survey were in the water (Fig. 13 a, b, c). Wind speeds were cumulated within each separate octant of the compass and the total wind run for each octant was plotted as a vector. The resultant wind-run vectors represent the directions which could have influenced the drift of the cards. Allowance is made for up to 45° deflection to the left of the wind stress caused by the Coriolis force. Because a drift card floats in the top centimetre of water where response to wind shear is rapid, it is felt that this simplistic approach is justified.

The wind-run vectors, even allowing for the maximum possible Coriolis deflection, ran strongly counter to, or obliquely counter to, the direction taken

by the cards. The minimum velocities of the cards therefore represent the speed of the Southland Current at the surface without reinforcement from the wind. With the wind running against the mean flow, the minimum velocities are, if anything, low. The highest velocities (Table 2) resulted from card trajectories that ended near Cape Campbell. Since part of their trajectory was outside the study area their speeds are not, strictly speaking, valid for this study. However, the third highest velocity, 12.9 cm s^{-1} , and several other high values, were obtained entirely within the confines of the study area (Table 2).

It is important to note that of the fastest ten cards, all but one were released south of Banks Peninsula. The fastest card in Pegasus Bay was released in the south-east sector where it became entrained by the Southland Current. The minimum speed of this card was significantly higher than those obtained for the fastest cards within the bay (Table 3).

Anomalous drift patterns were individually investigated in the light of available wind data. The vectors and arrival times of the west-travelling cards in Pegasus Bay do not suggest an obvious gyre. Wind vector diagrams (based on data from Harewood, the closest weather station; N.Z. Meteorological Service 1976 unpublished records) for the fastest card from each of the two west-going groups offshore (Fig. 13b) reveal a significant westward wind component was active at the time. Cards released along the western side of Pegasus Bay, less than 3 km from shore, moved onshore within hours along westerly to south-south-westerly vectors. Recovery rates were 70–90%. The wind recorded at the time of release was blowing 13 km h^{-1} from 050°T , suggesting the trajectory of the cards was also influenced by wind.

Both the drift card and *Landsat* data complement Brodie's (1960) and Dawson's (1954) observations of the circulation in Pegasus Bay. The Southland Current flows past the mouth of the bay with a minimum velocity of 13 cm s^{-1} while net circulation within the bay remains weak. Although an eddy system may be set up, it is not permanent, and the slow-moving surface water in the bay shows a significant response to local wind patterns.

An interesting situation developed with cards from

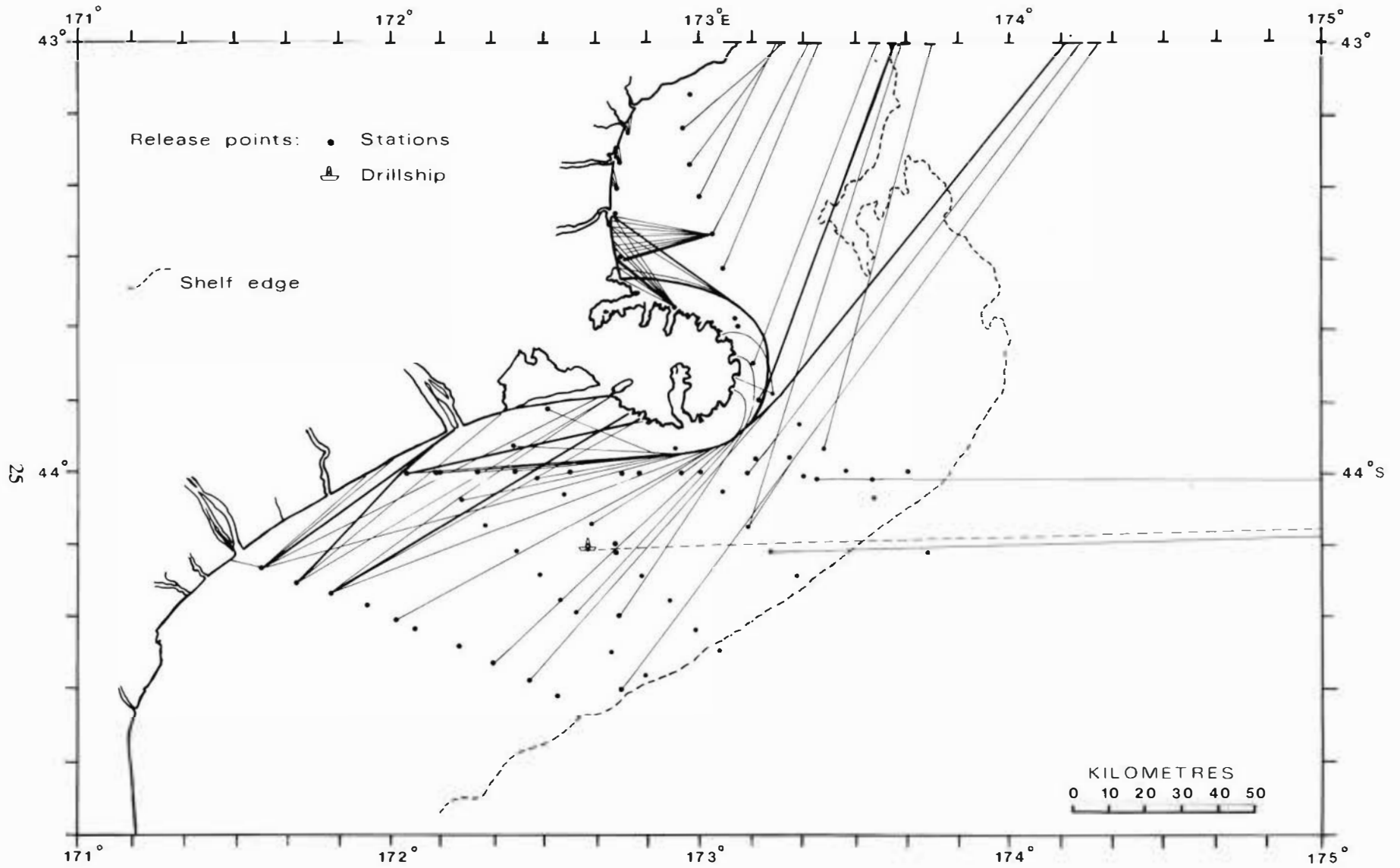


FIG. 12. Vectors between release and recovery points of drift cards. Thin lines represent the vectors of one or two drift cards; thick lines represent the dominant drift vectors; dashed line is the card trajectory from the drillship *Glomar Tasman*.

TABLE 2. Minimum speeds of 10 fastest drift cards.

NZOI Stn No.	Release Point		NZOI Card No.	Minimum Velocity (cm s ⁻¹)	Release Area	Direction of Travel
	Latitude S	Longitude E				
H367	44° 07.1'	172° 38.8'	N1803	9.4	Canterbury Bight	North-east
J404	43° 31.8'	173° 04.0'	I1214	10.3	Pegasus Bay	North-east
J414	44° 00.0'	172° 59.0'	U1343	11.6	Canterbury Bight	North-east
J416	44° 00.5'	172° 45.0'	U1393	14.4	Canterbury Bight	North-east
J427	44° 20.5'	172° 01.0'	U1842	12.9	Canterbury Bight	North-east
J430	44° 15.0'	171° 42.0'	U1957	10.7	Canterbury Bight	North-east
J430	44° 15.0'	171° 42.0'	U1990	8.7	Canterbury Bight	North-east
J431	44° 13.0'	171° 35.5'	X1006	9.9	Canterbury Bight	North-east
J434	44° 00.0'	172° 10.0'	X1079	9.0	Canterbury Bight	North-east
J440	44° 00.0'	172° 35.0'	X1241	18.3	Canterbury Bight	North-east

TABLE 3. Minimum speeds of drift cards released in Pegasus Bay.

NZOI Stn No.	Release Point		NZOI Card No.	Minimum Velocity (cm s ⁻¹)	Release Location in Bay	Direction of Travel
	Latitude S	Longitude E				
J400	43° 12.3'	172° 56.8'	I1060	5.5	Northern	North-east
J401	43° 17.3'	172° 57.9'	I1088	3.4	North central	North-east
J402	43° 21.9'	172° 59.9'	I1130	0.7	Central	North-east
J403	43° 26.9'	173° 02.1'	I1168	1.5	South central	West
J403	43° 26.9'	173° 02.1'	I1177	3.2	South central	North-east
J404	43° 31.8'	173° 04.0'	I1214	10.3	South-eastern	North-east
J454	43° 37.4'	172° 55.2'	X1281	7.1	Southern	North-west

the *Glomar Tasman* oil rig (Fig. 12). The rig was in an area from which on previous occasions good card returns had been obtained along the beach to the north. However, of 450 releases only one was found, and that on the Chatham Islands. Wind vector diagrams, covering the period from the first release to seven days after the last release (Fig 13c), have a strong easterly component; this suggests that the cards were transported eastwards beyond the coastal current system.

NEAR-BOTTOM CURRENTS

The movement of water near the seabed was investigated with the aid of 860 Woodhead drifters (Woodhead and Lee 1960) released at the locations shown in Fig. 14. The drifters were tagged with standard NZOI drift cards, additional weight being added to compensate for the added buoyancy of the card. They were emplaced on the seabed in bunches of ten using soluble anchors made of 4.5 kg blocks of agricultural salt. Thirteen percent of the drifters were recovered, mainly from beaches. Others were retrieved offshore by bottom-trawling fishing boats.

Interpretation of seabed drifter returns is not simple and several factors must be taken into account. The most obvious factor is friction between the drifter stem

and the seabed which results in a low minimum velocity for the transporting current. Harden Jones *et al.* (1973), using data published by Ramster (1965) and Phillips (1970), suggested that at water speeds between 20 and 103 cm s⁻¹ a Woodhead drifter moves at 70–90% of the water velocity, and this percentage decreases when current speeds fall below 20 cm s⁻¹. This implies that the paths of the drifters reflect the direction of the fastest current and not that of the true residual current (Ramster 1965). Turbulence at high current speeds also affects drifters. Harden Jones *et al.* (1973), using a negatively buoyant drifter in a zone of fast tidal currents (100 cm s⁻¹), found that at slack water and low current speeds the drifter remained on the seabed whereas at higher speeds it was periodically suspended, these suspensions reaching a maximum height of 12 m above the seabed and lasting for 15 minutes, during which the drifter travelled at speeds as high as 94 cm s⁻¹. It is reasoned that the turbulence associated with the higher current speeds suspended the drifter, thus making it more susceptible to transport as it was beyond the dominant influence of the frictional boundary layer. It is evident that drifters are more likely to follow the direction of fast, turbulent flows than slower, less turbulent flows.

The effect of wave surge cannot be ignored. Morse *et al.* (1968) and Halliwell (1973) suggest wave action in shallow water may transport drifters. Waves in the

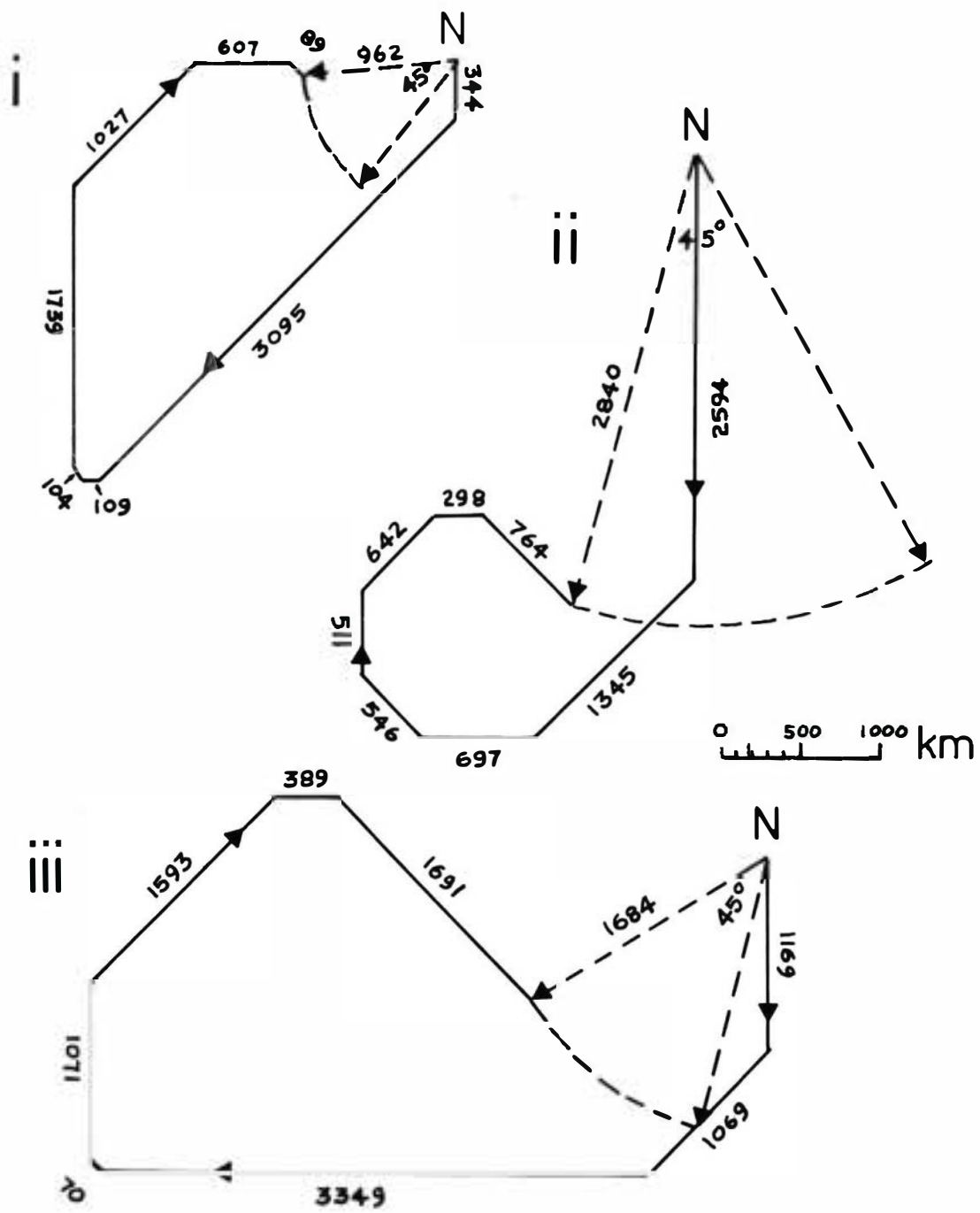


FIG. 13a. Progressive vector diagrams for wind run cumulated by octants during the period of travel of the ten fastest drift cards; 1000 h 9 October to 1000 h 7 November 1973. Observations from meteorological stations (i) Kaikoura; (ii) Winchmore; (iii) Harewood.

presence of linear currents may also be important; surge will presumably lift the drifters from the frictional boundary layer near the seabed, thus making them more susceptible to transport by linear currents.

If, as already mentioned, the fastest unidirectional flows on the Canterbury shelf act most frequently in the

same general direction as the Southland Current, the problem of reconciling the major trend of the drifters to that of the current regime on a broad scale is simple. Although the tides flow with equal velocity in either direction, both the Southland Current and dominant storm currents flow north-eastwards and this is



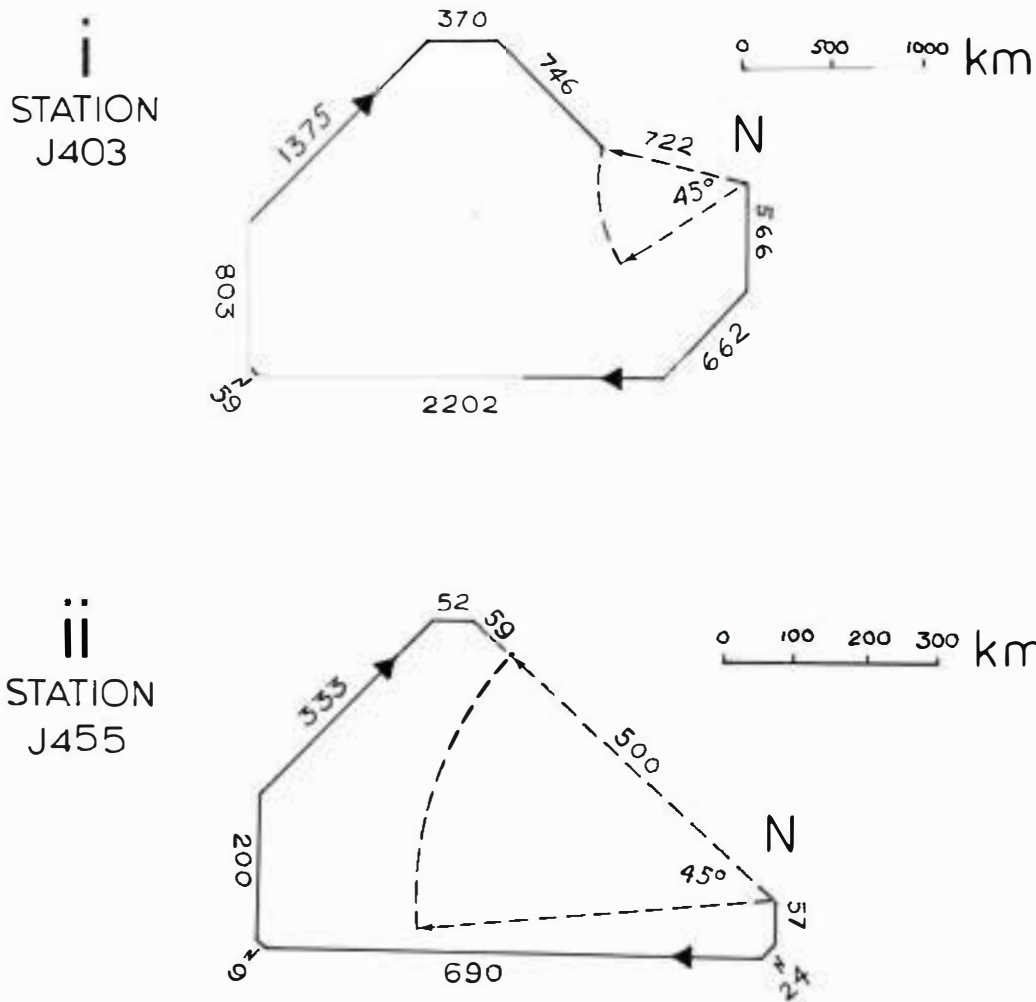


FIG. 13b. Progressive vector diagrams for wind run cumulated by octants during the periods of travel of the fastest card in each of the two west-going groups in Pegasus Bay. (i) 0900 h 13 October to 1600 h 16 October, 1973; (ii) 1000 h 9 October to 0700 h 29 October, 1973.

reflected in the vectors of the seabed drifters (Fig. 14).

The highest minimum velocity of the seabed drifters was 2.5 cm s^{-1} or 20% of that obtained for drift cards. This difference reflects both the lowering in speeds of currents as they approach the frictional boundary layer near the seabed and the friction between the drifter and the seabed (Harden Jones *et al.* 1973).

Besides the general northward trend of the vectors, another important trend is obvious, and that is the divergence between the onshore direction taken by the inner shelf drifters and the offshore direction taken by the middle and outer shelf drifters. This trend agrees well with the divergent sediment transport directions predicted from theory.

A small number of drifters in the south-west corner

of Canterbury Bight travelled southwards, e.g., one drifter finished 18 km south-west of the dropping locality after 25 days. A southward set has been reported from time to time by vessels steaming into Timaru. It seems likely, therefore, that a significant southward flow exists, which may be a permanent countercurrent produced by interaction of the mean flow with Banks Peninsula.

A lack of returns can be significant. Although 160 drifters were released in central and outer Pegasus Bay, there were no returns. This could mean that either the drifters were carried eastwards out of the bay or, more likely, bottom currents on the mud-floored bay are too feeble, sporadic, or random to move the drifters to a shore recovery site.

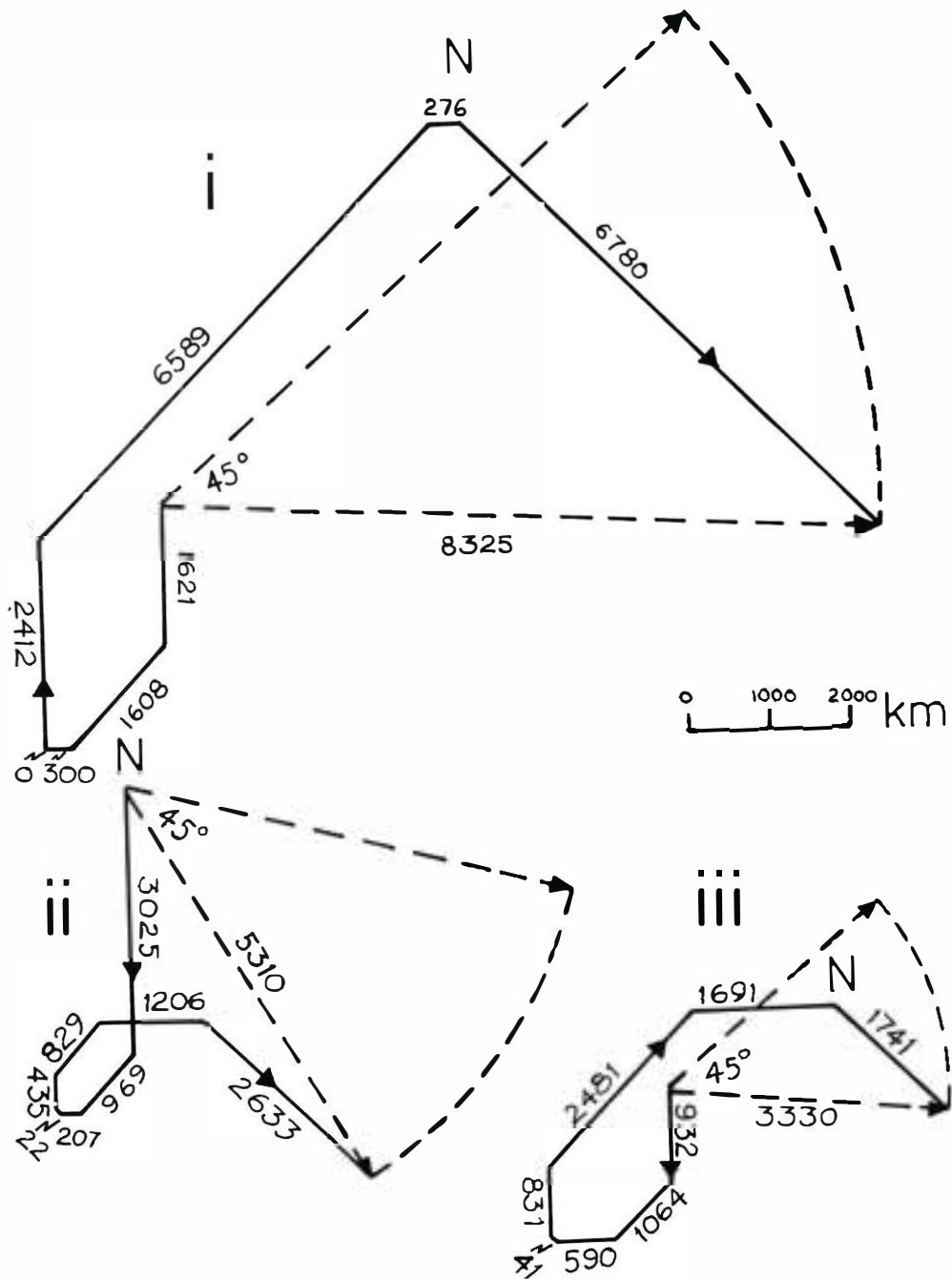


FIG. 13c. Progressive vector diagrams for wind run cumulated by octants during the period from the first card released from the stationary drillship to seven days after the last card release; 0000 h 8 July to 2400 h 4 August 1975. Observations from (i) *Glomar Tasman* drillship; (ii) Winchmore meteorological station; (iii) Harewood meteorological station.

CONCLUSIONS

1. The hydraulic regime on the Canterbury shelf consists of swell, meteorologically induced currents, tides, the Southland Current and, possibly, internal

waves. Theoretical evaluation of the velocities of the currents suggests they all play some role in transporting sediment.

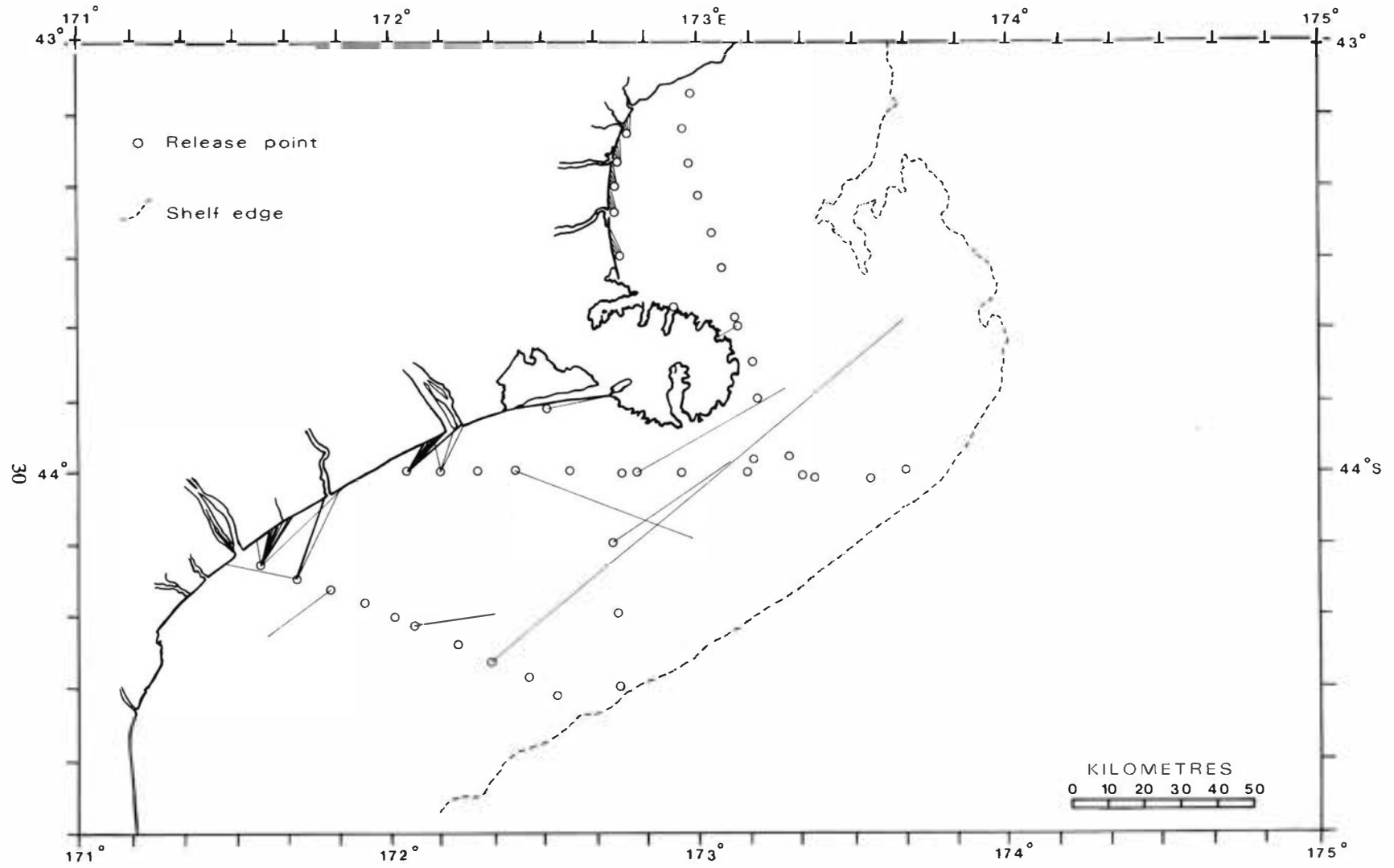


FIG. 14. Vectors between release and recovery points of seabed drifters. Thin lines represent vectors followed by one drifter; thick lines represent dominant drift vectors.

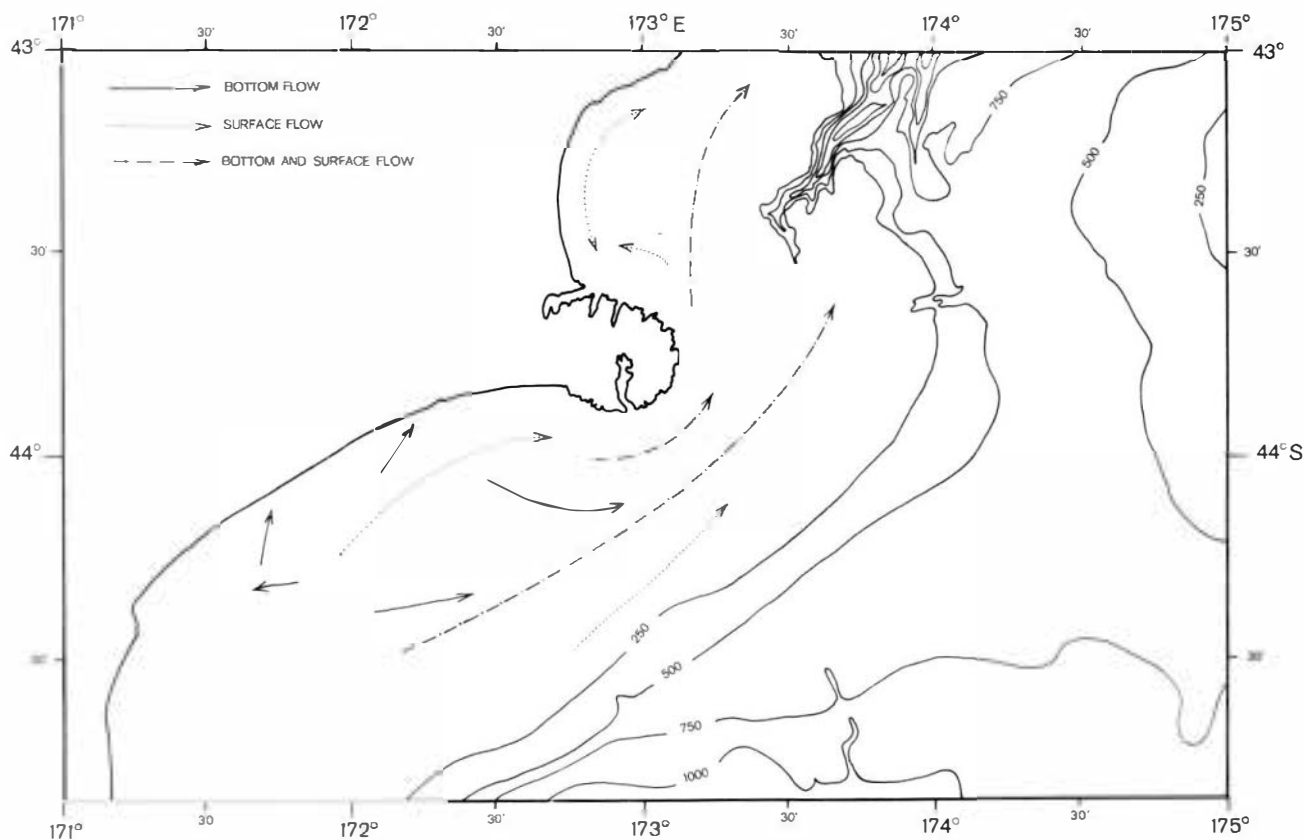


FIG. 15. Direction of net water flow on the Canterbury shelf based on drift cards, seabed drifters, and *Landsat II* photographs.

2. During calm weather the regime is dominated by tides, the Southland Current, and swell, which collectively effect a net north-eastward transport of inner shelf sediments. Transport in deeper water is negligible except on the tide-dominated shelf off Banks Peninsula and possibly at the internal wave-influenced shelf edge.

3. Currents induced by meteorological disturbances greatly reinforce existing motions to produce more widespread and higher rates of transport. Inner shelf sediments are moved almost continuously by the prevailing swell, wind waves, and wind drift. These same motions also transport middle shelf sediments, but less frequently. The principal storm currents on the outer shelf are wind-induced barotropic flow and, in

the presence of a strong pycnocline, a baroclinic flow. Either one, when reinforced by tides and the Southland Current, may transport bedload or wave-suspended bedload.

4. Gales and storms are mainly from the south and bring about an overall north-eastward transport of bedload coupled with an onshore component over the inner continental shelf and a slight offshore component over the remainder of the shelf. The prevailing directions of transport were substantiated by drift card and seabed drifter experiments (Fig. 15). Northerly disturbances may bring about a reversal of the trend on the wind drift-dominated inner shelf, but in deeper waters movement to the south is insignificant, or in some circumstances, replaced by northerly transport.

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and the general public for their assistance with the drift card and seabed drifter experiments. Finally, thanks go to R. A. Heath and K. B. Lewis for their critiques of the manuscript, Mrs R. H. Herzer and M. Yiannoutsas for draughting diagrams, and Mrs R. M. Thompson and Miss G. Marsden for typing services.

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