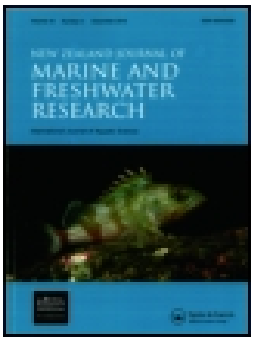

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ROLE OF MEAN CIRCULATION, TIDES, AND WAVES IN THE TRANSPORT OF BOTTOM SEDIMENT ON THE NEW ZEALAND CONTINENTAL SHELF

L. CARTER and R. A. HEATH

New Zealand Oceanographic Institute, Department of Scientific and Industrial Research, P.O. Box 8009, Wellington

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ABSTRACT

Evaluation of velocity data on water movements over the New Zealand continental shelf has revealed that the mean circulation by itself is too slow to induce transport of bottom sediments. Tides generally have higher velocities, but are still not the main transporting agent except in the tide-dominated Cook and Foveaux Straits. Waves have the potential to stir sediments on the inner and middle shelf (less than about 70 m deep) during annual storms, and probably down to 130 m depth during the maximum 25-y storm.

For sediment transport to take place, energies of at least two of the major water movements would have to complement one another. Optimum conditions for transport probably occur during storm periods when wave-suspended sediment is readily moved by tides and the mean circulation.

The direction of transport is mainly along the continental shelf and is largely in response to prevailing weather patterns coincident with the direction of the mean circulation and strongly reinforced by the appropriate phase of the tide.

INTRODUCTION

Correlation between the dispersal of sediments on the continental shelf and the direction of water movements is often reported in the literature. For the New Zealand shelf, sediment dispersal patterns deduced from the distribution of sand and gravel grain sizes have been related to the Southland Current on the Otago shelf (Andrews 1973) and to the Westland and D'Urville Currents off northwest Nelson (van der Linden 1969). Pantin (1961) used mineral tracers to detect directions of sediment movement in Cook Strait, which he equated with the strong tidal flow in this area. Likewise, Summerhayes (1969a) related the distribution of heavy minerals around northernmost New Zealand to possible eddy systems derived from the East Auckland and West Auckland Currents. However, although this correlational procedure is commonly used, there has been no detailed examination of the roles played by the major components of water movement in the transport of sediments on the continental shelf.

Although the greatest transport of sediment probably occurs in the littoral and uppermost neritic zone (e.g., Swift 1969, Kirk unpublished 1970), we will not deal with this hydrologically complex region here, but rather with the area further offshore encompassing most of the continental shelf. Firstly, bottom stresses and current speeds necessary to initiate sediment movement are mentioned in terms of simple theory and pre-existing experimental data; this allows estimates of minimum current speeds in the actual situations, which are described later in the paper. Secondly, the preferred theory of sediment transport rate is discussed, with regard to the mean circulation, tides, and wave-generated currents of the New Zealand shelf waters. These water motions are discussed in terms of their potential to move sediment. Case histories are used from two areas where sediment dispersal and water circulation are relatively well known. Finally, a broad picture of sediment transport is presented for the entire New Zealand shelf.

SOURCES OF DATA

Analyses of coastal currents around New Zealand have mainly been based on the use of the geostrophic method, changes in water mass properties and drift card experiments. Although the direction of the mean circulation has been reasonably well established (Fig. 1) by these methods (*see* e.g., Brodie 1960; Garner 1969; Heath 1973a) each method is subject to strong limitations so far as deriving absolute speeds. Those speeds, summarised in Fig. 3, are for mean flow at the *surface*, but the speeds are probably lower near the bottom. The most widely spread set of direct current measurements around New Zealand are those made with current poles over single tidal periods by the Royal New Zealand Navy at many sites shown on the 1:200,000 hydrographic charts published by the Navy Hydrographic Office. Other sources of direct current measurements are the log ship and current meter observations of Gilmour (1960) in Cook Strait and the subsurface current

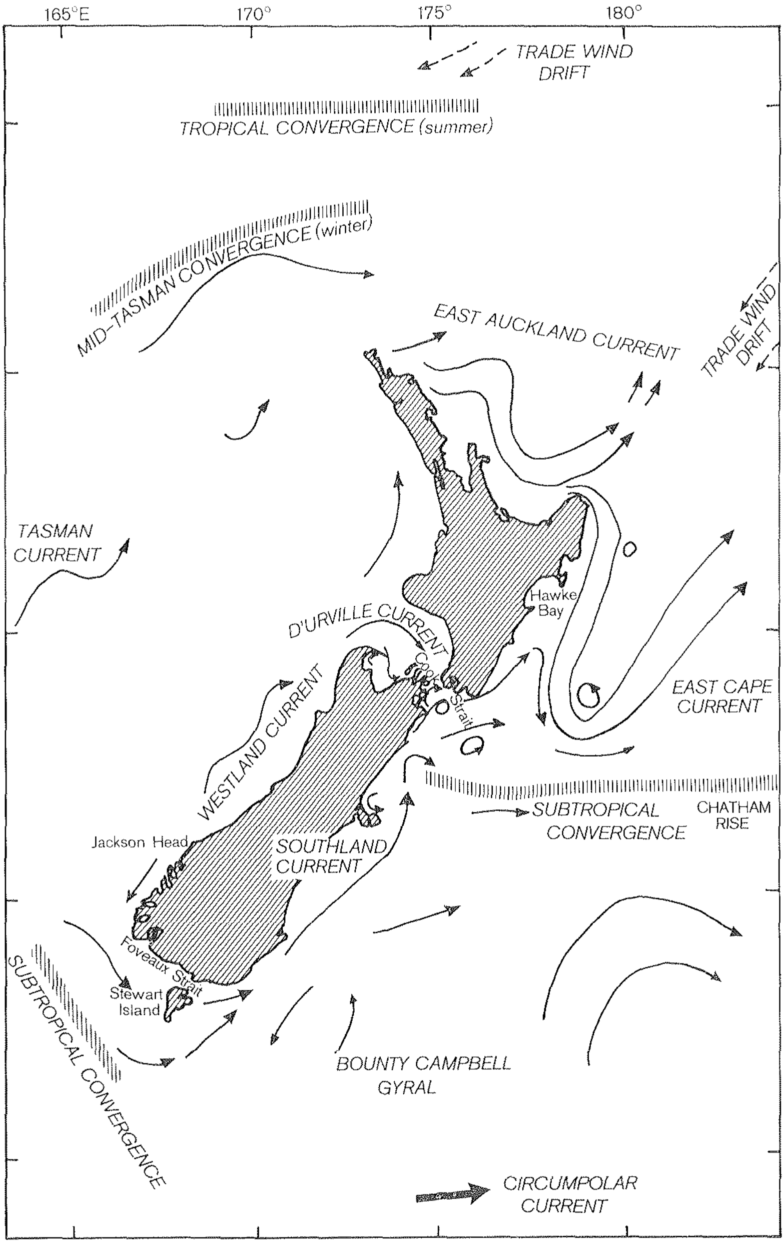

FIG. 1—Mean circulation for the New Zealand region, after Heath (1973a).

TABLE 1—Computed speeds at the surface and near the bottom of waves on the middle Taranaki (i.e., central western) shelf, based on wave records from the oil drilling rigs *Discovery 2* and *Sedco 135-F* (* = as for maximum amplitude)

POSITION		DEPTH	PERIOD OF RECORD		MAXIMUM AMPLITUDE SWELL				MAXIMUM SPEED		MAXIMUM PERIOD SWELL				MAXIMUM SPEED	
Lat. °S	Long. °E	(m)	Start	Finish	Height (m)	Period (s)	Direction (°T)	Date	Surface (cm.s⁻¹)	Bottom (cm.s⁻¹)	Height (m)	Period (s)	Direction (°T)	Date	Surface (cm.s⁻¹)	Bottom (cm.s⁻¹)
38.5	173.3	175	7–10–68	19–1–69	9	8	250	19–11–68	353	.006	6	10	320	17–11–68	186	0.4
39.3	173.5	120	20–1–69	27–3–69	6	9	140	11–2–69	211	1.0	*	*	*	*	*	*
39.6	173.5	110	15–10–69	31–12–69	7.5	8	230	18–10–69	294	0.6	2.5	14	010	30–11–69	57	11.4
39.2	173.7	90	1–1–70	21–1–70	6	9	130	7–1–70	211	4.5	*	*	*	*	*	*
39.5	173.5	110	22–1–70	22–2–70	6	8	120	8–2–70	235	0.5	3.5	10	130	10–2–70	112	2.3
40.3	173.4	75	7–3–70	26–3–70	3.5	5	70	14–3–70	217	0.0016	1	14	350	25–3–70	24	9.0
38.2	174.2	95	31–3–70	4–5–70	5	7	240	9–4–70	227	0.076	3	12	240	18–4–70	79	11.1
40.0	173.2	110	10–5–70	16–7–70	6	8	230	25–5–70	235	0.5	*	*	*	*	*	*
42.2	171.2	25	29–7–70	4–8–70	4.5	8	030	2–8–70	180	72	3	10	240	4–8–70	112	64
40.3	172.2	110	13–8–70	6–9–70	7.5	7	020	27–8–70	340	0.08	4	11	260	8–8–70	107	5.6
45.2	171.1	37	24–10–70	6–12–70	5.5	7	190	13–12–70	160	14	1.5	10	010	1–12–70	48	23

drogue measurements at ten coastal positions around the South Island by Heath (1973b). The only direct evidence for the direction of flow near the seafloor is provided by recently completed sea-bed drifter experiments on the Canterbury Shelf (R. H. Herzer, NZOI, pers. comm.).

Wave data are sparse, with the most comprehensive records coming from oil rigs on the Taranaki shelf. These records extend over a period of 2 y and note the height, period, and direction of wind waves and swell at 6-hourly intervals (Table 1). Additional wave data come from (i) a few recording stations maintained by harbour authorities (e.g., Watters 1953); (ii) shipboard observations (NZOI files) and (iii) hindcast statistics (Glenn *et al.* unpublished 1973).

RELATIONSHIP BETWEEN BEDLOAD SEDIMENT TRANSPORT AND BOTTOM STRESS

GENERAL PRINCIPLES AND DISCUSSION OF PREVIOUS WORK

The critical or minimum stress needed to initiate sediment movement under different circumstance of grain size and difference in relative density between the sand and water have been measured both in the laboratory (e.g., Inman 1963) and in the marine environment (e.g., Sternberg 1967, 1971). The critical stress required for the appropriate sediment characteristics are displayed in competency curves (e.g., Hjulstrom 1939, Sundborg 1956). Assuming that sediment movement has been initiated, transport can be evaluated using the often quoted theory of Bagnold (1963) in which the rate of mass transport of sediment (Q) is related to the power expended by the fluid moving over the bottom (ϕ) by

$$\frac{\rho_s - \rho_0}{\rho_s} gQ = \mathrm{k}\phi \qquad (1)$$

where ρ_s, ρ_0 are the densities of sediment and water respectively, k the coefficient of proportionality related to the proportion of the total power used in sediment movement, and g the acceleration of gravity (see e.g., Sternberg 1972). k has been shown experimentally by Bagnold (1963) to reach a constant value in flumes and rivers depending on the relative roughness of the bottom. However, field measurements made in the marine environment by Kachel & Sternberg (1971) indicate that k varies as a function of the excess boundary stress

$$\frac{\tau_0 - \tau_c}{\tau_c}$$

where τ_0 is the actual stress, and τ_c the critical stress. To date, however, this has not been fully tested under marine conditions with short period waves present. Values of k (Kachel & Sternberg 1971) and the subsequent curves given by Sternberg (1972), which are based on the flume tests of Guy *et al.* (1966), are for uniform flows. With a known value of the actual stress τ_0, and the critical stress τ_c (derived from competency curves), the value of k can be found from the curves of Sternberg (1972). The rate of sediment transport can then be calculated using equation (1). Of the three methods available for calculating the stress, both the direct calculation of the Reynolds stress (involving correlations between fluctuating velocity components) and the velocity profile method (requiring installation of several current meters to determine the vertical shape of the current profile) use data which are not available to us. However we can use the experimental result that the boundary stress $\underline{\tau}$ is proportional to the square of the mean speed and the fluid density, i.e.,

$$\underline{\tau} = C_{100}\, \rho_0 |\underline{U_{100}}| \underline{U_{100}}$$

where $\underline{U_{100}}$ is the mean velocity 1 m above the sea bed, ρ_0 the water density, and C_{100} the drag coefficient. Values of C_{100} for Reynolds numbers (*Re*) are given, for instance, by Sternberg (1972). He found that for values of U_{100} (mean speed) above 15 cm.s^{-1} ($Re \simeq 1.5 \times 10^5$) the flow is essentially hydrodynamically rough, and C_{100} is nearly constant, having a value of 3×10^{-3}. This speed of 15 cm.s^{-1} is well below the threshold speed for sand transport. Therefore in the calculations a constant value of 3×10^{-3} will be used for C_{100}.

EVALUATION OF THE RATE OF SEDIMENT TRANSPORT

The relevant quantity to be calculated to derive sediment transport rates is $k\Phi$ – the power expended in moving the sediment. In the case of a mean current velocity $\underline{U_0}$ and tidal velocity $\underline{U_T} \cos \Omega\, t$, the term $k\Phi$ in the direction of the unit vector $\underline{n}$ is given by

$$k\Phi = \frac{\rho\, C_{100}}{T} \Bigg[\int_{-t_1}^{t_1} \mathrm{fn}\left(\frac{\tau_0 - \tau_c}{\tau_c}\right) \{ (\underline{U_0} + \underline{U_T} \cos \Omega\, t) . (\underline{U_0} + \underline{U_T} \cos \Omega\, t)\} \quad \{\underline{U_0} + \underline{U_T} \cos \Omega\, t\} . \underline{n}\, dt$$
$$+ \int_{T/2 - t_2}^{T/2 + t_2} \mathrm{fn}\left(\frac{\tau_0 - \tau_c}{\tau_c}\right) \{ (\underline{U_0} + \underline{U_T} \cos \Omega\, t) . (\underline{U_0} + \underline{U_T} \cos \Omega\, t)\} \quad \{\underline{U_0} + \underline{U_T} \cos \Omega\, t\} . \underline{n}\, dt \Bigg] \qquad (2)$$

where $\pm\, t_1$ and $T/2 \pm t_2$ are cut-off times outside which the flow speed is insufficient to initiate sediment transport.

Taking the simple case where $\underline{U_0}$, $\underline{U_T}$ and $\underline{n}$ are all parallel, then equation (2) reduces to

$$k\Phi = \frac{\rho\, C_{100}}{T} \Bigg[\int_{-t_1}^{t_1} \mathrm{fn}\left(\frac{\tau_0 - \tau_c}{\tau_c}\right) (U_0 + U_T \cos \Omega\, t)^3 + \int_{T/2 - t_2}^{T/2 + t_2} \mathrm{fn}\left(\frac{\tau_0 - \tau_c}{\tau_c}\right) (U_0 + U_T \cos \Omega\, t)^3 \Bigg] dt \qquad (3)$$

with

$$t_1 = \frac{T}{2\Pi}\left[\cos^{-1}\left(\frac{U_c - U_0}{U_T}\right)\right]; \quad t_2 = \frac{T}{2\Pi}\left[\cos^{-1}\left(\frac{+U_c + U_0}{U_T}\right)\right]$$

where U_c is the lowest speed at which sediment will still be eroded, T the tidal period, Ω the tidal angular frequency, and $k = \text{fn}\left(\frac{\tau_0 - \tau_c}{\tau_c}\right)$.

[We are considering sediment transport past a single point and therefore Eulerian rather than Lagrangian current measurements are appropriate. When studying the trajectory of sediment movement, Lagrangian measurements are generally preferred. However, as sediment is not transported at the same speed as the water flow, (unless no large spatial change in water velocity occurs, in which case either type of measurement would suffice), neither measurement is appropriate for quantitative analysis. Many sediment dispersal patterns indicate a mean transport direction similar to the water flow pattern. This only indicates that sediment movement takes place when various components of the water flow reinforce one another. Quantitative analysis within such a system is probably best, with Eulerian current observations at numerous points.]

For k we will assume a constant value. The justification for this is that although k increases rapidly with an increase in the ratio τ_0/τ_c, on the continental shelf, the change in τ_0/τ_c (for $\tau_0 > \tau_c$) is generally small, with the sediments moving only during a short period around peak ebb and flood tide. The largest contribution to the integral occurs near these peak times and therefore a representative value of k near this time of the tidal cycle will be used. With this assumption, carrying out the integration in equation 3 gives

$$k\Phi = k\,\rho\,C_{100}\left\{2U_0^3\left(\frac{t_1 + t_2}{T}\right) + \frac{3U_0^2 U_T}{\Pi}\sin\Omega t\right]_{t_2}^{t_1} + \frac{3U_0 U_T^2}{4\Pi}\sin 2\Omega t\right]_{t_2}^{t_1} + 3U_0 U_T^2\left(\frac{t_1 + t_2}{T}\right) + \frac{U_T^3}{\Pi}\left(\sin\Omega t - \frac{\sin^3\Omega t}{3}\right)\Bigg]_{t_2}^{t_1}\right\} \quad (4)$$

With short period waves (mainly swell) also present, evaluation of $k\Phi$ becomes more complicated (Fig. 2). For waves with bottom velocity $U_w \cos \omega t$ and angular frequency ω, the start of the cut-off in erosion for the flood tide ($\omega >> \Omega$) occurs when

$$U_T \cos\theta_1 - U_w + U_0 = U_c \quad (5)$$

and for the ebb tide when

$$U_T \cos(\Pi - \theta_2) + U_w + U_0 = -U_c \quad (6)$$

The respective cut-off times $t_1 = \dfrac{\theta_1}{\Omega}$ and $t_2 = \dfrac{\theta_2}{\Omega}$ are then given by

$$t_1 = \frac{1}{\Omega} \cos^{-1} \left(\frac{U_c + U_w - U_0}{U_T} \right)$$

$$t_2 = \frac{1}{\Omega} \left\{ \Pi - \cos^{-1} \left(\frac{-U_c - U_w - U_0}{U_T} \right) \right\}$$

With times greater than these cut-off values, sediment will remain unmoved for increasingly longer in the swell cycle, e.g., for the first cut-off time (near flood tide) the subsequent times of cut off (t_2, t_3, t_{n+1}) are given by

$$U_T \cos \{\theta_n + \Omega (\Delta t_{n+1})\} + U_w \cos \omega (\Delta t_{n+1}) = U_T \cos (\theta_n) - U_w$$

i.e., $t_2 = t_1 + \Delta t_2$, $t_3 = t_2 + \Delta t_3$, $t_{n+1} = t_n + \Delta t_{n+1}$.

If U_0, the mean speed, was zero, this would not create a problem, for then the complicated part of the cycle near high tide, (when the speed associated with the swell is just enough to cause the stress to alternatively be above and then below the critical level needed for erosion) is balanced by the similar period near low tide. Of course, as visualised here, the process including swell is very much simplified, as cessation of erosion will not be instantaneous at the cut-off time, especially if sediment particles are being suspended. Alternatively, if U_0 is large, these complicated parts of the tidal cycle will not compensate. As can be seen from Fig. 2, a measure of the asymmetry in the transport produced at these times is given by the difference between θ_1 and θ_2. On adding equations (5) and (6) while assuming $\Delta \theta = \theta_1 - \theta_2$ is small we have

$$\Delta \theta = \frac{2 U_0}{U_T \sin \theta_2}$$

and we see that for U_0/U_T small, which is the usual case, $\Delta \theta$ is also small. Further, the error in the power arising from neglecting these complicated times of the tidal cycle will be minimised because the power depends on the cube of the speed, and most of the contribution will then come near the peak tidal flows, which will not generally be when the waves are producing an erosion cut-off.

The amount of power expended $k\phi$ for the case with swell present is given by

$$k\Phi = \frac{k \rho C_{100}}{T} \left\{ \int_{-t_1}^{t_1} (U_0 + U_T \cos \Omega t + U_w \cos \omega t)^3 \, dt \right.$$

$$+ \int_{T/2 - t_2}^{T/2 + t_2} (U_0 + U_T \cos \Omega t + U_w \cos \omega t)^3 \, dt$$

$$= k \rho C_{100} \left\{ \phi + 3 U_0 U_w^2 \left(\frac{t_1 + t_2}{T} \right) + \frac{3 U_T U_w^2}{2\Pi} (\sin \Omega t) \Big]_{t_2}^{t_1} \right\} \quad (7)$$

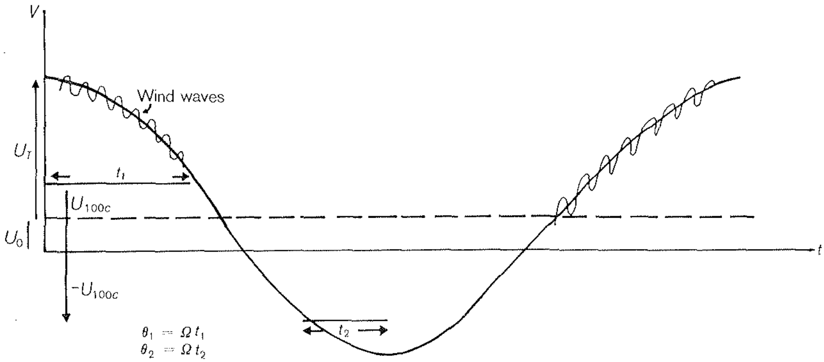

FIG. 2—Current speed (U) with time (t) over a tidal cycle; U_T the amplitude of the tidal speed, U_0 the speed of the mean flow, U_{100c} the critical speed below which sediment will not be eroded, Ω the tidal angular frequency. For simplicity, wind waves are superimposed on only part of a tidal cycle.

where $k \rho C_{100} \phi$ is the value without swell, but the cut-off times $t_1 = \theta_1/\Omega$, $t_2 = \theta_2/\Omega$ are those for the case with swell.

This brief analysis illustrates why in even moderate flow conditions sediment movement is highly time dependent. Alternatively, substantial sediment movement might only occur during the maximum 25-yearly storm. It is evident that we not only require a detailed knowledge of the speed, direction, and time dependence of each component making up the total water velocity, but also a knowledge of the long-term storm climate.

For the case when the bottom velocities of wind waves are substantial, a further simple analysis similar to that given previously, can be made for the cut-off times t, taking into account the wind-wave time dependence, e.g.,

$$U_T \cos \Omega t + U_w \cos \omega t = U_c - U_0$$

In this case most sediment movement will occur when the tidal flow is largest. However the time-averaged sediment transport will be small, unless the mean flow is large.

MEAN CIRCULATION ON THE CONTINENTAL SHELF

The mean circulation around New Zealand (Fig. 1) is strongly influenced by an easterly drift from the south Tasman Sea. On meeting the shelf and coast this flow branches near Jackson Head into

(i) a southern component, which flows through Foveaux Strait and around Stewart Island to continue north over the shelf and slope off the east coast of the South Island as the Southland Current, and

(ii) the Westland Current, which moves north at least as far as Cape Egmont.

The Southland Current branches near Kaikoura, where one component meanders to the east and the other continues north to southern Cook Strait, where the flow is modified by mixing with the D'Urville and East Cape Currents. This modified Southland Current travels eastwards and then back southwards south of Hawke Bay (Heath 1972).

Off northern Cook Strait the Westland Current has an offshoot, the D'Urville Current, part of which sweeps directly into Cook Strait whereas the remainder follows a circuitous clockwise route south of Cape Egmont (Heath 1969).

The flow out of the Tasman Sea also gives rise to the East Auckland Current which moves south-eastwards along the east coast of the North Island, between North Cape and East Cape (Barker & Kibblewhite 1965). Near East Cape the main flow of the East Auckland Current turns north, while the rest turns in a clockwise direction around East Cape, giving rise to the south-trending East Cape Current.

VARIABLE WATER MOTIONS

Superimposed on the mean circulation are motions induced by tides and meteorological disturbances. The latter include wind waves, swell, wind-induced currents, and barometric storm surge. Because of a lack of data on some of these components, we will deal mainly with tidal and wave motions. The theory of motions that can occur are covered in most books on physical oceanography e.g., Neumann & Pierson (1966) and a summary in the geological literature is given by Weggel (1972). To obtain estimates of bottom current velocities associated with grounding swell and short period waves we have used the well-known simple wave theory to enable extrapolation downwards from surface observations. In this paper a brief summary of the simple theory rather than the isolated equations is presented.

TIDAL FLOW

Of the variable motions, the tidal flow is the most regular and in general dominates the mean flow (Fig. 3), particularly where tidal flow is constricted, as between the three main islands of New Zealand (Gilmour 1960; Heath 1969, 1973b). The local tidal cycle is commonly asymmetric, with ebb and flood phases having different durations. Higher tidal speeds tend to occur during the shorter phase of the tidal cycle.

WAVES

For a wave with small vertical displacement of the free surface (ζ) travelling in a positive x direction

$$\zeta = A \cos (kx - \omega t)$$

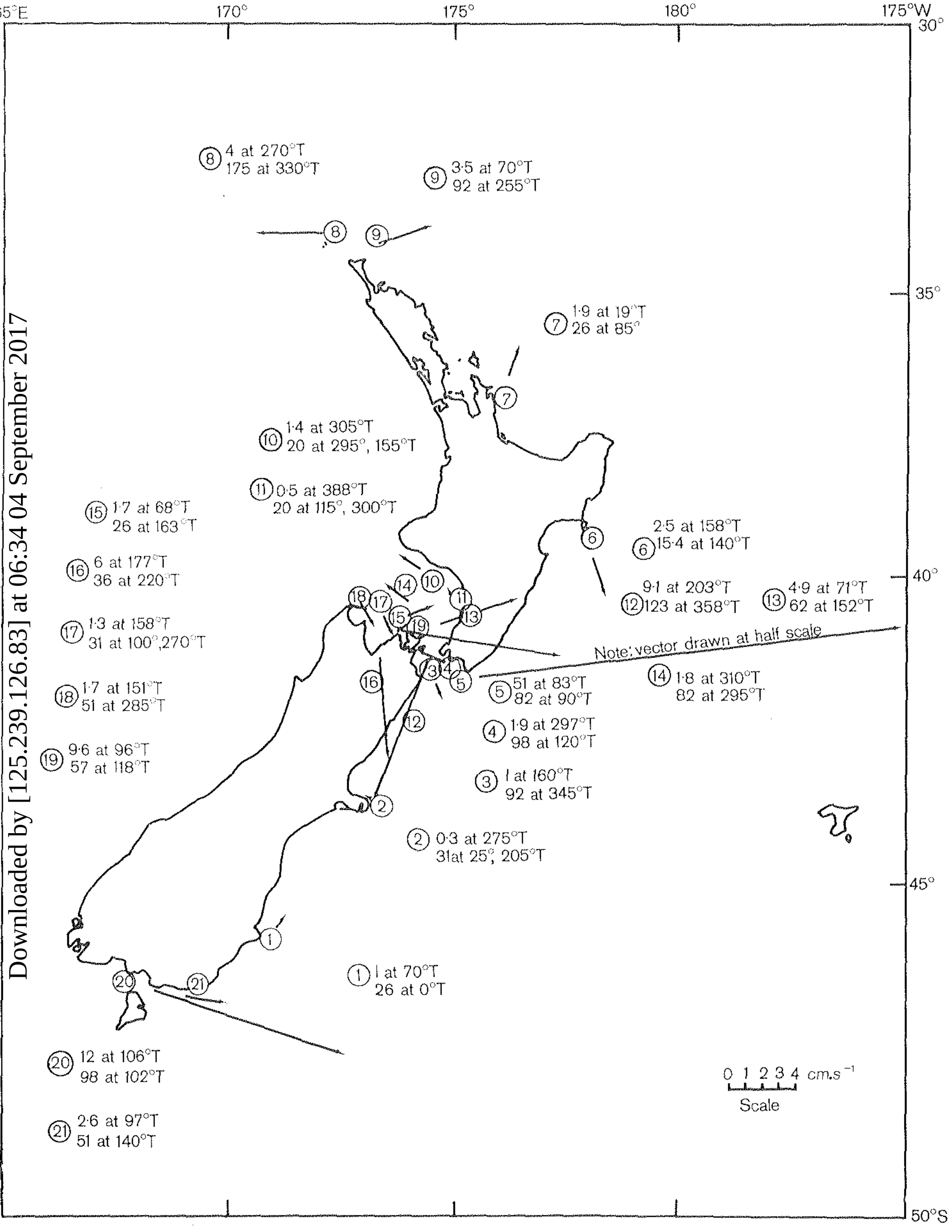

FIG. 3—Time-averaged mean velocity (smaller value) and maximum velocity of **surface** flow measured at stations occupied by the Royal New Zealand Navy over single tidal cycles. Arrows indicate the mean velocity. Velocity measurements are from 1:200,000 hydrographic charts: Station 1, Hydrographic Branch 1952a; Station 2, Hydrographic Branch 1953a; Stations 3, 4, 5, Hydrographic Branch 1953b; Station 6, Hydrographic Branch 1958; Station 7, Hydrographic Department, 1943; Stations 8, 9, Hydrographic Office 1972; Stations 10, 11, 12, 13, 14, Hydrographic Branch 1960a; Stations 15, 17, 18, 19, Hydrographic ... 20, 21, Hydrographic Branch 1952b; Station 16, Hydrographic

where A is the amplitude, k the wave number, ω the angular frequency, t the time. In water of depth d, the velocity potential ϕ is given by

$$\phi = \frac{A\omega}{k} \frac{\cosh k\,(Z + d)}{\sinh kd} \sin\,(kx - \omega t)$$

(see, e.g., Phillips 1966). The horizontal component of velocity is given by

$$\frac{\partial \phi}{\partial x} = U = A\omega \frac{\cosh k\,(Z + d)}{\sinh kd} \cos\,(kx - \omega t) \qquad (8)$$

The factor $A\omega \dfrac{\cosh k\,(Z + d)}{\sinh kd}$ corresponds to U_w in equation (7).

Using the available measurements with a known period and depth, the wave number can be calculated from the dispersion relationship

$$\omega^2 = g\,k \tanh kd$$

and hence the bottom speed can be calculated from equation (8).

The most extensive set of wave records at hand are from oil drilling rigs formerly located on the middle shelf off Taranaki (see Fig. 5). Speeds at the seafloor and sea surface calculated from the maximum range and largest period swell observed from the oil rig records are shown in Table 1.

WATER MOTIONS AND SEDIMENT TRANSPORT AROUND NEW ZEALAND

SPEEDS TO INITIATE SEDIMENT MOVEMENT

Speeds of near-bottom currents required to move sediment of different grain sizes can be determined from the competency curves of, for example, Sundborg (1956) and Hjulstrom (1939). However, such curves should be used with discretion, as the threshold mean speed U_{100c} (speed 1 m above the seafloor required to initiate movement of bottom sediments) is not only related to sediment grain size but also to grain density and hydraulic shape and to seafloor morphology. On the New Zealand continental shelf there is a great range of sediment types and substrates (see, e.g., Carter in press).

SEDIMENT TYPES

The shelf is mantled mainly by modern and relict terrigenous and biogenic sediments with a wide range of grain sizes (Fig. 4).

Modern terrigenous sediments dominate the Westland and Wairarapa–Hawkes Bay shelf, as well as nearshore zones of other sections of the shelf (e.g., van der Linden & Norris 1974; Pantin 1966; Lewis 1973). Near shore, medium to fine sands prevail, whereas muds dominate middle and outer shelf positions. As the shapes and densities of the sands approximate those of sands used experimentally to determine the competency curves (except for dense iron sand deposits of New

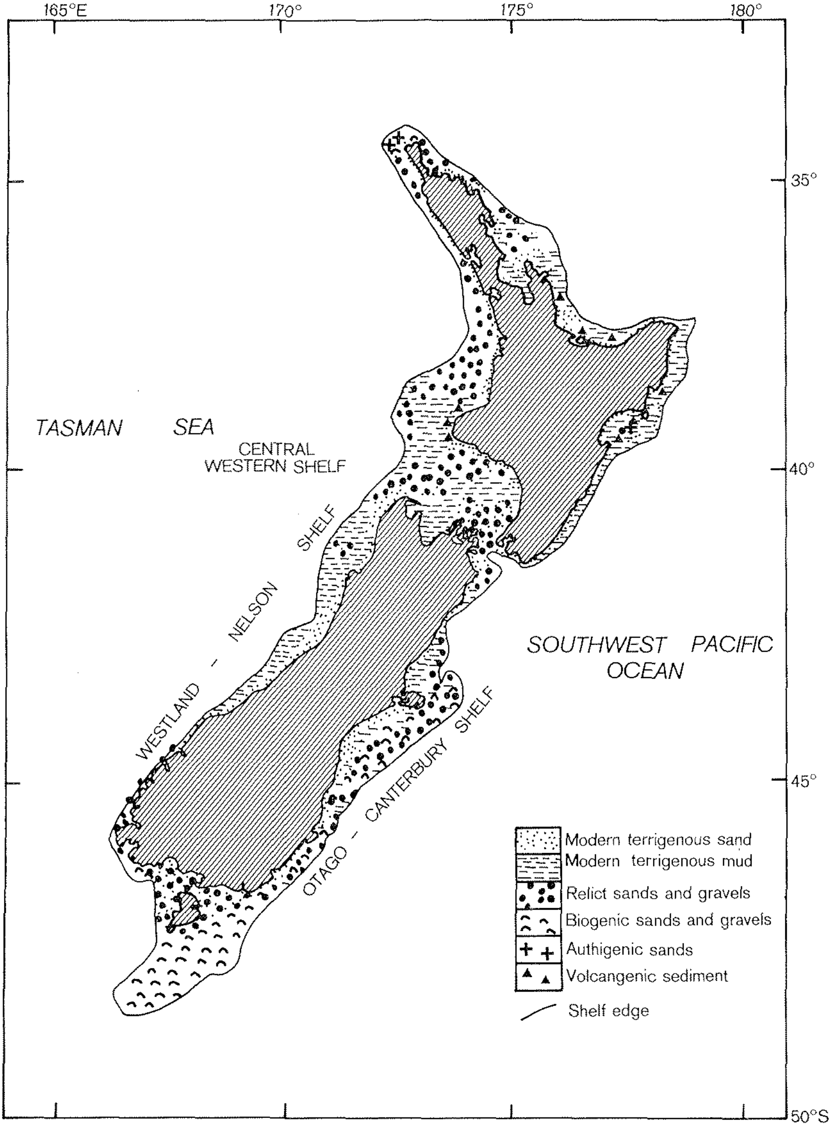

FIG. 4—Distribution of main sediment types on the New Zealand continental shelf (in legend, read *volcanogenic* for *volcangenic*)

Zealand's west coast), U_{100c} can be directly related to grain size. For modern sands U_{100c} is 35–50 $cm.s^{-1}$ based on the curves of Hjulstrom (1939) and Sundborg (1956).

Relict terrigenous sediments occupy middle and/or outer shelf positions, particularly off Otago–Canterbury (Andrews 1973; R. H. Herzer, NZOI pers. comm.), Southland (Cullen 1967), Taranaki (McDougall & Brodie 1967), and northernmost New Zealand (Summerhayes 1969a). Again, fine to medium sands prevail, although gravels are locally important. Values of U_{100c} for these sands will no doubt be similar to the values for modern sands, although U_{100c} for gravels will be considerably higher.

Calcareous biogenic sediments are common in areas where modern terrigenous sedimentation is low, such as middle to outer shelf locations off Southland (Cullen 1967), Otago (Andrews 1973), and Northland (Summerhayes 1969a). Application of U_{100c} versus grain size curves cannot be applied with confidence to biogenic sands on account of their variability in density and hydraulic shape. Published competency curves are based on behaviour of quartz or other terrigenous grains of similar density, and such grains generally have a restricted range of hydraulic shapes. By comparison, biogenic clasts have variable densities depending on the structure of the shell material and its degree of chemical and physical alteration, e.g., Blatt *et al.* (1972). Shape is also varied, with clasts of branching bryozoa, plate-like pelecypod shells, rod-shaped echinoid spines, conical gastropod shells, and variously shaped foraminiferal tests (see, e.g., Summerhayes 1969a, Andrews 1973). Such grains tend to react to currents differently from equivalent-sized terrigenous grains. Recent experimental work (K. B. Lewis, NZOI, pers. comm.) demonstrated that when mixed biogenic and terrigenous sands (from the middle shelf off Wanganui) are subjected to current speeds approaching 30 $cm.s^{-1}$, the first grains to move are *medium* to *coarse* sand-sized molluscan fragments, followed by *fine* quartz and feldspar sand.

MORPHOLOGY OF SUBSTRATE

Any departure from a smooth bed form, through development of sedimentary structures (e.g., ripple marks), the presence of rocky exposures, or the presence of a variety of different grain sizes, will effectively alter the U_{100c}/grain size relationship, chiefly through instigation of turbulent flow, e.g., observations by Sternberg (1971) showed that U_{100c} values of 36.8–38.8 $cm.s^{-1}$ were sufficient to move fine sand (mean size 0.42 mm) over a rippled substrate, whereas over a pebble gravel substrate a similar speed (39.0 $cm.s^{-1}$) was capable of moving coarse sand (mean size 1.09 mm). Around New Zealand a wide variety of substrates are present, including rough rocky bottoms around Northland (Summerhayes 1969b), mixed gravel sand substrates in Foveaux Strait (Cullen 1967), and rippled sands in northern Cook Strait (Estcourt 1968), all of which complicate the picture of sediment transport.

We will take a simplistic view and use published competency curve data, while bearing in mind that such curves are approximate for sediments that deviate from experimental criteria used to determine the curves, due to variability in properties of sediment grains and substrate.

ABILITY OF WATER MOVEMENTS TO TRANSPORT SEDIMENT

Sediment movement is influenced by the three main water movements: the mean circulation, tides, and waves, and each of these will be briefly considered, although when evaluating sediment transport the combined effect of the individual motions must be taken into account. Overall the mean circulation is weak, and most of the energy for transport must be supplied by the tides and storm-driven water movements. It should be emphasised that, as the sediment transport depends on the cube of the velocity (see equations 1, 3, 7), high current speeds over a short time transport more sediment than slower currents operating for a longer time; for this reason most transport might occur in infrequent periods of storm conditions. Further, as previously mentioned, sediment transport is likely to be in the direction of the mean circulation (i.e., the direction in which the components reinforce one another), unless there is a strong asymmetry in the distribution of storm directions opposite to the direction of mean flow.

MEAN CIRCULATION: In general, direct measurements reveal the mean circulation is too weak to instigate transport of very fine sand or coarser sediment (Fig. 3). An exception is off the southwest coast of the South Island, where Heath (1973a) reported a mean flow of 64 $cm.s^{-1}$ at 100 m depth.

TIDAL FLOW: Most tidal velocities have been determined at the sea surface (Fig. 3). Unless there is an internal tide generated, tidal currents will be less near the seafloor, e.g., Gilmour (1960) demonstrated that tidal current speeds in Cook Strait can decrease by as much as two-thirds near the bottom. It is evident from Fig. 3 that tides are too weak to move sediment on the open shelf, but where tidal flow is constricted, as in Cook and Foveaux Straits, it is the dominant force, reaching speeds of 50 $cm.s^{-1}$ in Cook Strait (Gilmour 1960).

WAVES AND SWELL: Records from offshore oil rigs show that the speed at the bottom associated with waves is not sufficient to instigate movement of bedload at the depth range recorded (90–175 m; Table 1). However, at shallower depths, bottom surge speeds would be increased, as indicated by speeds from the shallow station (37 m) on the south Canterbury shelf (Table 1).

It is only in a few isolated places that any one of the aforementioned components of water motion has sufficient impetus to move bedload. Consequently, if sediment movement is to take place, then it must be under the combined energies of these components. To illustrate how different components would interact, two situations have been chosen (central western shelf and Otago–Canterbury shelf) on the basis that these two areas have the best coverage of sediment and hydrological data (Table 2).

TABLE 2—Calculated rates of mass transport Φ for the central western shelf and Otago–Canterbury shelf based on actual and extrapolated values of the mean speed U_0, speed of the tidal flow U_T, and speed of swell and waves 1 m from the seafloor U_w. To demonstrate the influence that variation of the aforementioned values has on the magnitude of Φ, examples from N.W. Nelson and S. Canterbury are used, e.g., Φ for S. Canterbury is calculated for normal mean flow and for high rates of flow when the mean is strongly influenced by prolonged winds blowing from the south (Hydrographic Branch 1952b). Other notations are A, T, d, representative values of the swell amplitude and period, and depth of water; θ_1, θ_2 the cut-off times for sediment erosion; U_{100c} the threshold mean speed; $\frac{\tau_0 - \tau_c}{\tau_c}$ the ratio of the bottom stress, with τ_0 the actual stress, and τ_c the minimum stress to produce erosion; $K\left(\frac{\tau_0 - \tau_c}{\tau_c}\right)$ the appropriate ratio of the power expended into sediment transport; ρ_s sediment density; ϕ_1, ϕ_2 sediment transport rates with and without swell respectively; L shelf width (* = U_{100c} not exceeded).

POSITION	DOMINANT SEDIMENT	U_{100c} (cm.s^{-1})	U_0 (cm.s^{-1})	U_T (cm.s^{-1})	U_w (cm.s^{-1})	A (m)	T (s)	d (m)	θ_1 (deg.)	θ_2 (deg.)	$\frac{\tau_0-\tau_c}{\tau_c}$	K	ρ_s	ϕ_1 (g.cm^{-1}.s^{-1})	ϕ_2 (g.cm^{-1}.s^{-1})	L (km)	Φ (ton.y^{-1})
CENTRAL WESTERN SHELF																	
Taranaki	Fine sand	35	1.5	20	11	2.5	14	95	*	*	–	–	2.5	0	0	100	0
W. Cook Strait	Med. sand	40	10	50	10	8	11	110	37	*	2 06	0 02	2 5	$1\ 10\times10^{-2}$	$1\ 12\times10w^{2}$	50	1.7×10^{6}
N.W. Nelson	Med. sand	40	1.5	51	0	–	–	–	41	35 5	0.72	0 005	2 5	1.16×10^{-4}	No swell	50	1.7×10^{4}
		40	1.5	51	5	4	11	110	31.5	24.5	1.06	0.01	2.5	1.12×10^{-4}	$2\ 9\times10w^{4}$	50	4.5×10^{4}
OTAGO–CANTERBURY SHELF																	
S. Canterbury	Fine sand	35	5	25	23	1.5	11	37	*	*	–	–	2.5	0	0	50	0
		35	50	25	23	1.5	11	37	71	*	6 8	0.08	2.5	3.23×10^{-2}	4.13×10^{-2}	50	6.4×10^{6}

CENTRAL WESTERN SHELF

Textural and mineralogical criteria indicate sediment is transported along the Westland and Nelson shelf to the northeast, with at least part of the load moving into western Cook Strait (Fig. 5; Furkert 1947, van der Linden 1969). The bedload is mainly fine and medium sand, and hence would require near-bottom current speeds in excess of about 35–40 $cm.s^{-1}$ to initiate grain movement. Van der Linden (1969) regarded the Westland Current as the main transporting agent, but this current and its offshoot, the D'Urville Current, have low mean speeds ranging between 1.3 and 9.6 $cm.s^{-1}$ (Heath 1974a). By contrast, tides have maximum surface speeds of up to 51 $cm.s^{-1}$ (Table 2) and undoubtedly play a major role in initiating sediment movement. Superimposed on this circulation are effects of storm-driven waves and currents. Unfortunately wave and current data are sparse, but the few measurements at hand (Watters 1953, McIntosh 1958) suggest that the prevailing south-westerly storms are capable of stirring sediment at least on the inner and probably on the middle shelf. Maximum sediment transport along the Westland–Nelson shelf can be expected when south-westerly storms are coincident with peak tidal flow, which is to the northeast. Further, as pointed out above, the transport will be much greater when all three components reinforce one another, i.e., in the direction of the mean circulation (Westland Current).

Within Cook Strait, tidal forces prevail as a result of tidal constriction between the North and South Islands and of the time delay between high tide on one side of the Strait and on the other (Heath 1974b). Near-bottom tidal speeds of up to 50 $cm.s^{-1}$ (Gilmour 1960) are certainly capable of moving medium to fine sand, as attested by underwater photographs of active current ripples (Estcourt 1968) and by the sediment-tracer experiment of Pantin (1961). Again, additional impetus for sediment transport will be provided by storm-driven winds and waves, for which the Cook Strait area is notorious (Garnier 1958).

Further north, on the broad Taranaki shelf, effects of tides are dissipated, although tides are still probably a major force, having maximum surface speeds of up to 21 $cm.s^{-1}$ (Heath 1974a). The mean circulation is poorly defined, but it appears that the Westland Current extends to just south of Cape Egmont, where an offshoot swings to the south and follows the Taranaki coast. Judging by low speeds of the Westland Current off northwest Nelson (1.3–9.6 $cm.s^{-1}$), it is probably a minor component of the velocity field. Wave data from oil drilling rigs off Cape Egmont show that waves create bottom speeds of up to 11.1 $cm.s^{-1}$ at 95 m depth (Table 1), which is insufficient to stir sand in the area of recording, but will move sand in shallower sections ($<$70 m) of the shelf. The records embrace a period of only 2 y, whereas hindcast wave statistics compiled by Glenn *et al.* (unpublished 1973) for the same area suggest that waves generated during the maximum 25-y storm are capable of creating bottom speeds of 35 $cm.s^{-1}$ at 130 m depth.

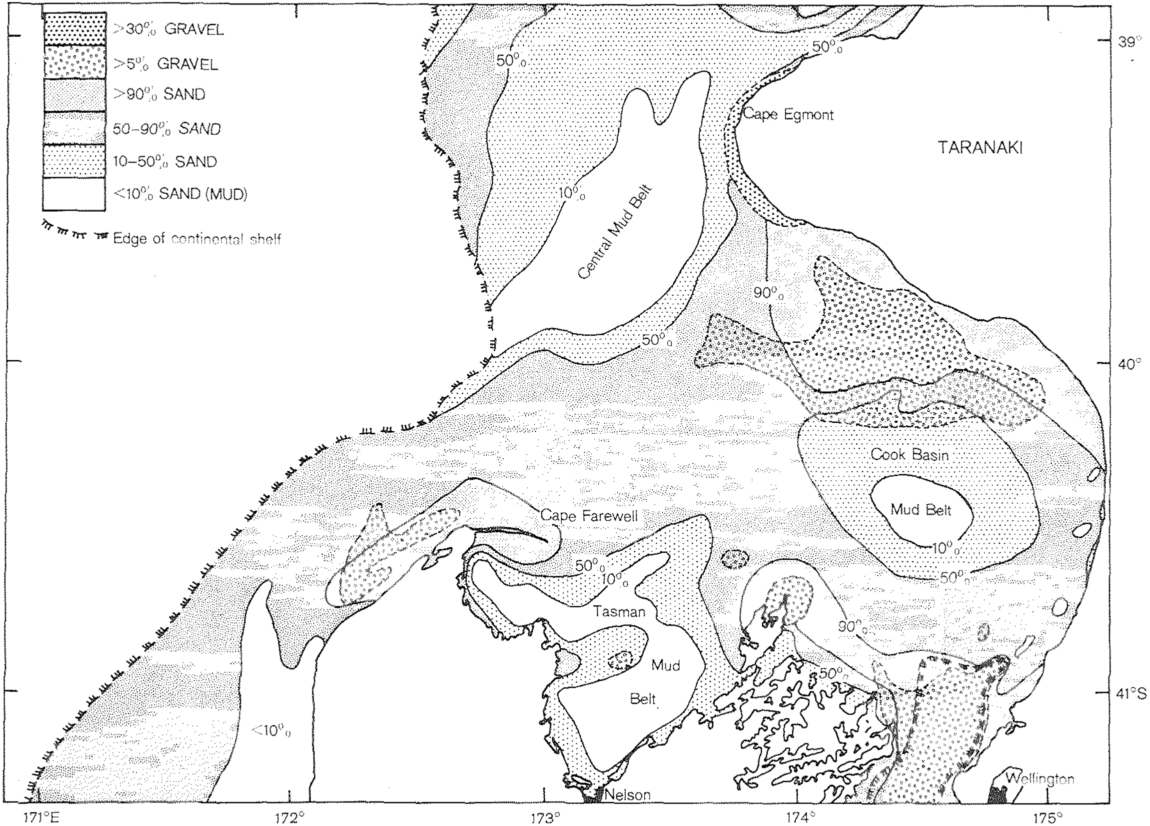

FIG. 5—Distribution of sediment types on the central western shelf of New Zealand. Modified from Lewis & Eade (1974).

OTAGO–CANTERBURY SHELF

The north-eastward movement of sand along the Otago–Canterbury shelf has been established on the basis of mineralogic criteria (Marshall 1905, Bardsley unpublished 1972) and textural parameters (Andrews 1973, Carter & Ridgway 1974, R. H. Herzer, NZOI pers. comm.). Direct evidence of the northeast-trending bottom current comes from recently completed seabed drifter experiments (R. H. Herzer, NZOI, pers. comm.). Woodhead drifters (Woodhead & Lee 1960) were released across the shelf off Banks Peninsula, from where they moved mainly northeast for distances of up to 120 km before retrieval.

Andrews (1973) attributed sediment movement to the Southland Current, but this current has a *surface* speed of only 7–8 $cm.s^{-1}$ (Heath 1972). By contrast, tides have far higher speeds, such as the 51 $cm.s^{-1}$ for the north-flow flood tide off Otago (Hydrographic Branch 1952a). Sediment movement is also strongly influenced by storm-driven components, which are mainly from the southwest. Prolonged south-westerly winds may increase surface current speeds by as much as 50 $cm.s^{-1}$ (Hydrographic Department 1958). Bottom surges created by grounding swells have speeds sufficient to stir sediment on the middle shelf (Carter & Ridgway 1974, and oil rig data, Table 1).

It would appear, then, that sediment movement on this shelf is initiated by tides and to a lesser extent by storm-driven components, with the mean circulation (Southland Current) playing a minor role. After sediment movement has been initiated, the direction of transport is probably controlled by storm-driven components associated with south-westerly storms.

DISCUSSION AND CONCLUSIONS

REGIONAL VIEW OF SEDIMENT TRANSPORT

Over much of the South Island shelf, bedload transport is along the shelf to the northeast (Fig 6). Although this direction is coincident with that of the Westland and Southland Currents, these currents by themselves have insufficient impetus to move fine sand, the main forces being tides and wind and storm-driven components. Between the main islands, in Cook and Foveaux Straits, sediment transport appears to be back and forth over the shelf in response to the strong tidal components in these areas. What little is known of the North Island shelf regime suggests sediment movement is mainly influenced by tides and storm-driven components, with the mean circulation again playing a minor role. In fact, off northernmost New Zealand, sediment dispersal is opposite to the mean circulation currents, i.e., the West Auckland and East Auckland Currents (Summerhayes 1969a).

Geologically determined dispersal patterns only indicate the direction of the overall sediment transport vector. In reality, sediments probably travel irregular paths over the continental shelf, especially since a major force behind the movement appears to be tidal in character, e.g., paths traced by an artificial sand (concrete plus iron sand) in the tide-dominated Cook Strait area, were backwards and forwards, presumably in response to different phases of the tidal cycle (Pantin 1961). Most shelf sands around New Zealand do, however, show a noticeably directional dispersal trend. Such a directional force would obviously be more noticeable if the main driving forces associated with sediment transport were combined in the same direction, as is envisaged for the Otago–Canterbury shelf where a combination of south-westerly storm conditions, the north-east-trending Southland Current and tidal current result in a north-eastward movement of sediment.

It may be argued that some dispersal patterns are relict, having developed during some lower stand of sea level, when presumably there would be more energy available for sediment transport mainly through increased wave energy in the shallower depths. Indeed it is likely the dispersal patterns of coarse sands and gravels are relict, for speeds of present day water motions are too slow to move such grain sizes except in the nearshore surf zone. However, these water motions can move finer bedload material, and consequently the patterns displayed by medium and finer sands are probably modern.

The quantity of bedload moved is very much dependent on the hydraulic regime. Under calm conditions it is unlikely that much bedload is moved except in areas of strong tidal flow. In combination with storm-induced motions, however, the transporting capability of the mean circulation and tides is greatly enhanced. Storm waves are probably responsible for stirring sediment over much of the shelf in a manner similar to that described for the Gulf of Mexico (Curray 1960), the western Canadian shelf (Carter 1973), and the western European shelf (Hadley 1964, Draper 1967). Because of the oscillatory motion of the waves, they are not major transporting agents in themselves. Their main effect is to stir sediment, which once in suspension requires less force (supplied by tides, mean circulation, or other horizontal moving components) to move along the shelf. Storms producing waves and swell capable of stirring sediments over the entire shelf are not annual events. Wave data (Table 1) from off Cape Egmont show that speeds capable of moving bedload were not attained on the middle shelf during the 2-y period of recording. Hindcasts by Glenn *et al.* (unpublished 1973), however, indicate that such speeds could be attained during the maximum 25-y storm.

The envisaged situation is, then, one of minor sediment transport under non-stormy conditions, greater sediment movement during annual storms particularly on the inner to middle shelf, and greatest sediment transport during maximum 25-y storms, when waves are of such a magnitude as to create sediment-stirring bottom surges at most shelf depths.

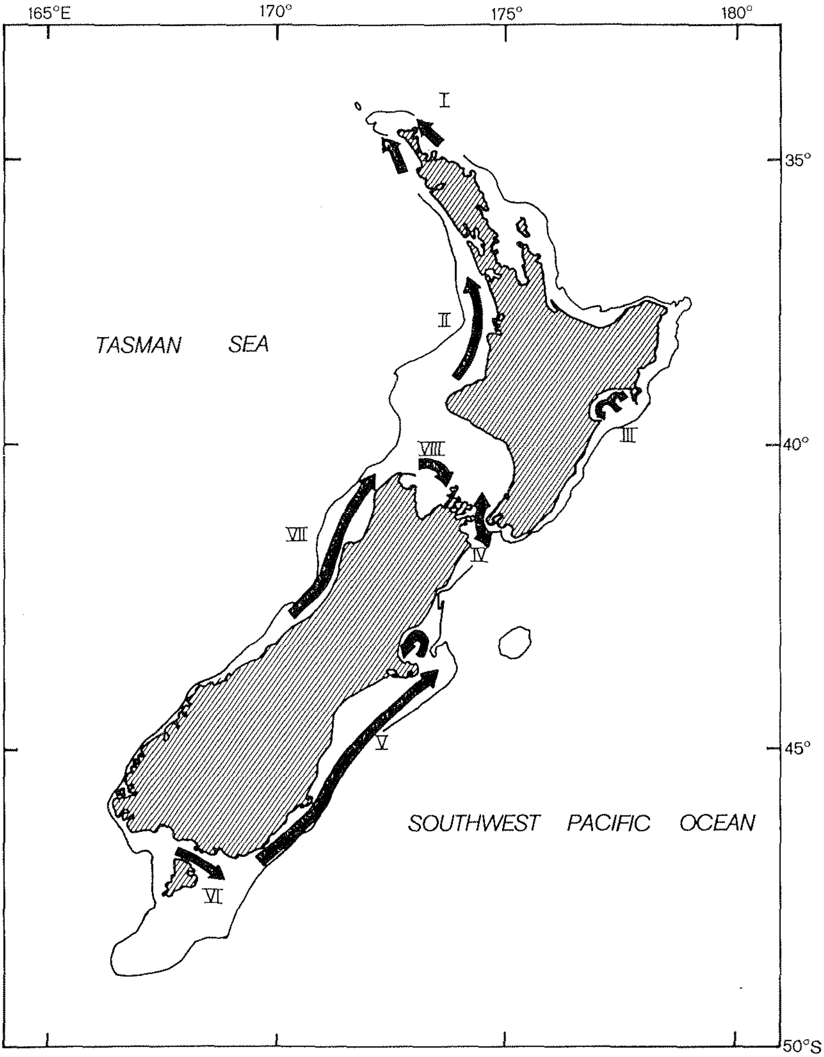

FIG. 6—Directions of sediment transport over the continental shelf based on the present study and the works of (I) Summerhayes (1969a); (II) McDougall & Brodie (1967); (III) Pantin (1966) and Ridgway & Stanton (1969); (IV) Pantin (1961); (V) Andrews (1973), and R. H. Herzer (pers. comm.); (VI) Cullen (1967) and Heath (1973a); (VII) Furkert (1947), and van der Linden (1969); (VIII) Reed & Leopard (1954), and van der Linden (1969).

MUD TRANSPORT

This paper has concentrated on the transport of sand and gravel rather than mud (silt and clay), because published studies of sediment dispersal deal mainly with the coarser sediments, and because the dynamics of mud transport are poorly known. Nevertheless, the abundance of mud over the New Zealand continental shelf necessitates comment on these sediments, although a paucity of data forces some comments to be speculative.

The presence of mud on the shelf is dependent upon the hydraulic regime and sediment supply. In areas receiving high volumes of terrigenous sediment, such as off Westland and Hawkes Bay–Wairarapa, mud blankets nearly all the shelf except shallow bank tops and the nearshore surf zone (see Pantin 1966, Lewis 1973; van der Linden & Norris 1974). It is suggested that the sediment input into these areas exceeds sorting capabilities of the shelf hydraulic regime, and consequently mud prevails. By contrast, on the Otago–Canterbury shelf, the hydraulic regime apparently can cope with the mud load, which largely appears to have moved beyond the shelf to accumulate on the continental slope and beyond. Exceptions occur on the downcurrent sides of major promontories, e.g. Otago and Banks Peninsulas, where muds are entrapped by suspected eddy systems (see Andrews 1973; R. H. Herzer, NZOI pers. comm.).

Much of the mud reaching the shelf is riverborne and, once entering the sea, part of this load may flocculate out and accumulate nearshore as suggested by Lewis (1973) and Scruton & Moore (1953). Being nearshore, the floccules are vulnerable to resuspension by grounding waves, as demonstrated by Scruton & Moore (1953), who recorded a hundred-fold increase in turbidity near the Mississippi Delta during storms. The resuspended sediment, as well as grains of non-floccule character that have remained in suspension, would then be readily transported by the mean circulation, tides, and storm-driven components. As with bedload, the directions of transport appear to be along the shelf, e.g., Lewis (1973) records a southerly transport of mud parallel to the Hawkes Bay coast in a response to the mean circulation and probably to tides; mud dispersal patterns on the Westland–Nelson shelf are coincident with the direction of main water movements (van der Linden & Norris 1974).

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