0278-4343(94)00068-9

Suspended sediment transport and morphological response on a dissipative beach

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(Received 14 February 1994; accepted in revised form 23 June 1994)

Abstract—This paper reports on some results from a field experiment conducted under highly dissipative surf zone conditions, occurring during a storm in Lake Huron, Ontario, Canada. Measurements of sediment resuspension and sediment flux were conducted at a number of stations across a nearshore bar. This bar moved approximately 25 m offshore during the storm, at a rate of $\approx 2.5 \text{ m h}^{-1}$. The maximum erosion depth over the former bar crest was 0.92 m, with simultaneous accretion on the lakeward slope of the bar reaching 0.63 m. During the early hours of the storm, the accretion rate over the lakeward slope was approximately 0.05 m h^{-1} .

Mean sediment concentrations were up to 6.4 g l^{-1} , time series of sediment concentration were characterized by low-frequency fluctuations and the sediment resuspension was strongly constrained by infragravity waves. The major mechanism responsible for the offshore sediment flux was an offshore directed mean current, which reached a velocity of -0.34 m s^{-1} . However, the reason for the erosion and the offshore migration of the bar was the strong spatial gradient in the cross-shore suspended sediment flux. This gradient was induced mainly by the infragravity waves, which transported sediment onshore at the former bar crest where the oscillatory flux balanced the mean flux, and offshore over the lakeward slope, reinforcing the mean flux at this location.

INTRODUCTION

Measurements of suspended sediment concentrations in wave-dominated coastal environments have often been conducted using traps (e.g. Kraus *et al.*, 1981; Kraus, 1987; Antsyferov and Kosyan, 1990) or suction samplers (e.g. Nielsen *et al.*, 1982; Nielsen, 1984; Bosman, 1986). These methods provide estimates of time-averaged sediment concentrations at discrete heights above the bed, which, multiplied by mean current velocities, have been used to calculate (mean) suspended sediment transports. Wave motions have been perceived as a stirring mechanism, making the sediment available for transport by any mean currents (e.g. Bagnold, 1963; Dally and Dean, 1984; Stive and Battjes, 1985).

Within the past decade, the advent of fast-response optical (OBS; Downing *et al.*, 1981) and acoustic backscatter sensors (Hanes *et al.*, 1988; Vincent *et al.*, 1991; Hay and Sheng, 1992) have increased our ability to investigate the processes of sediment resuspension and transport in natural environments. Recent studies have shown that the mechanism of resuspension indeed depends on oscillatory wave motions, with the configuration of the

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bed as an important secondary control. Measurements have usually been conducted outside the surf zone and have shown that over a rippled bed, the suspension of sediment occurs as discrete events on a time scale comparable to the time scale of the incident waves (e.g. Sternberg *et al.*, 1985; Hanes and Huntley, 1986; Doering and Bowen, 1988; Green and Vincent, 1990; Osborne and Greenwood, 1992a,b). These events are due to the ejection of sediment laden vortices from the ripples into the water column (Vincent and Green, 1990; Osborne and Greenwood, 1993), a process which has also been observed in the laboratory (e.g. Tunstall and Inman, 1975; Sleath, 1982; Blondeaux and Vittori, 1990). However, when bedform steepness declines, the sediment concentration at a given elevation above the bed decreases due to the less intense vortex shedding (Vincent *et al.*, 1991). Under "flat bed" conditions, sediment resuspension is more highly influenced by diffusion. A further complication to the prediction of sediment concentrations outside the surf zone is the presence of wave groups. Concentrations have been shown to grow and decay on a time scale corresponding to these groups; the fluctuations have been attributed to the inherited turbulence effect associated with the wave group (e.g. Doering and Bowen, 1988; Hanes, 1990; Osborne and Greenwood, 1992a,b, 1993).

Under energetic conditions within the surf zone, major suspension events may occur on a time scale much longer than the incident waves (e.g. Downing, 1985; Beach and Sternberg, 1988, 1991; Aagaard and Greenwood, 1994, in press). In this case, the long time scales are not due to wave groups, as the groups decay shoreward of the breakpoint. Instead, they are a response to the presence of infragravity waves within the surf zone itself.

Wave-induced oscillatory currents can also be extremely important to the transport of sediment. Owing to the "flux coupling" between (oscillatory) current velocity and sediment concentration (Jaffe *et al.*, 1985), wave motions may give rise to a directed net suspended sediment transport. Jaffe *et al.* (1985) documented the importance of wave motions in causing an onshore migration of a nearshore bar, even in the presence of offshore directed mean flows, and they demonstrated that the net cross-shore transport of suspended sediment ($\langle q_s \rangle_n$) can be computed as:

$$\langle q_s \rangle_n = 1/n \sum_{uc} = \langle u \rangle \langle c \rangle + \langle u'c' \rangle \quad (1)$$

where the term $\langle u \rangle \langle c \rangle$ is the sediment transport accomplished by mean currents, computed as the product of the time-averaged velocity and the time-averaged sediment concentration. The magnitude of the oscillatory term $\langle u'c' \rangle$ depends on the correlation between fluctuations in concentration and oscillatory velocity.

Recently, a number of studies have been conducted employing such techniques. The majority of these studies have taken place outside the surf zone and they show that incident waves tend to transport the sediment shoreward, this transport to some extent being balanced by offshore mean flows and offshore oscillatory transport at infragravity frequencies (Huntley and Hanes, 1987; Doering and Bowen, 1988; Osborne and Greenwood, 1992a,b). The offshore infragravity transport is due to the bound long wave troughs (Longuet-Higgins and Stewart, 1964; Shi and Larsen, 1984) occurring beneath groups of large incident waves.

However, it has also been demonstrated that there is a strong vertical dependence in the transport associated with incident waves, and that this transport may change direction with height due to increasing phase lags between velocity and concentration with increasing elevation above the bed. These phase lags are due to the aforementioned generation of

sediment laden vortices over rippled beds (Vincent and Green, 1990; Vincent *et al.*, 1991; Osborne and Greenwood, 1993), or to the progressively increasing amounts of sediment lifted into the water column by the action of wave groups (Greenwood *et al.*, 1991; Osborne and Greenwood, 1993).

Only a few experiments have been conducted within the surf zone and measurements have usually been at a single point within this zone (Beach and Sternberg, 1988, 1991; Russell *et al.*, 1991; Osborne and Greenwood, 1992b; Davidson *et al.*, 1993). Whereas transport patterns are relatively well documented for the region seaward of the break-point, sediment transport appears considerably more complex within the surf zone. The oscillatory transport due to incident waves decreases due to wave breaking; mean currents gain increasing importance and are generally directed offshore, whereas oscillatory transport at infragravity frequencies tends to become increasingly important as energy levels increase. Infragravity transports display a spatial dependency, with directions and magnitudes apparently being related to the cross-shore structure of standing waves (Aagaard and Greenwood, 1993, 1994). Motions at such frequencies appear to assume significance under storm-induced dissipative conditions and with proximity to the beach face; they may become the major sediment transport mechanism in the inner surf and swash zones (Beach and Sternberg, 1988, 1991; Russell *et al.*, 1991; Russell, 1993).

Owing to the fact that most previous studies have relied on a single instrument station, cross-shore gradients in sediment dynamics and sediment transport are poorly documented. Aagaard and Greenwood (1994) reported on suspended sediment transport across a nearshore bar in a moderate storm during which the bar migrated shoreward. The majority of the transport was accomplished by mean currents and standing infragravity waves at discrete frequencies, the latter transporting the sediment towards surface elevation antinodes. Onshore directed mean currents and oscillatory infragravity transports over the bar crest appeared to be the reason for the onshore bar migration. During intense storm conditions, bars tend to migrate offshore (e.g. Sallenger *et al.*, 1985); however, empirical linkages between bar migration and a measured sediment transport have so far not been documented. In this paper, we document the migration of a nearshore bar and the associated suspended sediment flux, which occurred under highly dissipative surf zone conditions during a single storm event.

FIELD SITE AND INSTRUMENTATION

The experimental results reported here form part of the Canadian Coastal Sediment Transport programme (C-COAST; Greenwood *et al.*, 1990, 1991). The field site was located at Burley Beach within Pinery Provincial Park, Lake Huron, Canada (Fig. 1). This beach is exposed to winds and waves from directions between SW and N with relatively long fetches particularly towards the NNW of up to 300 km. Wave hindcasting for Kettle Point, immediately south of the study site, suggests a mean annual wave height of 0.9 m with a period of 3.4 s (Reinders *et al.*, 1989) with $H_{10\%}$, $T_{10\%}$ being 2.0 m and 5.8 s, respectively. Tides are negligible in this environment. Burley Beach is therefore dominated by wind waves, particularly from the west and northwest, while some swell may persist for short intervals after storms. The net longshore transport is directed towards the south and presently estimated at $68,000 \text{ m}^3 \text{ a}^{-1}$ (Reinders *et al.*, 1989). Owing to the artificial outlet of the Ausable River and the headland at Kettle Point (Fig. 1), the majority of this sediment accumulates at or close to the study site and large dunes have formed

Fig. 1. Location of Burley Beach, Pinery Provincial Park, Lake Huron, Canada.

within the park. The mean grain size of the sand in the nearshore zone is $180\text{--}200\text{ }\mu\text{m}$, and the profile is gently sloping with a mean gradient of $\beta = 0.009$. The shoreface exhibits three longshore bars of which the third (and outer) is very low and broad (Fig. 2).

Five profile lines were established, numbered Lines 140S, 60S, 0, 60N and 140N, with

Fig. 2. Shoreface profile, Line 0 (the instrument transect) surveyed on 24 September 1992.

Fig. 3. Location of the instrument stations B-I–B-IV at $x = 80, 90, 100$ and 127 m. H marks the location of the mobile sonar (HRRTSIII) at $x = 111.5$ – 116.5 m.

Line 0 being the sensor deployment line. These lines were surveyed before and after storms using a Geodimeter Total Station and a Raytheon echo-sounder. The main feature of the experiment was the deployment of a High Resolution Remote Tracking Sonar (HRRTS III; Richards and Greenwood, 1993), approximately 115 m from the baseline on the lakeward slope of the second bar (Fig. 2), with the sensor head at $z = 0.54$ m prior to the experiment. This sensor was used for measuring the two-dimensional bedform geometry as well as the mean bed elevation through time. At the four corner posts of the H-frame supporting the HRRTS (ranging from $x = 111.5$ – 116.5 m), as well as in the middle of this box, five sensor stations were deployed in order to measure the sediment flux. These stations each consisted of an electromagnetic current meter (Marsh-McBirney OEM512) and three fast-response (10 Hz) optical backscatterance sensors (D & A Instrument Co OBS-1P). The current meters were deployed at nominal elevations of $z = 0.4$ m above the bed with the optical sensors at $z = 0.05, 0.10$ and 0.15 m.

A further four sensor stations, numbered B-I through B-IV (Fig. 3) were deployed in a cross-shore array across the second bar, at $x = 80, 90, 100$ and 127 m from the baseline, respectively, with the distance from the baseline to the shoreline being approximately 20 m. These stations each consisted of a current meter and one or two optical backscatter sensors. Here, the current meters were at nominal elevations of $z = 0.47$ m (80 m), 0.36 m (90 m), 0.25 m (100 m) and 0.28 m (127 m), respectively. All current meters were oriented with positive axes onshore (u) and to the northeast (v). Furthermore, the 90 m station was equipped with a pressure transducer. Before and after the experiment, the current meters were calibrated in a large tow tank, while the OBS-sensors were calibrated in a Sediment Recirculating Facility using sand from the field deployment locations. Owing to a rather large background turbidity at the field site (due to fine particulates from the Ausable River mouth at Port Franks), OBS field offsets were determined as the 10th percentile of the sample values in each record. This magnitude was chosen because the cumulative distribution curve of the sample values usually exhibited a distinct break around this value. Such a procedure may introduce some error in the determination of absolute sediment transport rates, but not in the relative rates with which we are mainly concerned. All sensors were hardwired to an underwater data acquisition system (UDATS; Hazen *et al.*, 1987), and recorded at a rate of 4 Hz for periods of 55 min, with breaks of 5 min, throughout the storm. Winds were recorded at the top of the foredune using a Young Wind Monitor, part of a Portable Air Quality Monitoring Station (PAQMOS).

DATA ANALYSIS

Instrument records were truncated to 8192 sample values, corresponding to 34.1 min of data. The output from each channel was plotted to check data quality and records which contained obvious errors or noise were discarded. During the storm discussed here, morphological changes were large which resulted in a collapse of the instrument stations at B-I and B-II within the first 4–5 h of the event, while the sensors lakeward of B-III were progressively buried by sand. Therefore, only sediment flux results from the early part of the storm are presented in this paper.

Univariate statistical descriptions of velocity and sediment concentrations (mean, standard deviation etc.), as well as spectral and cross-spectral characteristics were computed using the BMDP/386 statistical software package. All spectra were computed using 50 degrees of freedom (df). Maximum oscillatory currents were estimated as $u_m = 2.8 u_o$, where u_o^2 is the vector sum of the record variance. Sediment flux rates were determined using equation (1). In this equation, the term $\langle u'c' \rangle$ can be computed using cross-spectral analysis techniques (Huntley and Hanes, 1987), as the cospectrum yields the cross-product of the in-phase components of velocity and concentration, as a function of frequency. The local net oscillatory suspended sediment transport rate can therefore be calculated from

$$\langle u'c' \rangle = \langle q_s \rangle_{osc} = \frac{\Delta f}{F} \int_0^T C_{uc(f)} dt \quad (2)$$

where $C_{uc(f)}$ is the cospectral variance at a given frequency, Δf is the resolution of the cospectrum, F is the cospectral frequency range and T is the record length. The cospectrum thus describes the contribution of different frequency bands to the direction and rate of the oscillatory suspended sediment flux.

THE STORM EVENT

The field experiment was conducted during the months of September–November 1992. This period was dominated by the passage of several low-pressure systems across Lake Huron. The event discussed here occurred on 16–17 October and caused significant morphological adjustment and specifically nearshore bar migration. The storm was associated with an intense low-pressure system which generated southwesterly and westerly waves, which in turn set up strong longshore currents owing to the oblique wave incidence. The event began rather suddenly during the morning of 16 October. Initially, winds were light and offshore, but they veered southwest and increased rapidly up to 17–20 m s^{-1} ; winds slowly decreased over the next 18–20 h (Fig. 4). Maximum offshore significant wave heights were approximately 2 m. Owing to the gently sloping shoreface, waves broke well seaward of the outer (third) bar and the surf zone became very dissipative with up to 9–10 bores observed simultaneously. Virtually no wave reformation occurred. Maximum recorded orbital velocities at the HRRTS-station reached 1.75 m s^{-1} , but were almost certainly depth limited; increasing values of u_m were correlated with increasing water depths as a result of wind and wave setup. Mean longshore and cross-shore current velocities reached +1.02 m s^{-1} and –0.34 m s^{-1} , respectively, at 1930 h on 16 October. The cross-shore currents were consistently directed offshore. Note that the

Fig. 4. Wind speed (W), maximum orbital velocity (u_m), mean cross-shore current ($\bar{u}$) and mean longshore current ($\bar{v}$) recorded during the storm on 16–17 October 1992. Velocities were measured on the lakeward slope of the second bar.

mean currents reached their maximum several hours prior to the storm peak, here defined by the maximum values of u_m which occurred at 0100 h, 17 October.

Figure 5 illustrates spectra of the cross-shore current, recorded by a current meter at the HRRTS-station (116.5 m) on 16 October. During the early hours of the event, presented by the spectrum recorded at 1300 h, incident waves of 0.2 Hz (5 s) and a local significant wave height of 0.73 m occurred. The period of the incident waves increased subsequently and by 2000 h the spectral energy peaked at approximately 0.14 Hz (7 s). A distinct feature of these spectra is the narrow-banded peak seen at a frequency of 0.009 Hz (the frequency varied slightly between 0.0087 and 0.0098 Hz; 102–114 s). This represents the peak frequency of the infragravity wave field recorded by the sensors in the array. In contrast to previous reports (e.g. Bauer and Greenwood, 1990; Aagaard *et al.*, 1994), infragravity waves were important at the very beginning of the event and their peak frequency

Fig. 5. Cross-shore velocity spectra recorded at the 116.5 m station on 16 October 1992. Time series were 34.1 min long. Note: np = number of sample values, df = degrees of freedom.

appeared to be rather stable through the rising limb of the storm. Another current meter, which was deployed some 5 m landward of the HRRTS kept working throughout the event. This instrument showed that after ≈ 2200 h (16 October), the infragravity band became significantly broader.

MORPHOLOGICAL CHANGES

Figure 6 illustrates the profile response along the instrument transect (Line 0). In effect, the second bar which was sinuous prior to the storm was eroded and a straight bar was formed further offshore. At the pre-storm bar crest up to 0.92 m of erosion occurred

Fig. 6. Shoreface profile change during the storm, measured along the instrument transect. The dashed line is the pre-storm survey, while the solid line is the post-storm survey.

Table 1. Net volumetric changes in the second bar–trough system measured in $\text{m}^3 \text{m}^{-1}$ of shoreline length

Profile	Erosion ($\text{m}^3 \text{m}^{-1}$)	Accretion ($\text{m}^3 \text{m}^{-1}$)	Balance ($\text{m}^3 \text{m}^{-1}$)
140S	14.91	11.95	−2.96
60S	17.97	18.20	+0.23
0	19.19	18.27	−0.92
60N	11.14	10.47	−0.67
140N	6.21	4.65	−1.56
Average	13.89	12.71	−1.18

(causing the collapse of the stations at 80 m and 90 m), while up to 0.63 m of accretion took place on the lakeward slope of the second bar (burying the instruments around the HRRTS). At the 111.5 m station (landward end of the HRRTS-track), 0.53 m of accretion occurred. Along the instrument transect which was surveyed immediately prior to the storm, the bar moved approximately 25 m lakeward. Table 1 documents the volume changes computed for the second bar–trough system from all five profile lines. The sediment budget was calculated using the intersection of the pre- and post-storm profile lines in this region, i.e. in the case of Line 0, between $x = 70$ and 160 m. Using this method, the average amount of erosion of the pre-storm bar crest was $13.9 \text{ m}^3 \text{m}^{-1}$ of shoreline, while $12.7 \text{ m}^3 \text{m}^{-1}$ accreted on the lakeward slope of the bar. The sediment budget for this profile segment was essentially in balance. While acknowledging the possible errors associated with the method of computation, it can be argued that the second bar was reorganized without any significant net sediment losses or gains to or from cross-shore or alongshore sources.

Profile changes appeared to occur very rapidly following the onset of the storm. Figure 7 illustrates the average bed elevation change as determined by the HRRTS, as a function of time. The sonar, which was deployed on the lakeward slope of the bar showed that accretion occurred extremely rapidly and essentially linearly from the onset of the event until around 2200 h on 16 October, a couple of hours after the time when mean cross-shore currents peaked (see Fig. 4), and approximately corresponding to the time when the infragravity wave band became broader. However, incident wave heights (as represented by u_m) were still increasing at 2200 h (Fig. 4). The initial accretion rate on the lakeward slope was $\approx 5 \text{ cm h}^{-1}$. Subsequent to 2200 h, the accretion rate dropped sharply to $\approx 1 \text{ cm h}^{-1}$, and by 0700 h (17 October) the bed showed some limited accretion ($\approx 10 \text{ cm}$) and then became stable. These results suggest that the bar approached a quasi-equilibrium position after 2200 h.

Assuming that the positions of the bar at 2200 h (16 October) and at the end of the event were more or less identical, and that Fig. 7 also represents a measure of the rate of erosion in the trough, as well as the speed of bar migration along Line 0, we can estimate that prior to 2200 h, the bar moved offshore (along this line) with a speed of $\approx 2.5 \text{ m h}^{-1}$, and an associated trough erosion rate of $\approx 1.5 \text{ m}^3 \text{m}^{-1} \text{h}^{-1}$. These figures are similar to the maximum rates measured on an ocean shoreface at Duck, North Carolina, as reported by Birkemeier (1985).

Fig. 7. Bed elevation changes recorded at the HRRTS station, 16 and 17 October 1992. The sensor head was at $z = 110$ cm. The asterisk marks the bed position, as recorded by divers prior to the storm.

SPATIAL GRADIENTS IN SUSPENDED SEDIMENT CONCENTRATION

In this highly dissipative environment, sediment suspension appeared more homogeneous spatially than in the moderately dissipative surf zone reported on by Aagaard and Greenwood (1994). This homogeneity persisted both horizontally and vertically and was probably due in part to the importance of low-frequency (infragravity) oscillations, which appeared to exert a major control on suspension events. Figure 8 illustrates 34.1 min time series of cross-shore velocity ($z = 0.32$ m), low-pass filtered cross-shore velocity and suspended sediment concentrations recorded on 1402 h, 16 October at 116.5 m. Pronounced infragravity oscillations are readily apparent in the low-pass filtered time series of cross-shore velocity. The time series of sediment suspension also clearly show a very strong low-frequency modulation, occurring on a time scale of approximately 1.79 min (i.e. 0.009 Hz). Moreover, the three time series of sediment concentration follow more or less identical trends with no discernible lag time vertically at least at this scale of resolution. The only major difference is that the absolute magnitudes of concentration decrease with increasing elevation above the bed. While suspension events at the incident wave time scale were still important, they were subordinate to the infragravity modulations, indicating that the infragravity motions exerted an important control on the process of sediment resuspension.

Comparing the time series of sediment concentration with the low-passed time series of cross-shore velocity (using a filter cut-off at 0.02 Hz) reveals that the majority of the large suspension events at this location were associated with the offshore phases of the

Fig. 8. Time series of (from bottom upwards) cross-shore current velocity (u), lowpassed velocity (u_{lg}), and sediment concentrations at $z = 0.05, 0.10$ and 0.15 m, recorded at the 116.5 m station, 1402 h, 16 October.

infragravity oscillations. However, a simple linear dependency between concentration magnitude and current velocity does not seem to exist.

Figure 9 illustrates sediment concentration time series recorded at 1402 h, 16 October from the stations at 80, 90, 111.5, 116.5 and 127 m, respectively. The optical sensor at the 100 m station was already buried at this time. These time series illustrate the increasing dominance of the infragravity component towards the shoreline. At the lakeward end of the array (127 m), low frequency oscillations are clearly visible; suspension events at the

Fig. 9. Time series of sediment concentration at the (from bottom upwards) 127, 116.5, 111.5, 90 and 80 m stations, 1402 h, 16 October.

incident wave time scale were equally important, however, at this position at the lakeward toe of the second bar. Suspension events occurring on infragravity time scales ($f < 0.067$ Hz; approximately the frequency of the spectral trough in the current records) accounted for 54% of the time series variance at this station. As incident wave energy decayed across the bar, the influence of wave motions at these frequencies decreased, while infragravity oscillations assumed progressively greater importance. The infragravity fraction of the fluctuations increased from 78% (116.5 m), 76% (111.5 m), to 84% (90 m) and 86% (80 m). Thus, at the latter position, only 14% of the time series variance was associated with the incident wave frequencies.

Corresponding spectra of sediment concentration are shown in Fig. 10(a). The spectra illustrate an unequivocal statistically significant peak (5%-level) around 0.009 Hz, while

Fig. 10. Variance density spectra of: (a) sediment concentration at 80, 90, 111.5, 116.5 and 127 m; and (b) cross-shore velocity at 90 m and 116.5 m, 1402 h, 16 October. Note: np = number of sample values, df = degrees of freedom.

small (mostly insignificant) peaks in the incident wave band are 1–2 orders of magnitude smaller. The spectra also reflect the increasing infragravity importance towards the shoreline (80 m), where the variance is almost an order of magnitude larger than at 127 m, even though the bed was eroding at the former and probably accreting at the latter, thus respectively increasing and decreasing the distance between the optical sensors and the bed. Suspension events at incident wave time scales are clearly relatively more important at 127 m and the effect of incident waves on sand resuspension decreases shoreward.

Figure 10(b) shows cross-shore velocity spectra from 90 m and the HRRTS-station (116.5 m). Over this distance, infragravity variance increased dramatically, while incident wave variance was almost constant. In a similar manner to the spectra of sediment

Table 2. Coherence and phase of suspended sediment concentrations and cross-shore current at 0.009 Hz; 16 October 1400 h

Sediment concentration	Coherence	Phase (degrees)
111.5 m/80 m	0.70	152
111.5 m/90 m	0.79	161
111.5 m/100 m	0.55	-2*
111.5 m/116.5 m	0.86	-10
111.5 m/127 m	0.60	-3
Cross-shore current	Coherence	Phase (degrees)
111.5 m/90 m	0.82	-2

*Records from 1300 h.

concentration, the cross-shore velocity spectra also exhibit infragravity peaks at 0.009 Hz. It is therefore probable that the suspension events at infragravity frequencies were constrained by these long waves.

Another important aspect of the time series of sediment concentration (Fig. 9) is the cross-shore spatial homogeneity at infragravity frequencies. Major suspension events are virtually identical in shape and time of occurrence at both 80 and 90 m on the one hand, and at 100, 111.5, 116.5 and 127 m on the other. Table 2 documents the coherence and phase between suspended sediment concentration measured at each sampling station, relative to that at 111.5 m, using the 0.009 Hz frequency band. While suspension events at 0.009 Hz were virtually in phase across the outer part of the array, a significant phase jump occurred between the 100 and 90 m stations; concentration magnitudes shifted by almost 180° of phase. All measured phases were highly significant; the 95% confidence level of zero coherence is 0.28 (Jenkins and Watts, 1968). This suggests that: (a) the role of bedforms in introducing phase lags was restricted at these long time scales; and (b) a radical change in the hydrodynamic forcing by the infragravity wave field occurred somewhere between 100 and 90 m. However, Table 2 also demonstrates that there was no phase shift of the cross-shore velocity (at 0.009 Hz) between 111.5 and 90 m. This lack of phase shift suggests that the infragravity waves at 0.009 Hz were standing in the cross-shore direction. A scenario which might fit the observations (i.e. zero-phase in cross-shore velocity; antiphase in sediment concentration) would be a standing wave field, the 0.009 Hz frequency having a node in surface elevation (but not an antinode) between 111.5 and 90 m, with sediment preferentially transported by oscillatory velocities towards antinodes located shoreward of 90 m and seaward of 111.5 m. Such an hypothesis was proposed by Aagaard and Greenwood (1994) to account for the direction of oscillatory suspended sediment transport at infragravity frequencies; in that study, the hypothesis was substantiated by measurements of the cross-shore infragravity wave surface elevation using pressure transducers. In the present study unfortunately only one transducer (at the 90 m station) was functional.

Figure 11 is an enlargement of a segment of some of the records from 1402 h, 16 October; from top to bottom the panels show the raw time series of cross-shore velocity, low-pass filtered cross-shore velocity and sediment concentration (nominally at $z = 0.05$ m) at 90 m, as well as sediment concentration (nominally at $z = 0.10$ m), low-pass filtered cross-shore velocity and raw time series of cross-shore velocity at 111.5 m. First, the figure again illustrates the relatively greater efficiency of incident wave frequencies in resuspending sediment at 111.5 m. This is partly because this station was located further lakeward (with consequently less incident wave dissipation), and partly because the nominal elevations of the optical sensors were no longer applicable. The OBS at 111.5 m was in reality close to $z' = 0.05$ m (the lowermost OBS, nominally at $z = 0.05$ m was becoming buried at this time), while the OBS at 90 m was considerably higher than $z = 0.05$ m owing to erosion of the bed. The small thickness of the incident wave boundary layer means that suspension at these frequencies is more conspicuous close to the bed (e.g. Wright *et al.*, 1994). The figure also highlights the zero phase lag between the current velocities at low frequencies and the 180° phase shift between the suspension time series. Finally, the velocity time series illustrate the strongly asymmetric shape of the surf bores, particularly at 90 m.

An important detail is the fact that suspension events appear to precede the maximum current velocity associated with the low-frequency oscillations. In general, major suspension events are initiated prior to velocity reversals in the low-passed velocity records. In this particular case, the lead time for event initiation appears to be some 12–15 s, i.e. a phase lead of approximately 45 – 55° , at a frequency of 0.009 Hz. Cross-spectral analysis of the entire record showed that the phase lead of concentration magnitude relative to velocity magnitude was 31° and 13° , respectively, at 90 m and 111.5 m. Allowing for a short, but finite time for the sediment to be transported vertically to the elevation of the sensors, this figure compares favourably with the phase lead of bed shear stress, relative to free stream velocity which is 45° for laminar flows and somewhat smaller for turbulent flows (e.g. Nielsen, 1992). Jonsson and Carlsen (1976) recorded phase lags of 18 – 31° for rough turbulent flows. These observations suggest that the major suspension events occurring on a time scale of approximately 100 s in Figs 8 and 10 could be due to infragravity wave-induced shear stresses. However, the fact that additional phase lags with increasing elevation were small weakens this argument. The OBS-sensors located 0.05 m higher in the water column recorded phase lags of 33° and 17° , respectively, at 90 m and 111.5 m. Another possibility could be that suspension events were linked to the surface wave structure of the infragravity waves. In the case of depth-limited wave breaking, long wave crests would tend to increase local incident wave heights and orbital velocities (e.g. Abdelrahman and Thornton, 1987). This hypothesis is more likely as shear stresses associated with incident waves are considerably larger than the stresses associated with low-frequency waves. Long-wave modulation of orbital velocities at incident wave frequencies is not readily visible in the time series shown in Fig. 11. To investigate this in more detail, these time series were high-pass filtered and separated into half-cycles associated with long wave surface elevation crests and troughs. At the 111.5 m station, significant orbital velocities at incident wave frequencies were 0.1 – 0.2 m s $^{-1}$ larger under long wave crests than under long wave troughs. At the 90 m station, the difference was on the order of 0.1 m s $^{-1}$. Even though these differences are small, it seems that long-wave modulation of incident wave velocities was responsible for at least part of the concentration modulation at infragravity wave timescales.

Fig. 11. Time series of velocity and concentration recorded on 16 October, 1402 h. From bottom upwards: cross-shore current velocity (u), lowpassed velocity (u_{ig}) and sediment concentration ($z = 0.10$ m; $z' = 0.05$ m), all at the 111.5 m station; sediment concentration ($z = 0.05$ m; $z' = 0.14$ m), low-passed velocity (u_{ig}) and cross-shore velocity (u), at 90 m. z is nominal elevation, z' is true elevation.

Fig. 12. Cospectra of u and c recorded at 90 m (a) and 111.5 m (b) at 1402 h, 16 October. True elevations (z') were 0.14, 0.19 m (90 m), and 0.00, 0.05 and 0.10 m (111.5 m), respectively. The cospectra have 50 degrees of freedom. Note: the low frequency peaks at ≈ 0.009 Hz are statistically significant at the 5% level.

SPATIAL GRADIENTS IN SUSPENDED SEDIMENT FLUX

Oscillatory suspended sediment flux was computed using cross-spectral analysis (Huntley and Hanes, 1987). Cospectra from the pre-storm bar crest (90 m) and the lakeward slope (111.5 m) are presented in Fig. 12(a) and (b). These spectra are representative of the early part of the storm event, when sediment concentration measurements were also available (i.e. until approximately 1800 h, 16 October). The elevations given for the OBSs in the figures are nominal elevations. In Fig. 12(b), the cospectrum for the lower OBS has also been presented, even though this sensor was clearly becoming buried at this time, as witnessed by an increasing mean concentration; the oscillatory nature of the concentration

series was maintained, however. These spectra illustrate the dominant influence of infragravity frequencies in the oscillatory sediment flux, with large cospectral peaks being centered at approximately 0.009 Hz. As might have been expected from the foregoing discussion, the flux at infragravity frequencies was in opposite directions at the two locations. On the crest of the bar [Fig. 11(a)], infragravity motions transported suspended sediment onshore, while an offshore flux occurred on the seaward slope of the bar [Fig. 12(b)]. In contrast, suspended sediment flux due to incident waves was significantly smaller and consistently directed onshore as shown by the frequencies between 0.15 and 0.2 Hz. Sediment flux due to the incident waves also decreased with elevation above the bed. Transport rates at all other frequencies were very small.

The statistical significance of these peaks in sediment flux can be ascertained through an analyses of phase relationships (Davidson *et al.*, 1993). Phase shifts between velocity and concentration greater than 270° or less than 90° will result in onshore sediment flux, while phases between 90° and 270° will cause an offshore flux. Ninety-five per cent confidence bands for the phase estimates were computed using the coherence estimates (Jenkins and Watts, 1968). The phase shifts for the infragravity peaks identified in Fig. 12(a) and (b) thus become $330^\circ \pm 7^\circ$ ($z = 0.05$ m) and $329^\circ \pm 7^\circ$ ($z = 0.10$ m) for the 90 m station, and 179° ($z = 0.05$ m), 172° ($z = 0.10$ m) and $171^\circ \pm 8^\circ$ ($z = 0.15$ m) for the 111.5 m station. The cospectral peaks at infragravity frequencies were thus significant in terms of both direction and magnitude; the spectral peaks at incident wave frequencies were also statistically significant.

Net suspended sediment flux was computed using equation (1). However, owing to the fact that the instruments were inside the bottom boundary layer associated with the mean currents, the mean flux ($\bar{u}\bar{c}$) must be corrected for the differences in elevation between the current meters and the concentration sensors. Assuming a logarithmic current profile, the velocity at the level of the OBS (u_2) relative to the measured velocity (u_1) can be computed as:

$$u_2 = u_1 \frac{\ln(z_2/z_0)}{\ln(z_1/z_0)} \quad (3)$$

where z_2 , z_1 are the elevations of the OBS and the current meter, respectively, and z_0 is the bed roughness length. Under flat bed conditions, z_0 is often estimated as $d/30$ (e.g. Nielsen, 1992), where d is sediment grain size. However, in nature the bed is rarely if ever perfectly flat. Measurements of z_0 under natural conditions are scarce; however, Sherman and Greenwood (1984) reported values of 0.0003 m under "flat bed" conditions in natural surf zones (see also Greenwood and Sherman, 1988). This value of z_0 is adopted here.

According to equation (3), and accounting for the fact that the bed had accreted by 0.05 m at 111.5 m, u_2/u_1 at this station was 0.73 at the level of the middle OBS (nominally $z_2 = 0.10$ m, but in reality $z'_2 = 0.05$ m). The estimation of z'_2 and z'_1 at 90 m is somewhat more difficult. At this time the bed had accreted by approximately 10% of the total accretion measured at the 111.5 m station; assuming a similar erosion rate at 90 m, the bed would have eroded by approximately 0.09 m at 1400 h (total amount of erosion: 0.92 m). The elevation of the lowest OBS would thus become $z'_2 = 0.14$ m and $u_2/u_1 = 0.84$. Mean current velocities measured at 90 and 111.5 m have been reduced using these values.

Oscillatory sediment transport rates have not been reduced, since the optical sensors were almost certainly outside the bottom boundary layer associated with the incident waves, and probably in the uppermost part of the boundary layer associated with the

Fig. 13. Suspended sediment flux rates at 90 m and 111.5 m, 1402 h, 16 October. Shown are fluxes due to infragravity wave motions (inf.grav; black), incident wave motions (inc; horizontal stripes) and mean currents ($\bar{u}\bar{c}$; dots), as well as the net sediment fluxes.

infragravity waves. According to Grant and Madsen (1986), the thickness of the turbulent oscillatory boundary layer in the presence of currents can be approximated as:

$$\delta \approx \kappa u_* / \omega \quad (4)$$

where κ is von Karman's constant (taken as 0.4), ω the radian frequency of the waves, and u_* is the shear velocity given by:

$$u_z = (u_* / \kappa) \ln (z / z_0) \quad (5)$$

where u_z is velocity measured at elevation z . The velocity records were separated into a low-frequency band ($f < 0.067$ Hz) and a high-frequency band ($f > 0.067$ Hz), and the significant orbital velocities were computed as $u_s = 2(u_o^2 + v_o^2)^{1/2}$. At the 90 m station (at 1400 h), u_s was 0.40 m s^{-1} at low frequencies, and 1.07 m s^{-1} at high frequencies, yielding $\delta = 0.16 \text{ m}$ ($f = 0.01$ Hz), and $\delta = 0.03 \text{ m}$ ($f = 0.17$ Hz). Corresponding values of δ at 111.5 m were 0.08 m and 0.02 m, respectively. According to these computations, the lowest optical sensor at 90 m and the middle sensor at 111.5 m were in the upper part of the infragravity wave boundary layer (at 90 m at the very top), while all sensors were outside the incident wave boundary layer (except the lowest OBS at 111.5 m which was being buried). The lack of correction for oscillatory transports thus does not seem to be critical, particularly as equation (4) is only an approximation, and the computation of δ relies very much upon the definitions used.

Computed sediment flux rates and direction at 90 and 111.5 m are presented graphically in Fig. 13. The estimates are based on data from the lowest concentration sensor at 90 m ($z' = 0.14 \text{ m}$) and the middle sensor ($z' = 0.05 \text{ m}$) at 111.5 m. This figure illustrates the oscillatory suspended sediment flux induced by infragravity waves ($f < 0.067$ Hz) and incident waves ($f > 0.067$ Hz), as well as the mean transports ($\bar{u}\bar{c}$) and finally the local net suspended sediment flux rate. Interestingly, a large gradient in net sediment flux seems to have existed between the two positions, with the local net flux being essentially zero ($+0.004 \text{ kg m}^{-2} \text{ s}^{-1}$) at 90 m ($z' = 0.14 \text{ m}$), and directed offshore ($-0.490 \text{ kg m}^{-2} \text{ s}^{-1}$) at

111.5 m ($z' = 0.05$ m). This flux gradient was a result of the opposing directions of sediment flux induced by the infragravity waves at the two positions. While incident waves and mean currents transported sediment onshore and offshore, respectively, without any significant spatial gradients in these terms (although the mean flux decreased slightly on the lakeward slope), the sediment flux at infragravity frequencies was directed onshore at 90 m and offshore at 111.5 m. Thus, at the location of the pre-storm bar crest, there was a balance between the onshore-directed oscillatory flux and the offshore-directed mean flux, with the flux at infragravity frequencies amounting to approximately 75% of the mean flux. At 111.5 m, the sediment flux attributable to infragravity waves was directed offshore augmenting the effects of the mean currents, although at this station they were clearly subordinate to the mean flux ($\approx 40\%$). While the detailed spatial variation in the flux gradient between the 90 and 111.5 m stations is not well documented, the evidence provided earlier points to an infragravity wave surface elevation node between the 90 and 100 m stations (Table 2). We would expect maximum infragravity flux to occur at the node and therefore the flux gradients would increase and then decrease between the 90 and 111.5 m stations since the infragravity fluxes would appear to be the important constraint on the spatial flux gradients. If this was indeed the case, this would account for the erosion at 90 m and the accretion at 111.5 m (Fig. 13).

It is clearly problematic to compare the net sediment flux at two different elevations; here, flux rates were computed at $z' = 0.14$ m (90 m station) and $z' = 0.05$ m (111.5 m station). This problem is particularly pertinent if the oscillatory sediment flux reverses with elevation above the bed. However, such was not the case during this event. While we are unable to compare sediment flux rates at similar elevations or compute vertically integrated sediment transport rates, it is obvious from the cospectra [Fig. 12(a) and (b)] that oscillatory flux rates were qualitatively similar with increasing elevation above the bed; this is particularly evident at infragravity wave frequencies. Furthermore, computing the sediment flux at 111.5 m using the OBS at $z = 0.15$ m ($z' = 0.10$ m), the net flux at this elevation was $-0.405 \text{ kg m}^{-2} \text{ s}^{-1}$, which is only 17% less than the flux at $z' = 0.05$ m. Thus, while flux rates obviously decreased with elevation above the bed, the flux directions were consistent (at least in the lower part of the water column) for a given cross-shore location. The flux divergence identified between the 90 and 111.5 m stations appears, therefore, to be real and not an artifact of sensor location relative to the bed.

At 1800 h, the only concentration sensor still functioning was the uppermost OBS at the 116.5 m station. This instrument was located very close to the bed at this time ($z' \approx 0.03$ m), and indicated that the mean flux had increased by approximately 30%, while the infragravity flux rate had decreased to about 60% of the rates recorded at 1402 h (Fig. 14). Furthermore, the direction of sediment flux at incident wave frequencies had reversed; the result was a very large offshore net flux at 116.5 m ($z = 0.15$ m; $z' = 0.03$ m) of approximately $-0.730 \text{ kg m}^{-2} \text{ s}^{-1}$. Whether the zero net sediment flux condition at 90 m persisted at this time can not be ascertained.

DISCUSSION AND CONCLUSIONS

This study has demonstrated that nearshore profile changes may occur very rapidly. During a storm, a lacustrine nearshore bar of significant size (≈ 0.7 m high) migrated some 25 m offshore. Estimates from the accretion rates leading to the build up of the displaced bar profile indicated that the bar migration speed was on the order of 2.5 m h^{-1} with the

Fig. 14. Suspended sediment flux rates at the 116.5 m station, 1800 h, 16 October. Infragravity wave fluxes (inf.grav; black), incident wave fluxes (inc; stripes) and mean flux rates ($\bar{u}\bar{c}$; dots) are shown, as well as the net flux rate.

erosion rate of the former bar being approximately $1.5 \text{ m}^3 \text{ m}^{-1} \text{ h}^{-1}$. These magnitudes are comparable to the largest rates measured in an ocean environment (Birkemeier, 1985) and differ significantly from earlier reports on bar migration in other protected, storm-dominated, coastal environments, e.g. the Great Lakes (Greenwood, 1986; Davidson-Arnott and McDonald, 1989; Bauer and Greenwood, 1990) and Denmark (Aagaard, 1990; Aagaard *et al.*, 1994). These earlier studies concluded that nearshore bars in such environments are comparatively stable with horizontal adjustments on the order of 5–10 m during storms. Sediment transport rates may be quite large, while net changes remain small (Greenwood, 1986). Bars are therefore often in a state of dynamic equilibrium with storm conditions, at least in part because morphological changes between storms are small owing to low wave energy conditions in general and the lack of prolonged periods of swell in particular. In contrast, the present study has documented the topographic response in a situation where the morphology would appear to be significantly out of equilibrium with a particular set of environmental conditions. This response may be almost instantaneous with very little lag time, and the adjustment towards a new equilibrium may occur quickly provided that the energy level is large and sustained. Once a new equilibrium is approached (in this case approximately 9–10 h after the beginning of the storm) a quasi-stable morphology is restored.

It is of interest to speculate why the accretion rate on the lakeward slope of the bar (and the offshore bar migration) decreased sharply at 2200 h (16 October). While we have no data on sediment flux rates at this time (due to instrument collapse and/or burial), it is suggestive that mean offshore currents were decreasing, their maximum having been reached 2–3 h previously. However, wave heights over the second bar did not reach a maximum until at least 0100 h (17 October). The fact that the mean current velocity was leading local wave height by several hours could have been due to the offshore migration of the second bar which decreased the gradient of the lakeward slope (e.g. Fig. 13) and thus local set-up gradients. Assuming that the offshore directed mean current was an undertow,

smaller setup gradients should theoretically result in lower undertow velocities (e.g. Svendsen, 1984).

A contributing factor to the reduced accretion rates after 2200 h could have been lower sediment transport rates associated with the infragravity waves. These waves appeared significantly more broad-banded after ≈ 2200 h. While this study has demonstrated that large rates of sediment flux can occur with a reasonably narrow infragravity spectrum, Aagaard and Greenwood (1994) showed that in the case of a broad-banded spectrum, net infragravity flux rates may be quite small, due to interactions between different wave modes. It should be emphasized that while the sensors all registered a narrow spectral peak at ≈ 0.009 Hz during the beginning of the storm, this does not automatically imply that this peak was significant across the entire nearshore. However, the fact that the spectrum at infragravity frequencies appeared narrow-banded in the velocity as well as in the sediment concentration records [Figs 5 and 10(a) and (b)] across the sensor array ($x = 80\text{--}127$ m; i.e. at distances of 55–100 m from the shoreline) might suggest that this was indeed the case. Furthermore, a white shoreline spectrum would produce a peak frequency spread of 0.003 Hz across the array. We observed a frequency spread of 0.0011 Hz. This also suggests that the spectral structure seen by the instruments was not just an artifact of the amplitude variation of a cross-shore standing wave. Finally, the fact that the infragravity waves as recorded by the current meters definitely became more broad-banded after 2200 h suggests that a significant change in the wave field occurred as the storm and the morphology evolved.

While the mean currents would appear to be the single most important mechanism in the offshore sediment flux, the primary reason for the rapid erosion of the second bar was probably the large offshore-directed gradient in suspended sediment flux at the bar crest, which occurred during the early hours of the storm. The flux divergence was primarily the result of opposing oscillatory fluxes at infragravity frequencies (≈ 0.009 Hz). At the pre-storm bar crest, where the net flux was essentially zero, the onshore-directed infragravity and incident wave flux balanced the offshore-directed mean flux. On the lakeward slope, the sediment flux at infragravity frequencies was directed offshore, augmenting the effect of the mean currents, but clearly subordinate. Sediment flux divergence caused by an onshore-directed flux at incident wave frequencies and an offshore-directed mean flux appeared to be small. If the flux gradients observed during the early hours of the storm are representative for the event, we must conclude that infragravity waves may be important to morphological change in the surf zone by imposing spatial gradients on sediment transport associated with mean currents.

The flux divergence attributed to the infragravity waves (at ≈ 0.009 Hz) could well be induced by a cross-shore standing wave structure, with a node in surface elevation close to the pre-storm bar crest. Unfortunately, the existence of this node cannot be confirmed from surface elevation records, as only one pressure sensor was functioning satisfactorily. However, several lines of evidence support this hypothesis. First, cross-spectral analyses of the cross-shore velocity field revealed an in-phase relationship across the bar at 0.009 Hz (Table 2). This is indicative of standing wave motion at this frequency, without an antinode (but possibly a node) in surface elevation between 90 m (pre-storm crest) and 111.5 m (pre-storm lakeward slope). Second, cross-spectral analysis of the time series of sediment concentration revealed that a phase shift of nearly 180° occurred at low frequencies between 90 and 100 m. At this spatial scale, it is difficult to suggest an alternative to a shift in the direction of the hydrodynamic forcing. Such shifts might conceivably occur at

standing wave nodes (see below). Third, the theoretical cross-shore structure of the measured infragravity waves can be computed using the method proposed by Holman and Bowen (1979), according to which the effective nearshore slope felt by a 0.009 Hz, mode 1 edge wave would be 0.0158, and the first node would have occurred at $x = 96.5$ m. This suggests that a node in surface elevation may very well have been located between the 90 and 100 m stations. The cross-shore length scales of the infragravity waves may indeed be important constraints on the cross-shore length scales of the nearshore profile (and hence the position of bars) by perturbing a more or less spatially uniform offshore directed sediment transport associated with mean currents. This would apply particularly in those cases where the infragravity waves have a narrow frequency range.

This hypothesis requires clarification of the fact that suspension maxima occurred only once during every infragravity wave cycle (Fig. 11) and that a preferred direction of infragravity wave sediment flux existed. Such directionalities have been reported earlier by Beach and Sternberg (1988, 1991), Davidson *et al.* (1993) and Aagaard and Greenwood (1994). Noda (1968) showed that at positions between standing wave nodes and antinodes, oscillatory velocities in the bottom boundary layer were asymmetric with larger velocities directed towards antinodes. It is possible that such velocity (and thus shear stress) asymmetries in the standing infragravity wave boundary layer could account for the larger sediment concentrations observed, at times when the oscillatory velocity was directed towards antinodes, causing a net transport towards the antinodes. An alternative scenario would be an interaction between infragravity wave surface elevation and incident wave velocities. Orbital velocities at incident wave frequencies were on the order of $0.1\text{--}0.2\text{ m s}^{-1}$ larger under infragravity wave crests than under troughs. A 180° phase shift in sediment concentration (suspended by the incident waves) would occur if sensor locations were on either side of a standing wave node. In this scenario, infragravity wave velocity asymmetries would also be required to cause a net sediment transport towards surface elevation antinodes.

Sediment flux patterns across the second bar in this highly dissipative environment appeared less complicated than a similar situation in a moderately dissipative surf zone, reported on by Aagaard and Greenwood (1994). In the latter case, mean currents and sediment flux associated with such currents were directed landward up the seaward slope of the second bar and across the bar crest. Landward of the bar crest, an offshore directed mean flux occurred. Onshore directed oscillatory flux at incident wave frequencies were significant, particularly on the bar crest and seaward slope; sediment flux at infragravity frequencies was highly variable both spatially and temporally as several infragravity modes were present with temporally varying intensities and thus transport capabilities. The situation was further compounded by the occurrence of far infragravity (shear?) waves. The net result was a sediment flux convergence on the landward slope of the second bar, causing an onshore migration of the bar. In the present case, mean currents were consistently directed offshore across the bar and the infragravity frequency band was dominated by a single spectral peak across the array, resulting in what appeared to be a spatially more homogeneous transport regime.

A number of field studies (e.g. Jaffe *et al.*, 1985; Beach and Sternberg, 1988, 1991; Osborne and Greenwood, 1992a,b; Aagaard and Greenwood, 1994, in press) have emphasized the importance of oscillatory currents in suspended sediment transport. This conclusion is endorsed by the present study. While the mean flux was clearly the single most important transport mechanism, cross-shore sediment flux gradients (and thus

transport divergence) were related to oscillatory wave motions, particularly at infragravity frequencies. Foote *et al.* (1993) similarly argued that interactions between long waves and mean flows were significant elements in the generation of velocity asymmetries and thus net sediment transport within the surf zone. Numerical models for sediment transport and beach profile change which fail to include oscillatory terms must therefore be inadequate.

Oscillatory currents were also clearly important in the mobilization and resuspension of sediment, making this sediment available for transport by mean currents. This study has not been able to determine whether the shear stresses associated with the infragravity waves or incident wave shear stresses modulated by a standing infragravity wave field was the more important mechanism in the sediment suspension. Until we develop a sensor which can reliably measure current and/or sediment velocities within infragravity (and incident) wave boundary layers under field conditions, our understanding of the processes of sediment resuspension will remain inadequate and computations of net sediment transport rates will only be very rough estimates.

Acknowledgements—This is a contribution from the Canadian Coastal Sediment Transport Programme (C-COAST) supported by Strategic and Operating Grants awarded to B. Greenwood from the Natural Sciences and Engineering Research Council of Canada. T. Aagaard was supported by an International Postdoctoral Fellowship from the same body, and by the Danish Natural Science Research Council, grant no. 11-8164. We would like to thank all of those who assisted with the data collection, especially R. W. Brander, D. Rockwell, K. Jagger, R. Atkins and R. Richards (University of Toronto) and Dr A. J. Bowen, D. G. Hazen and S. McLean (Dalhousie University). John Jønsson and O. B. Berthelsen (University of Copenhagen) assisted with the graphics. This paper benefited significantly from reviews by Dr D. A. Huntley and the anonymous referees. Thanks are also due to the staff at Pinery Provincial Park for their assistance, especially Mr Kobayashi, Mr D. Matheson, Mr T. Crabe and Mr K. Armstrong; a special thanks to Marge. Finally, the Atmospheric Environment Service, Environment Canada, and Mr John Deary in particular, are thanked for assistance with recording meteorological variables.

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