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SHORT PAPER

Scale of tsunami-generated sedimentary structures in deep water

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To date, no attempt has been made to explore possible constraints on maximum water depths for tsunami-generated sedimentary structures of various sediment grain sizes, in terms of estimated wavelength and amplitude of potential bedforms. This paper explores some possible quantitative limits on threshold velocities for sediment movement at specified water depths, assuming Airy wave theory which treats tsunami waves as shallow-water, small-amplitude, long-wavelength waves. Such a treatment suggests that the passage of tsunami waves is unlikely to entrain sediment coarser than fine sand to coarse silt in water depths in excess of 200 m. Although fraught with theoretical difficulties and a lack of observations of modern sedimentary bedforms attributable to known tsunamis, this paper is presented to stimulate debate rather than provide definitive answers.

'Tsunami' is the Japanese word for 'harbour wave', derived because of the amplification of water level commonly associated with the passage of such waves from the open sea into restricted coastal embayments. Tsunamis are generated by catastrophic, geologically instantaneous, displacements of the seafloor. The largest tsunamis are related to seismogenic activity associated with large-scale faulting and displacement of the seafloor over large areas, rather than to explosive volcanism.

Meteorite impacts in or near the sea can generate tsunamis, as postulated for the 'tsunami deposit' at the Cretaceous–Tertiary boundary, Brazos, Texas (Bourgeois *et al.* 1988). Bourgeois *et al.* (1988) calculated that a 10 km diameter meteorite impacting in the ocean up to 5000 km away could have produced the conditions necessary for deposition of the sandstone bed (two amalgamated beds) in Texas. They noted that a more proximal impact in shallower water or of a smaller object could produce the same effect.

Of the total number of seismic events affecting the seafloor, relatively few actually produce detectable tsunamis. The most tsunami-prone region of the world today is the Pacific Ocean basin which is rimmed by active continental margins, island arcs and marginal basins.

During a tsunami-generating seismic event, the vertical displacement of the seafloor may be distributed over a wide area from which waves propagate outwards. After a few hundred kilometres of travel from the epicentre, the waveforms appear to propagate as radially-spreading waves. In the Pacific Ocean, where water depths are generally about 5.5 km, typical wave speeds are about 230 m s^{-1} (*c.* 15% speed of sound), with observed nearshore wave periods of 15–100 minutes (Apel 1987).

In Japan, there is a large database on tsunamis (inundation or run-up heights), and it has been found that the inundation values vary not only regionally but also locally for a single event within a 10–20 km stretch of coastline. This variation is a function of wave refraction, resonance and non-linear effects due to topographic irregularities on various scales (Kajiura 1983).

Tsunamis have long wavelengths (of the order of 200 km), and in the open ocean small amplitudes (decimetres); thus tsunamis are essentially undetectable by ships at sea. As tsunamis approach the shelf and coast, however, the abrupt shoaling of water depths causes a dramatic increase in wave amplitude as the uniformly distributed wave energy (from top to bottom) is compressed. Tsunami waves may be enhanced if they are in phase with the semidiurnal tidal cycle.

A critical problem concerns possible aspect ratios (amplitude/wavelength) of sedimentary structures produced by tsunamis, the solution to which depends upon obtaining reliable data relating wave height to water depth. Such data can then be used to infer wavelengths of tsunami deposits in specified water depths. Although data of this kind have been available for some time, they have not yet filtered into the geological literature to better constrain discussion of tsunami-generated sedimentary structures.

This paper considers, using Airy wave theory and some data from known historic tsunamis, the aspect ratios of sedimentary structures formed in certain grain-size populations, which may result from the passage of a tsunami over the seafloor. Secondary water movement, such as that generated by backflow from shallow-water and coastal set-up, together with offshore-directed surges and associated sediment gravity flows (e.g. turbidity currents), will be capable of generating a big range in bedform aspect ratios. It seems intuitively likely that back-flow surges could be of sufficient magnitude to move all normally available grain sizes and to produce a very heterogeneous assemblage of bedforms. Such secondary bedforms may form the bulk of tsunami-generated structures, but are beyond the scope of this paper.

Chilean tsunami, 24 May 1960. One of the best documented tsunamis of historic time was that which resulted from a major Chilean earthquake on 24 May 1960 (see papers in Takahasi 1961). That earthquake was of magnitude 8.5, originated on the continental shelf between Concepcion (37°S) and Castro (43°S) off Chile, and the wave front travelled across the Pacific Ocean as shown in Fig. 1. Off the Chilean coast, the earthquake caused local uplift of 1 m, and subsidence elsewhere of about 1.6 m (Takahasi & Hatori 1961).

A study of the arrival times for the tsunami along the coast of Japan (Takahasi & Hatori 1961) show that there was a time delay caused by the refraction of the wave front over the Izu–Bonin and Marianas island arcs. Inundation heights for the tsunami waves along the Japanese coast were up to 3.8 m, with reported inundation heights at Isla Macha and Mehuin, Chile, of more than 20 m and 15 m, respectively. Typical reported wave periods range between 40–80 minutes (table 4, Takahasi & Hatori 1961). The inundation height is not the same as the wave height prior to the tsunami wave surging and breaking onto the coast: the wave height will be considerably less, but its value for a given water depth offshore is rarely documented.

Tsunami waves are reflected not only at the coast but as a result of impinging upon submarine ridges. Wave

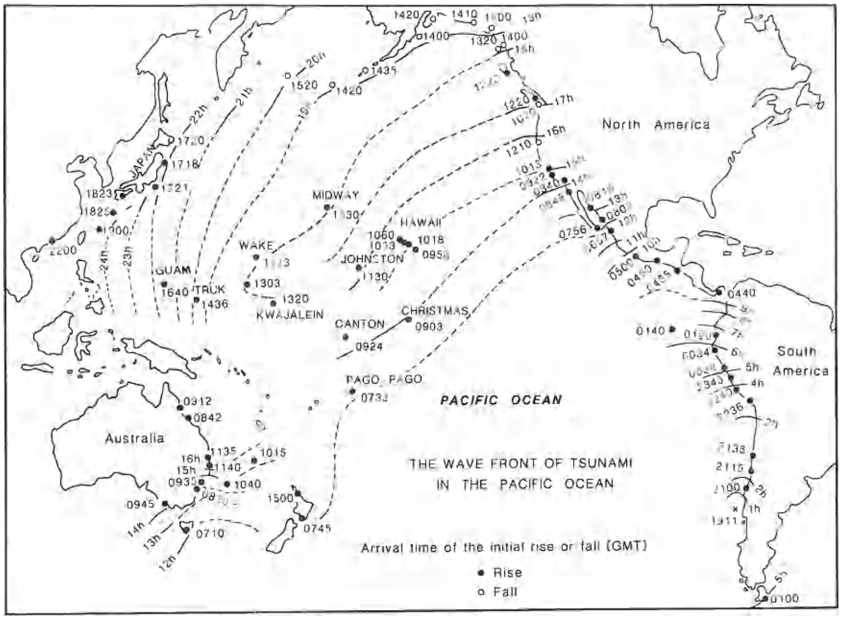

Fig. 1. Wave front of 1960 Chilean tsunami in the Pacific Ocean. Arrival times for the initial rise or fall of the mean water level shown as Greenwich Mean Time (GMT) from epicentre X. Redrawn from Takahasi (1961).

reflection appears most effective when the width of a ridge, at typical shelf depths, is 0.1–0.2 of the wave length (Takahasi & Hatori, in Takahasi 1961, p. 23). The Chilean tsunami of 1960, with a wavelength of about 500–800 km and wave period of up to 80 min, shoaled from 4500 m water depth to impinge on the Izu–Bonin island arc (ridge) with a width of 100–150 km. Since the ratio of ridge-width to wavelength was 0.1–0.15, the long period waves were reflected very effectively at the Izu–Bonin island arc.

Ida & Ohta (*In* Takahasi 1961, p. 108) calculated that the Chilean tsunami caused a wave of *c.* 40 cm amplitude from a wave of period 60 min in a water depth of 1000 m, about 35 km off the coast of Japan at Maisaka. Furthermore, they calculated that a wave of period 60 min would produce an approximately 65–70 cm amplitude wave in 200 m water depth (figs 6 & 7, p. 117, Ida & Ohta 1961).

Wave parameters, water depth. The long wavelength of tsunami waves compared to water depth, even in the deepest parts of the oceans (an order of magnitude difference), and their small amplitude, allows the application of shallow-water (sinusoidal) Airy wave theory. The equations used here are mainly those summarized in table 1-III by Allen (1984), Masuda & Makino (1987) and from Clifton & Dingler (1984).

In shallow water, the wavelength of a wave can be expressed as (Dyer 1986):

$$L = \frac{gT^2}{2\pi} \tanh(kh) \tag{1}$$

where L = wavelength in water depth h; T = wave period; tanh = hyperbolic function $(e^x - e^{-x})/(e^x + e^{-x})$; k = radian wave number $= 2\pi/L$.
For deep-water waves, where $h \ll L$ ($h/L > 0.25$), tanh (kh) tends towards unity, therefore eq. (1) simplifies to:

$$L_0 = \frac{gT^2}{2\pi} \tag{2}$$

where L_0 = wavelength in very deep water.

L can also be expressed as a function of L_0 and h (this formula is an approximation, Komar 1976 pp. 42–44), such that:

$$L = L_0[\tanh(k_0 h)]^{0.5} \tag{3}$$

where $k_0 = 2\pi/L_0$.
Wave height, H, in a known water depth, h, can be expressed as a function of the wave height in very deep water, H_0:

$$H = \frac{H_0}{(4k_0h)^{0.25}} \tag{4}$$

Equation (4) is based on the assumption that the energy flux of a wave (tsunami wave in this case) remains constant as the bottom shoals. In reality, there may be a significantly large amount of energy dispersion due to wave interaction with the bottom sediments, but this will serve only to reduce the water depths at which tsunami waves can mould bedforms on the seafloor. Dyer (1986 p. 99) expresses the energy flux per unit length of crestline of a wave, E, as $E = \rho g a^2/2$, where ρ = fluid density, g = gravitational acceleration, and a = wave amplitude. $a = H/2$ and the energy flux is thus proportional to cH^2, where c = wave celerity. For shallow water waves, $c = (gh)^{0.5}$, which gives:

$$H_1^2(gh_1)^{0.5} = H_2^2(gh_2)^{0.5} \quad \text{or} \quad H_1^4 h_1 = H_2^4 h_2 \tag{5}$$

The height of a tsunami wave in the open ocean, say in a water depth of H_1, is obtained such that:

$$H_1 = H_2(h_2/h_1)^{0.25} \tag{6}$$

where h_1 = ocean water depth; H_2 = wave height above the shelf; h_2 = shelf depth.
For example, if $H_2 = 6$ m, $h_1 = 5000$ m, and $h_2 = 10$ m, then H_1 is approx. 1.27 m.

While Airy wave theory is specifically applicable to relatively small amplitude waves in deep water, it does provide a reasonable approximation to measured orbital diameters, and near-bottom maximum velocities, for waves of finite amplitude in shallow water (Clifton & Dingler 1984). In Airy wave theory, the orbital diameter at the seafloor, d_0, is related to the wave amplitude, H, water depth, h, and wavelength, L, in a specified water depth as follows:

$$d_0 = \frac{H}{\sinh(2\pi h/L)} \tag{7}$$

where sinh is the hyperbolic function $0.5(e^x - e^{-x})$.
Calculating d_0 and assuming orbital wave theory (where d_0 is related to characteristic length of a bedform, λ) allows an estimate of the wavelength of a bedform (Miller & Komar 1980), using the empirical equation:

$$\lambda = 0.65 d_0 \tag{8}$$

This equation may be inapplicable to very long-wavelength, low-amplitude, bedforms for which no observational data are available. Clifton & Dingler (1984) suggest that orbital ripples are related to the mean grain size D, and form under conditions where the ratio of d_0/D is in the range 100–3000 or more. Orbital ripples have a wavelength that can be expressed as some function of the orbital diameter of the oscillatory fluid motion that generates the bedform. Finally, an estimate of the maximum horizontal orbital velocity near the seafloor for shallow-water, u_m, can be obtained (Wiegel 1964, table I-III in Allen 1984):

$$u_m = \frac{Hc}{2h} \tag{9}$$

where c = shallow-water wave speed = $(gh)^{0.5}$; g = gravitational acceleration; h = water depth; H = wave height.

Using the above equations and available data for measured wave heights in estimated water depths, it is possible to calculate the wave parameters H_0 and L_0, using eqs. (1) and (2) for the very deep open ocean, then substitute these values into eqs. (3) to (7) to estimate the wave parameters for any specified water depth.

Assuming a wave period, $T = 40$ min (2400 s), a wave height, $H = 6$ m, water depth, $h = 10$ m, wavelength in very deep water, L_0 from eq. (2) (based on data from the 1960 Chilean tsunami along the SE coast of Japan), then using eq. (4), the wave height in very deep water, $H_0 = 0.44$ m. Thus, a wave height of 6 m in coastal waters would be only of the order of decimetres out in the open ocean. Obtaining a value for H_0 (0.44 m) enables a calculation of the wave height in any given water depth, say 500 m, $H_{500} = 2.28$ m, from eq. (4).

The near-bottom orbital diameter, d_0, is calculated from eq. (5), i.e. for the conditions above, in 500 m water depth, with $L = 1.7 \times 10^5$ m obtained from eqs. (2) and (3), such that $d_0 = 122$ m. The equilibrium wavelength associated with this orbital diameter, from eq. (6), would be about 80 m. These values would give a maximum orbital velocity at the substrate, from eq. (7), $u_m = 0.16$ m s^{-1} for a water depth of 500 m.

Using reasonable values of tsunami wave height in

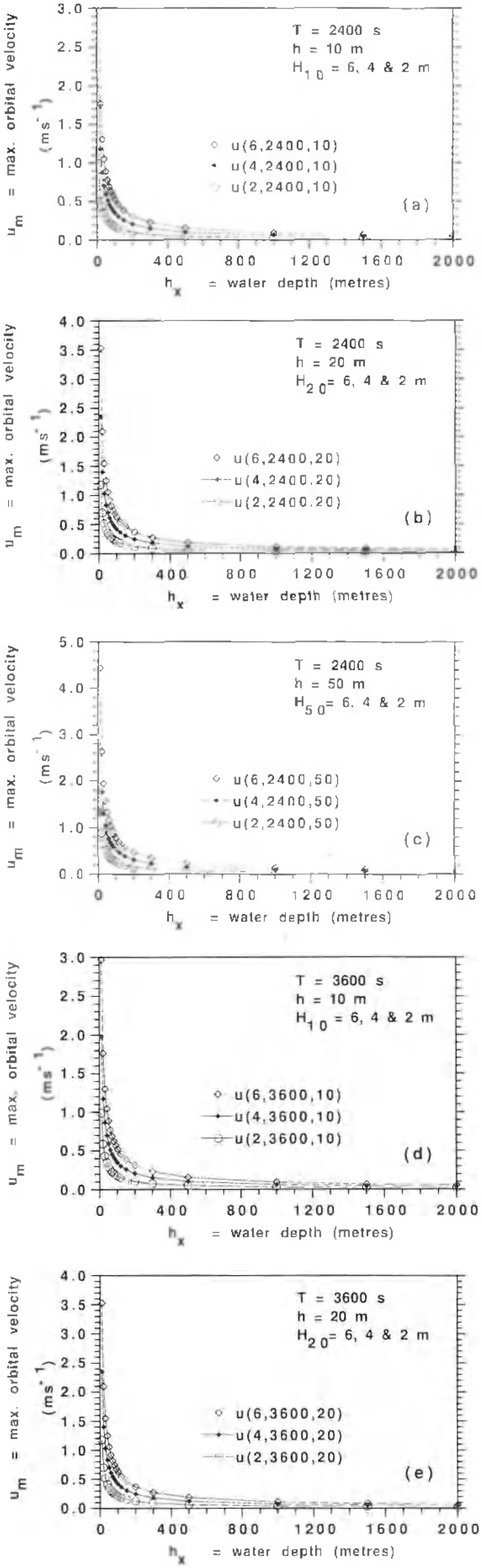

Fig. 2. Graphs of maximum orbital velocity, u_m in m s^{-1}, at the sea floor against water depth, h_x in metres. These graphs show that at depths greater than about 200 m maximum, velocities and related shear stresses at the seafloor would probably not be sufficient to entrain/transport sediment coarser than silt-grade. Cohesive muds would also be difficult to erode.

coastal waters, it is possible to construct a series of graphs for u_m against water depth, h (Fig. 2). The principal conclusions that can be drawn from such graphs are: (a) for reasonable values of H and T, it is unlikely that velocities at the seafloor would be sufficient to move sand greater than coarse silt to very fine sand in water depths greater than 200 m (i.e., below the shelf break), and (b) the wavelength of bedforms associated with the passage of tsunamis over the seafloor will typically be of the order of 50–100 m.

Discussion. Although, in theory, bedforms produced as a direct result of the passage of tsunami waves over the seafloor may possess predictable bedform aspect ratios that are related to the wave parameters, the coastal set-up of water and the subsequent back-surge offshore may be of sufficient magnitude to generate other sandy bedforms in any water depth. Undoubtedly, the back-surge of water can generate various sediment gravity (unidirectional) flows which interact with the essentially long-period (oscillatory) wave activity of the tsunami. Such structures probably possess attributes that are difficult to quantify and predict, but may be similar to those in 'tempestites' produced by storm-surge ebb currents.

The topography of the seafloor immediately prior to the passage of a tsunami wave will be an important factor in controlling the nature of any sedimentary structures which might be generated. A knowledge of bottom friction under wave action is required in order to fully understand sediment movement and wave modification. The case for laminar flow is reasonably well understood, whereas the structure of the turbulent frictional boundary layer under oscillatory flow is poorly investigated (Kajiura 1968), a case that is perhaps more appropriate to tsunamis. Bagnold (1946), using an empirical formula, presented a quadratic friction law for oscillatory flow in the presence of artificial ripples in which the friction coefficient decreases with increasing particle excursion distance, i.e. where there is an increase in wave period and/or the amplitude of the velocity oscillation. This theoretical and experimental work relates to shallow water wave conditions and there are insufficient data to assess the real applicability to oscillatory flow associated with the passage of tsunami waves.

Palaeocurrent analysis will provide the only reliable method of assessing the orientation and position of the shoreface/coastline relative to the shoaling direction of tsunami waves. The shear stress at the loose boundary, or sediment surface, associated with the passage of a tsunami will possess a strong velocity-time asymmetry that renders any predictions about the internal architecture of a macrobedform rather uncertain. The probable scale or aspect ratio for tsunami-generated bedforms, assuming Airy wave theory, should be kept in mind during field observations of suspected ancient analogues. Finally, we accept that Airy wave theory may not provide the best framework to describe tsunami-generated macroforms, and that for such structures $\lambda = 0.65d_0$ may also be inapplicable. More theoretical work must be tied to field observations before our predictive model can be improved.

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