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Modeling the development of marine terraces on tectonically mobile rock coasts

A.S. Trenhaile *

Department of Earth Sciences, University of Windsor, Windsor, ON, Canada N9B 3P4

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Abstract

A mathematical model was used to study the effect of glacially induced fluctuations in sea level on the formation of wave-cut terraces on tectonically mobile rock coasts. Two deep water wave sets were used to calculate breaker height and depth, which, along with surf zone width, bottom roughness, and the gradient of the submarine slope, dictated the force exerted at the shoreline. An erosional force threshold was employed to represent the variable strength of the rocks. The model also considered tidal range, the time that the water level spent at each intertidal elevation, and the protective effect of debris accumulation at the cliff foot. Three hundred runs were made with constant sea level and with different rates of rising and falling relative sea level. Rates of erosion increased with the rate of rising and falling sea level, but eventually decreased in some runs with very rapid changes in relative sea level. Fifty-five longer runs were also made with a Quaternary sea level model that consisted of 26 glacial cycles representing the period from 2 million to 0.9 million years ago, and nine cycles, of approximately twice the amplitude and wavelength, in the last 0.9 million years. These runs were made on a landmass experiencing slow (0.11 mm yr^{-1}) or fast (0.74 mm yr^{-1}) positive or negative changes in the elevation of the land. It was found that on rising landmasses, erosional coastal terraces are formed during interglacial stages, and on subsiding landmasses during glacial stages. The number of terraces increased with the rate of uplift and subsidence, and with the slope of the hinterland. Terrace gradient increased with tidal range, and decreased with rock resistance. The was an inverse relationship between terrace width and the strength of the rock.

Abbreviations: C, coefficient to represent differences in the force exerted at the waterline by plunging, spilling or surging-collapsing breakers; E_s, amount of submarine erosion each year (m); E_y, amount of intertidal erosion each year (m); F_b, wave force at the breakers (kg m^{-2}); g, acceleration due to gravity (m s^{-2}); h, water depth (m); h_b, breaker depth (m); H_b, breaker height (m); H_o, deep water wave height set consisting of five wave heights H_o1 to H_o5 (m); k, surf attenuation factor related to bottom roughness; L, deep water wavelength (m); M, coefficient (6.5×10^{-10} to 3.25×10^{-8} m^3 kg^{-1}) to convert the excess surf force into the amount of intertidal erosion during each model iteration; MHWN, mean high water neap tidal level; MHWS, mean high water spring tidal level; MLWN, mean low water neap tidal level; MLWS, mean low water spring tidal level; MT, mean tidal level; Q, a cliff-foot protection factor related to the amount and persistence of the debris at the cliff foot. When $Q = 0.2$, debris accumulation reduces the erosion rate at the cliff foot to one-fifth of the erosion rate of an unprotected cliff foot ($Q = 1$); s, submarine erosion depth decay constant (m^{-1}); S_f, surf force at the waterline (kg m^{-2}); S_{fmin}, threshold erosional strength of the rocks (kg m^{-2}); T, wave period (s); T_d, tidal duration value, the time each year that the water level occupies each intertidal elevation (h yr^{-1}); T_r, spring tidal range (m); W, hourly number of waves of each of the five deep water heights; Ws, surf zone width (m); β, gradient of the bottom from the breakers to the waterline (°); δ, initial surface gradient (°); ρ_w, unit weight of sea water (1025 kg m^{-3})

* Fax: +1-519-973-7081. *E-mail address:* tren@uwindsor.ca (A.S. Trenhaile).

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1. Introduction

Gently sloping marine terraces, bounded on the landward side by a steeper ascending slope, and on the seaward side by a steeper descending slope, are a common feature of many rock coasts. There are stacks and sea caves on some terraces, and shallow water marine and terrestrial deposits, often at considerable elevations above the present level of the sea. Marine terraces are up to several kilometres in width in western Europe and other tectonically stable regions, and they can extend up to 200 m or more in elevation, (Orme, 1964; Guilcher, 1974). These surfaces are much higher than the maximum elevation reached by the sea during Quaternary interglacial stages, and they were therefore probably formed during long periods of high sea level in the Tertiary (Haq et al., 1987). There are flights of elevated terraces, ranging up to 2 million years or more in age, along the steep hinterlands of the western coast of North America, and on other tectonically mobile coasts (Shepard and Wanless, 1971; Pillans, 1983; Lajoie et al., 1991). These terraces were generally formed during the Quaternary, and they were elevated by tectonic uplift to their present positions. Whereas Tertiary terraces in tectonically stable regions developed along coasts that are imbedded in, or are situated along the trailing edge of, tectonic plates, Quaternary terraces have often developed on tectonically mobile coasts along plate boundaries (Inman and Nordstrom, 1971; Davies, 1972).

A number of workers have modeled the long-term development of marine erosion surfaces under stable sea level conditions on slowly changing rock coasts (Flemming, 1965; Horikawa and Sunamura, 1967; Scheidegger, 1962, 1970; Sunamura, 1976, 1977, 1978a; Trenhaile and Layzell, 1981; Trenhaile, 1983, 1989; Trenhaile and Byrne, 1986). Only a few workers have considered the effects of changing sea level. They include Sunamura (1978b), who modeled the development of erosional continental shelves in a tideless sea during the Holocene transgression; Trenhaile and Byrne (1986), who investigated the role of Holocene changes in relative sea level on the evolution of intertidal shore platforms; and King (1963) and Scheidegger (1962, 1970), who modeled the effect of a steady rise and fall in sea level on steep rock coasts. More recently, Trenhaile (2001a) used a mathematical model to investigate the effect of glacially induced fluctuations in sea level on the evolution of intertidal shore platforms and erosional continental shelves during the last 2 million years.

There have been only a few attempts to model the development of coastal terraces in tectonically mobile regions. Cinque et al. (1995) and Anderson et al. (1999) studied the formation of marine terraces with glacial sea level changes on tectonically mobile coasts, using simple cliff erosion models which did not consider the vertical distribution of wave energy expenditure within the intertidal zone. Trenhaile (1989) modeled the effect of intertidal wave erosion on the evolution of continental shelves and elevated coastal terraces with changing sea level over five glacial–interglacial cycles during the middle and late Quaternary. The present paper uses a different intertidal erosion model and is the first to consider the development of terraces on tectonically mobile coasts over the entire Quaternary Period.

2. The model

The mathematical wave erosion model discussed here has been used previously to examine shore platform evolution under stable sea level conditions (Trenhaile, 2000), the role of weathering in platform evolution (Trenhaile, 2001b), the effect of different sea levels in the last two interglacial stages on contemporary intertidal shore platforms (Trenhaile, 2001c), and the evolution of shore platforms and erosional continental shelves on stable coasts during the Quaternary (Trenhaile, 2001a). The model uses basic wave equations to explore the interaction between wave dynamics, tides, coastal morphology, and rock erosion at the shoreline. The model considers neither marine deposition nor post-formational modification of terraces and cliffs by physical and chemical weathering, stream channel incision, and mass movement and other slope processes (Anderson et al., 1999). To avoid unnecessary repetition, only a fairly brief overview of the derivation and assumptions of the wave erosional model are presented here; the reader is referred to Trenhaile (2000) for a more detailed discussion.

The model was based on the assumption that mechanical wave action, by water hammer, high shock pressures generated by breaking waves, air compression in rock crevices, and shallow water abrasion, is usually the most important erosional mechanism in the stormy, wave-dominated environments of the middle latitudes (Everard et al., 1964; Trenhaile, 1987; Sunamura, 1978c). Wave mechanical processes are closely associated with a narrow zone above and below the fluctuating waterline, where there are alternations of air and water and where waves exert the greatest pressures. Most wave erosion therefore occurs at the water surface, at the surf–rock interface (Trenhaile, 1987). Although the water surface fluctuates within the intertidal zone, the mean water surface is most frequently at, or close to, the mean high and low water neap tidal levels; the water surface spends about one-third less time at the mid-tidal level, and much less time between the neap and spring high and low tidal levels (Carr and Graff, 1982; Trenhaile, 1987). The location of the mean water surface, and therefore the effect of mechanical wave erosional processes, is increasingly concentrated between the mean high and low water neap tidal levels as the tidal range (T_r) decreases (Carr and Graff, 1982).

The wave force at the breakers (F_b) was represented by (Coastal Engineering Research Center, 1984):

$$F_b = 0.5\rho_w h_b \tag{1}$$

where ρ_w (1025 kg m^{-3}) is the unit weight of sea water and h_b is the breaking wave depth. The critical ratio between breaking wave height (H_b) and breaking depth has a mean value of about 0.78. Therefore:

$$H_b = 0.78\ h_b \tag{2}$$

From Eqs. 1 and 2:

$$F_b = 0.5\rho_w(H_b/0.78) \tag{3}$$

The proportion of the breaker energy that reaches the surf–rock interface, and is therefore available for erosion, depends on the width and surface roughness of the surf zone. The surf force (kg m^{-2}) reaching the waterline (S_f) was represented by a decay function of the form:

$$S_f = 0.5\rho_w(H_b/0.78)e^{-k.\mathrm{Ws}} \tag{4}$$

where k is a surf attenuation constant, and Ws is the width of the surf zone ($h_b/\tan\beta$, where β is the gradient of the bottom extending from the breakers to the waterline). Incorporating the effects of a number of additional factors provided an excess surf force equation for horizontal intertidal erosion (m) at the waterline at the head of the surf zone (E_y):

$$E_y = M\sum_{W=1}^{5}(T_d W(0.5\rho_w C(H_b/0.78)e^{-k.\mathrm{Ws}} - S_{\mathrm{fmin}})) \tag{5}$$

where M is a scaling coefficient to convert the excess surf force into rock erosional units (values ranged from 6.5×10^{-10} to 3.25×10^{-8} m^3 kg^{-1}); T_d is the number of hours per year in which the water level is at a given intertidal elevation (the tidal duration value); C is a coefficient representing differences in the force exerted at the waterline according to the type of breaking wave; S_{fmin} is a threshold minimum surf force term used to represent the strength of the rock; and W is the hourly number of waves of each of the five deep water height categories in each of the wave sets (calculated from the wave period). Two wave sets were used in this study, representing west coast swell (wave set 1) and exposed storm wave environments (wave set 2) (Table 1). E_y is the total ero-

Table 1
Wave data used in the model runs

Wave set:	1					2				
Height (m)	1	2	3	4	5	1	3	5	7	9
Frequency (% per yr)	40	35	17	6	2.3	54	26	13	4.3	2.6
Period (s)	6	7.5	8.8	10	10.8	6	8.8	10.8	12	13

Wave set 1 was collected off the Galician coast (west coast swell environment) in northwestern Spain (Dirección General de Puertos del Estado). Wave set 2 was collected off Cornwall (storm wave environment) in southwestern Britain (Darbyshire, 1956). Wave period for a fully arisen sea were calculated for each area after Neumann (1953).

sion for the five wave height categories in the wave sets at each of five intertidal levels. Tidal duration values were obtained for macrotidal Swansea, Wales, and mesotidal Burnie, Tasmania (Carr and Graff, 1982). Breaker height was determined from the deep water wave data using Komar and Gaughan's (1972) expression:

$$H_{\mathrm{b}} = 0.39g^{0.2}(TH_{\mathrm{o}}^{2})^{0.4} \quad (6)$$

where g is the acceleration due to gravity, T is wave period and H_{o} is wave height in deep water.

Horizontal submarine wave erosion (E_{s}) was made proportional to the amount of erosion at the waterline, according to the expression:

$$E_{\mathrm{s}} = E_{\mathrm{y}}\mathrm{e}^{sh} \quad (7)$$

where s is a depth decay constant and h is the water depth.

Eqs. 5 and 7 were used to calculate the amount of erosion over each iteration interval at mid-tide (MT), at the mean high water spring (MHWS) and mean high water neap (MHWN) tidal levels, and at the mean low water spring (MLWS) and mean low water neap (MLWN) tidal levels. These levels are at vertical intervals of 2.25 m and 0.75 m in the macrotidal (9 m) and mesotidal (3 m) environments used in this study, respectively. Subtidal erosion, from the MHWN tidal level to a maximum depth of half the wavelength of the wave below the mean low water spring tidal level, was also calculated at 2.25 m intervals in macrotidal runs and at 0.75 m intervals in mesotidal runs. Model runs were made using eight variables, each of which was assigned one of two possible values (Table 2). The initial surface was linear, with a gradient (δ) of either 15° or 30°.

2.1. The constants

The constant k represented the rate at which surf energy is expended between the breaker zone and the waterline. Eq. 4 was used to calculate the surf force (S_{f}) generated at the waterline for a wide variety of surf zone widths, k values, and breaker heights. The results suggested that a k value of 0.1 could suitably represent high rates of attenuation across irregular surf zones, and 0.01 low rates of attenuation across smoother surf zones (Trenhaile, 2000). The only field data on wave attenuation across shore platforms has been collected by Stephenson and Kirk (2000), who found that as little as 5–7% of the wave energy arriving at the seaward edge of an approximately 70-m-wide platform in New Zealand reaches the cliff foot. These data are limited, however, and they were collected in a dominantly microtidal environment, on a gently sloping platform fronted by a distinct low tide cliff, which

Table 2
Variables and constants in the model runs

T_{r} (m)	k	s	Q	S_{fmin} (kg m^{-2})	δ (°)	Waves	Uplift/subsidence (mm yr^{-1})
3	0.01	1	0.2	20	15	lower	slow (0.11)
9	0.1	2	1	1000	30	higher	fast (0.74)

The columns are independent of each other and no horizontal correlation is implied between them.

causes shearing of the upper part of the incoming waves (Sanders, 1968). Therefore, k values of 0.01 and 0.1, which allow 50 and 0.1%, respectively, of the breaker wave force to reach the waterline across a 70-m-wide surf zone, were used to safely encompass the full range of behaviour on sloping, macro- and mesotidal platforms, lacking low tide cliffs, in vigorous wave environments. The same technique suggested that S_{fmin} values of 1000 kg m^{-2} and 20 kg m^{-2} were suitable to represent the minimum surf forces capable of eroding resistant and weak rocks, respectively (Trenhaile, 2001a). Eq. 7 was used to determine realistic values for s, the depth decay constant. Erosion rates at depths of 1 m and 5 m were 0.37 and 0.007, respectively, of the erosion rate at the waterline when $s = 1$, and 0.135 and 4.54×10^{-5}, respectively, of the erosion rate at the waterline when $s = 2$ (Trenhaile, 2000). The coefficient M converted the excess force exerted at the surf–rock interface into the amount of horizontal cliff and platform erosion. Eq. 5 was used to determine values for this coefficient that provided horizontal rock erosion rates for each iteration that fell within the range of annual rates that have been measured in the field (0–20 mm yr^{-1}) (for example, see tabulations in Kirk, 1977 and Sunamura, 1992).

One must be cautious in converting model iterations into real time units because of uncertainty over the precise values of the constants M, k, s, and S_{fmin}. The values used for these constants were reasonable (Trenhaile, 2000), however, and they generated erosion rates that were consistent with annual rates measured in the field. Furthermore, the wide range of values used to represent each variable in this study compensated for the effect of over- or underestimation of the annual rate of erosion. It should also be emphasised that the model was used to study the factors that influence terrace development, rather than to reproduce the precise details of a terrace series within a given time scale. Therefore, although Eqs. 5 and 7 were used to calculate the amount of erosion accomplished in one year (one model iteration), it was the relative, rather than the absolute, time that wave erosion occurred at different elevations that was most important.

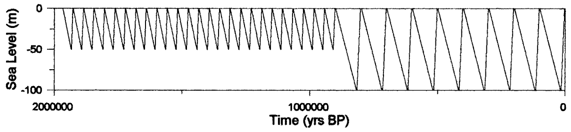

Fig. 1. The sea level model used in this paper. The scale of this diagram is too small to depict the sea level stillstands at the glacial and interglacial levels.

2.2. Relative sea level

The term ‘early Quaternary’ was used in this paper to refer to the period of frequent but fairly low amplitude glacial oscillations from 2 million to 900 000 years ago; the ‘middle Quaternary’ for the period from the end of the early Quaternary to the end of the last interglacial stage; and the ‘late Quaternary’ for the last glacial stage from 100 000 years ago to present.

Oxygen isotopic data from deep sea cores, speleothems and glaciers suggest that glacial cycles are ‘sawtoothed’, consisting of a long period of ice build-up and falling sea level during the glaciation phase, and a much shorter period of rising sea level during deglaciation. Glacial stages occurred at intervals of about 41 000 years between about 2 million and 900 000 years ago, and subsequently at intervals of approximately 100 000 years. The data also imply that since 900 000 years ago, the size of the ice sheets, and consequently the amplitude of the sea level cycles, has been about twice that in the early Quaternary (Ruddiman et al., 1986, 1989; Shackleton, 1987; Shackleton et al., 1990). Although actual sea level fluctuations were undoubtedly much less uniform, for computational purposes Trenhaile (2001a) used a simplified Quaternary sea level model to study the Quaternary development of tectonically stable rock coasts; that model was also used in the present paper to represent eustatic changes in sea level in the last 2 million years (Fig. 1). The main features of this sea level model were:

(a) There were 26 glacial cycles, each lasting 41 000 years, from 1 966 000 to 900 000 years ago, and nine larger ones, each lasting 100 000 years, in the last 900 000 years.

(b) The time allocated to each section of a glacial cycle was proportional to its wavelength.

Based on the last glacial stage, 3% of each cycle was allotted to glacial low sea level stands, and 3% to high interglacial stands; 82% of the total time was allocated to the slowly falling sea level at the beginning of glacial cycles, and 12% to the much faster rise in sea level following glacial minima.

(c) Sea level was at its present level in each interglacial stage, and at −49.5 m during glacial stages from 1 966 000 to 900 000 years ago, and at −99 m in the last 900 000 years.

The sea level model represented a simplification of average conditions during the Quaternary, rather than a faithful reproduction of the precise details of sea level changes according to one of the many disparate deep-sea core records. Consequently, whereas oxygen isotope substage 5e actually occurred between about 130 000 and 116 000 years ago, the peak of the last interglacial stage was from 103 000 to 100 000 years ago in the model.

The age and elevation of marine terraces on the western coast of North America suggest that rates of tectonic uplift usually range from about 0.11 to 0.74 mm yr^{-1} in areas characterised by strike-slip rather than by compressional tectonics (Ku and Kern, 1974; Muhs and Szabo, 1982; Muhs, 1983; Muhs et al., 1994). Therefore, to study the effect of eustatic, glacially induced fluctuations in sea level on tectonically mobile coasts, 55 model runs were made on coasts that experienced slow (0.11 mm yr^{-1}) or fast (0.74 mm yr^{-1}) positive or negative changes in the elevation of the land (Table 3). The effect of the variables on the morphology of shore platforms on tectonically stable coasts under constant and variable sea level conditions has been considered in detail elsewhere (Trenhaile, 2000, 2001a,c). Therefore, a more limited range of values was used in this part of the study in order to more fully explore the effect of glacial sea level oscillations on tectonically mobile coasts (Table 3).

Three hundred short model runs, lasting for 10 000 iterations (10 000 years), were also made to determine how erosion rates are influenced by the rate and direction of changes in relative sea

Table 3
Values used in the model runs, with glacial sea level oscillations, illustrated in this paper

Run	T_r	k	Wave set	M	S_{fmin}	s	Q	δ	Land movement
1	9	0.01	2	1	20	1	1	30	fast uplift
3	3	0.01	1	10	20	1	1	30	slow uplift
5	3	0.01	2	1	20	1	1	30	fast uplift
8	9	0.01	2	1	20	1	1	15	fast uplift
9	9	0.01	2	1	1000	1	1	15	fast uplift
12	9	0.01	2	1	20	1	1	30	slow uplift
15	9	0.01	2	50	20	1	1	30	slow uplift
17	3	0.01	2	1	20	1	0.5	30	fast uplift
18	3	0.01	2	1	1000	1	1	30	slow uplift
20	9	0.01	2	50	20	1	1	30	fast uplift
23	9	0.01	2	1	1000	1	1	30	slow uplift
25	3	0.01	2	10	20	1	1	30	slow uplift
29	9	0.01	1	1	1000	1	1	15	fast uplift
30	9	0.01	2	1	1000	1	1	15	slow uplift
31	9	0.01	2	1	1000	1	1	30	fast uplift
32	9	0.01	2	10	1000	1	1	15	fast uplift
33	9	0.01	1	1	20	1	1	30	fast uplift
34	3	0.01	2	1	20	1	1	30	slow uplift
46	9	0.01	2	1	20	1	1	15	slow uplift
47	3	0.01	2	10	20	1	1	30	slow subsidence
48	9	0.01	2	10	1000	1	1	15	fast subsidence
49	9	0.01	2	10	1000	1	1	15	slow subsidence
50	9	0.01	2	1	20	1	1	30	fast subsidence
51	9	0.01	2	1	20	1	0.5	30	slow subsidence

level (Table 4). These runs were made with relative sea level rising and falling at rates ranging up to 16.7 mm yr^{-1}, and, for comparative purposes, with constant sea level. The slower changes in sea level were similar to the mean rates of falling sea level during ice build-up (approximately 1–1.5 mm yr^{-1}), whereas the faster rates were similar to the mean rates of rising sea level during ice break-up (about 8–8.5 mm yr^{-1}). The model also represented intermediate and more extreme rates of sea level change that have been attained on tectonically mobile coasts where the land is also rising or falling, and for shorter periods associated with interstadial or more minor fluctuations in temperature.

3. Results

In addition to model runs for multiple glacial sea level cycles on rising and falling land masses, runs were also made to assess the contribution of uniformly rising or falling relative sea level on the development of tectonically mobile coasts.

3.1. Uniform changes in relative sea level

It has generally been accepted that rising sea level causes rapid erosion, and is therefore conducive to the formation of wide shore platforms and marine terraces (Ramsey, 1846; Gilbert, 1885; Davis, 1896; Bradley, 1958). This is because erosion of stable coasts over long periods has produced wide, gently sloping submarine shelves and intertidal shore platforms which attenuate the incoming waves. When there is an intertidal shore platform, water depth on the upper portion of the platform increases during high tide when sea level is rising (Fig. 2A), but the depth remains the same when sea level is falling (Fig. 2B,C). Under such circumstances, rising sea level would be conducive to rapid erosion because deeper water would permit waves to break closer to shore, and dissipate less energy in the subtidal and intertidal zones.

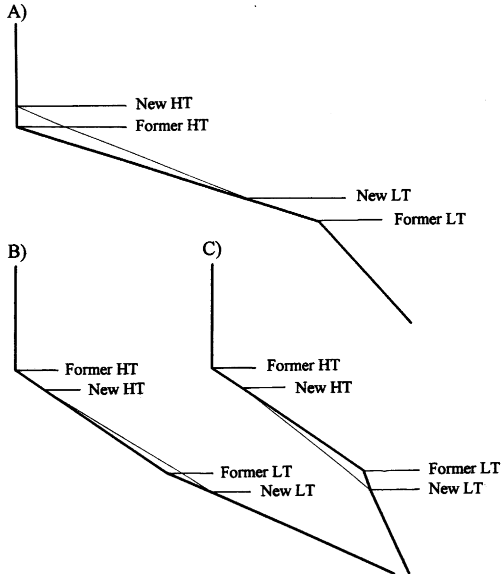

Fig. 2. Changes in intertidal gradient for: (A) a rise in sea level; (B) a fall in sea level when the intertidal gradient is greater than the submarine slope; and (C) a fall in sea level when the intertidal gradient is lower than the submarine slope. The thick line represents the previous supra-, inter- and subtidal surface. The thin line is drawn between the high (HT) and low (LT) tidal levels after a change in relative sea level.

There is no fundamental reason why erosion rates should vary according to whether relative sea level is constant, rising, or falling. The differences that do occur are therefore the result of erosion and slope reduction in the intertidal and subtidal zones. On tectonically mobile coasts, however, the sea may not be at any elevation long enough to produce very wide, gently sloping surfaces from steep initial slopes. The relationship between erosion and relative sea level changes in

Table 4
Values used in the model runs illustrated in this paper with constant and uniformly rising and falling sea level

Run	T_r	k	Wave set	M	S_{fmin}	s	Q	δ
10	9	0.01	1	1	20	1	1	30
20	9	0.01	2	1	1000	2	1	15
30	9	0.1	2	1	20	1	1	30
40	9	0.01	2	1	20	2	1	15

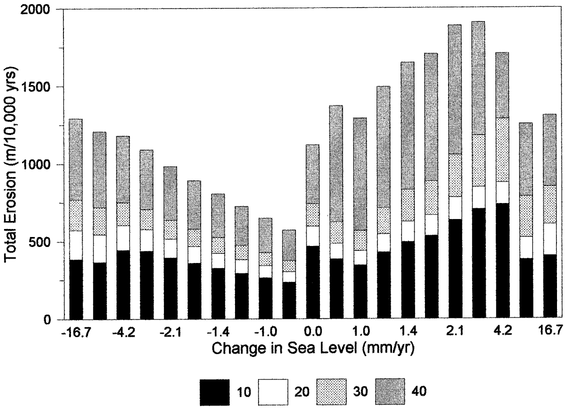

Fig. 3. Total amount of erosion accomplished by constant, rising and falling sea level for model runs 10, 20, 30, and 40 (Table 4).

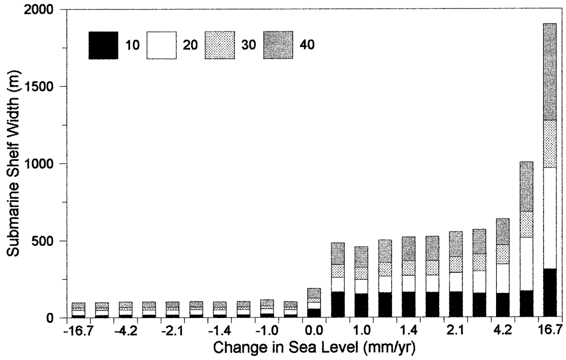

Fig. 5. The width of the submarine shelves produced in model runs 10, 20, 30, and 40 under constant and rising and falling sea level conditions (Table 4).

the model runs was quite complex; more erosion generally did occur with rising than falling sea level, but the precise relationships varied with the rate of change (Fig. 3). In general, the rate of erosion increased with the rate of rising and falling relative sea level, although there was an indication in some runs that erosion may eventually decrease with the most rapid rates of rising, and possibly falling, sea level, presumably because rapid changes in sea level do not allow much erosion to be accomplished at each level.

Trenhaile (1989) found that the intertidal width of simulated platforms was greatest when sea level was falling, but the much larger shelf, or subtidal surface, was much wider when sea level was rising: this was confirmed in the present study (Figs. 4 and 5). Platform width generally increased slightly with the rate of sea level rise, but it declined with the rate of fall, whereas shelf width generally increased with the rate of sea level rise. According to King (1963), erosion during a period of steadily falling sea level produces a subaerial slope parallel to the original surface, with an intertidal shore platform at its foot; this was modeled by Scheidegger (1962), Trenhaile (1989), and Cinque et al. (1995). The present study supported Scheidegger's conclusion that because the same amount of erosion occurs at each elevation, the erosion surface produced by both rising and falling sea level is parallel to the original surface. The model also confirmed Scheidegger's (1962, 1970) rejection of Popov's (1957) contention that a series of terraces can be formed by a uniform fall in sea level.

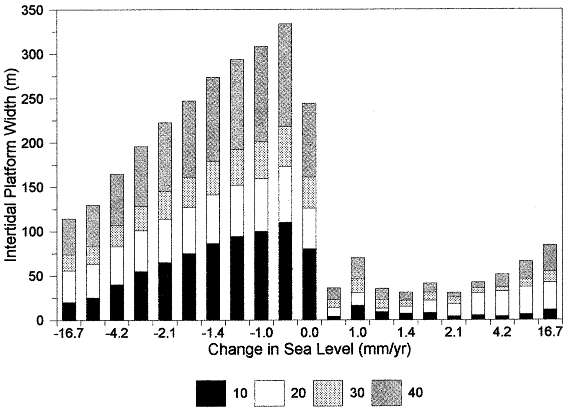

Fig. 4. The width of the intertidal shore platforms produced in model runs 10, 20, 30, and 40 under constant and rising and falling sea level conditions (Table 4).

3.2. Glacial cycles

Subaerial and submarine terraces can only develop and persist when the erosion accomplished at some elevation is significantly greater, in space or time, than the erosion above and below that elevation. On tectonically mobile coasts, it is theoretically possible that terraces could be the product of erosion during interglacial or glacial stillstands, or during periods of rising or falling sea level. They may also be modified or ultimately destroyed by erosion and terrace formation at

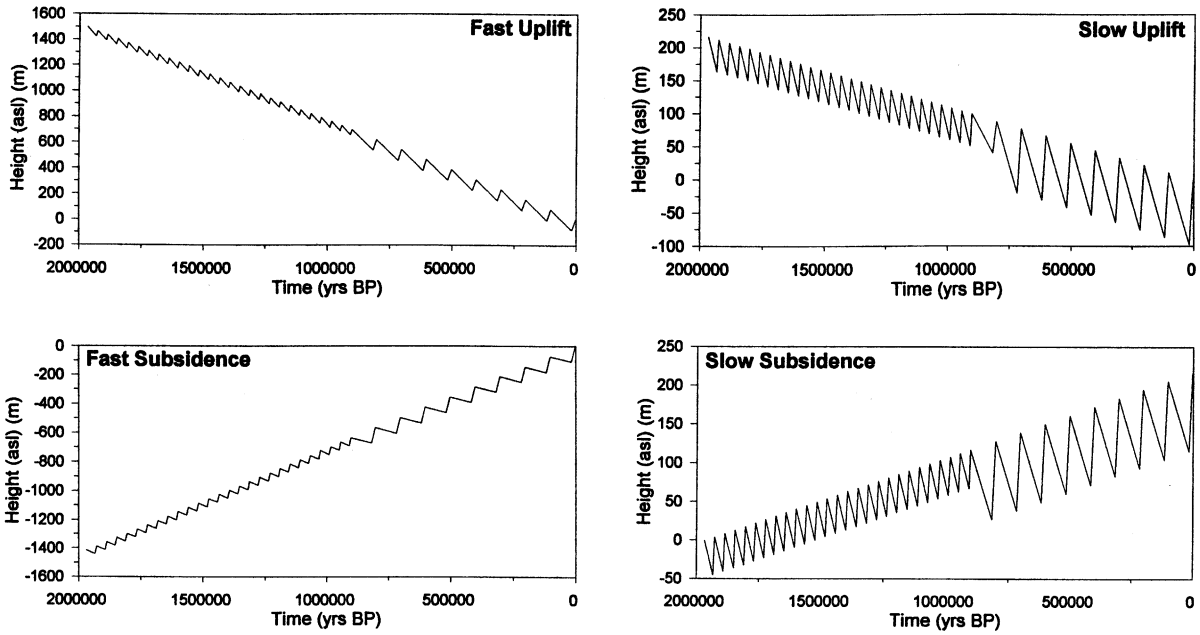

Fig. 6. The present elevations of palaeo-sea levels on emerging and subsiding coasts. The maximum and minimum of each 'saw-toothed' oscillation represents the interglacial and glacial stillstands, respectively.

lower elevations (Muhs et al., 1994; Anderson et al., 1999), particularly on steeply sloping coasts which experience slow uplift or submergence, and rapid wave erosion.

To account for the formation and distribution of simulated terraces by multiple, glacially induced oscillations in sea level, we must first consider the present elevation of the palaeo-sea levels generated in this model (Fig. 6). With rising land, the highest palaeo-sea levels were attained during the earliest Quaternary interglacials. The height of interglacial and glacial sea levels alternated at lower elevations, although they were not in chronological order. Only glacial stages are represented below the present level of the sea. With rapid uplift of the land, each early Quaternary interglacial and glacial sea level was about 30 m lower than its predecessor, and about 74 m lower in the middle and late Quaternary. The corresponding figures for slowly rising land were only 4.5 m in the early Quaternary, and 11 m in the middle and late Quaternary. Vertical differences between successive glacial and interglacial sea levels were the same for subsiding as for rising land. When the land was rising, however, glacial and interglacial sea levels occupied progressively lower elevations through time, but when the land was subsiding, glacial and interglacial sea levels occupied progressively shallower depths through time.

The relationship between wave attenuation and the gradient of the bottom implies that intertidal shore platforms become wider and more gently sloping until they attain a constant or equilibrium state (Edwards, 1941; Bird, 1968; Trenhaile, 1972); several mathematical models have supported this contention (Sunamura, 1978a; Trenhaile and Layzell, 1981; Trenhaile, 1983, 2000, 2001a). In the present study, the morphology of the terraces would be time-independent if they had attained equilibrium before being elevated above the level of the sea, whereas if they were not in equilibrium, their morphology would reflect, in addition to rock hardness, wave size and other morphogenic factors, the amount of time that marine processes had operated at that elevation. Very slow changes in terrace width, gradient and profile shape near the end of sea level stillstands, and terrace gradients that were consistent with the relationship between the slope of shore platforms and tidal range (Trenhaile, 1978, 1987, 2000), sug-

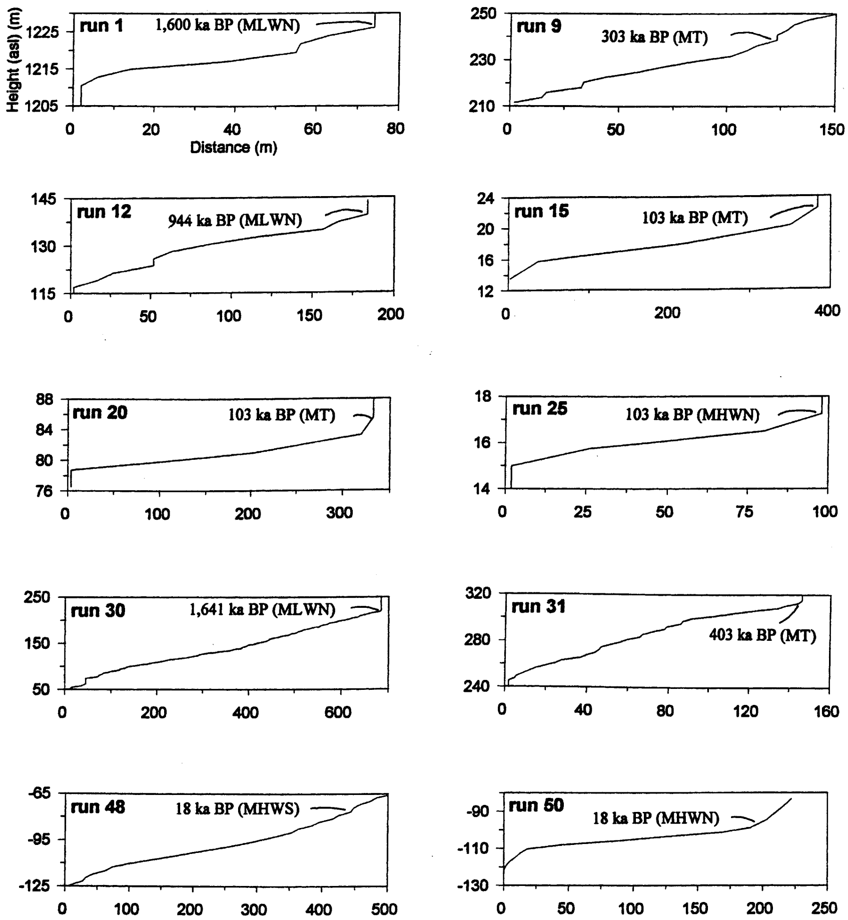

Fig. 7. Sample terrace profiles for rising and subsiding landmasses. The age (thousands of years before present) and tidal level of the inner edges at the time of formation are shown on each profile. The palaeo-tidal elevation of the poorly defined inner edges of terraces formed on subsiding land in runs 48 and 50 are approximations. The terraces on rising land were formed during interglacial stages, and on sinking land (runs 48 and 50), during glacial stages. The M and S_{fmin} values used in runs 9 and 30 tend to generate slow rates of erosion ($M = 1$, $S_{fmin} = 1000$) and fast erosion in runs 15 and 25 ($M = 50$ and 10, respectively, $S_{fmin} = 20$). The values used in other runs tend to produce intermediate rates of erosion. In all runs, however, the effect of other variables also needs to be considered.

gested that quasi-equilibrium was attained by some model terraces, particularly those formed during the longer glacial cycles of the middle and late Quaternary, and in runs with high M values, weak rocks, and other conditions that were conducive to rapid erosion. More rapid changes in morphology and steep gradients, relative to the tidal range, suggested that other terraces had not attained equilibrium at the end of model runs.

It is inconceivable that all the terraces in the field, formed under different morphogenic, geologic, and relative sea level conditions, were able to attain equilibrium in the limited time in which they were exposed to wave action. Therefore, the occurrence of equilibrium and non-equilibrium terraces in the model was probably consistent with field conditions, and with the occurrence of terraces with a wide range of widths and gradients. Furthermore, although some terraces did not attain equilibrium in the model, changes in terrace morphology decline through time, as the

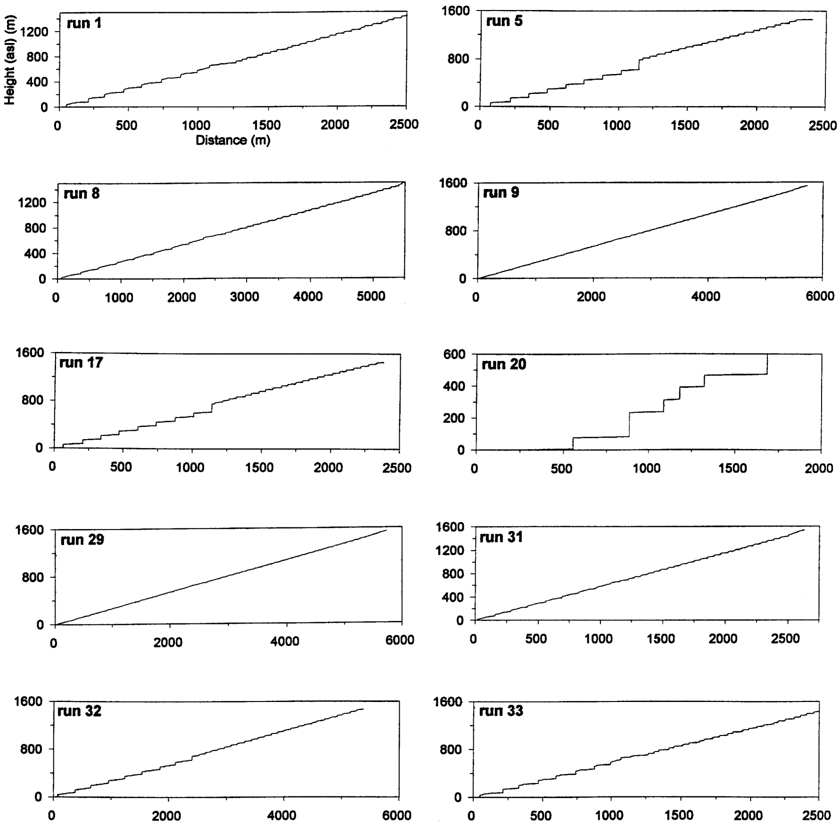

Fig. 8. Subaerial profiles produced in model runs with fast uplift (Table 3). The M and S_{fmin} values used in runs 9, 29 and 31 tend to generate slow rates of erosion ($M = 1$, $S_{fmin} = 1000$), and fast erosion in run 20 ($M = 50$, $S_{fmin} = 20$). Other runs used values that tend to generate intermediate rates of erosion. In all runs, however, the effect of other variables also needs to be considered.

terraces become progressively wider and more gently sloping, and most terraces had developed a morphology that was at least roughly similar to their equilibrium form.

3.2.1. Rising land

Palaeo-sea levels extended up to about 1460 m above present sea level on rapidly rising land, and up to about 225 m on slowly rising land (Fig. 6). Marine terraces were produced in all model runs, ranging in width from a few tens of metres up to more than 500 m, and with seaward gradients ranging from less than 1° up to 10° or more. Unlike intertidal shore platforms, which have a potential vertical extent equal to the tidal range, terraces on rising landmasses can be truncated, or in some cases extended, after their initial formation, by erosion at lower elevations. Two or more terraces may also coalesce, when wave erosion degrades or completely removes the low cliffs between them. Consequently, there was no relationship between terrace width and gradient. There was also no characteristic terrace profile. Many terraces were essentially linear, although they often contained several breaks in slope and slight ramps at the foot of the cliff (Fig. 7, runs 1, 9, 20 and 30). Some terraces were convex (Fig. 7, runs 12 and 31), however, and others were concavo-linear-convex, with a concave inner edge, or terrace–cliff junction, and a seaward slope consisting of either a smooth convex slope or a series of narrow, descending steps (Fig. 7, runs 15 and 25).

The general nature of the profiles that developed on rising land depended on the rate of up-

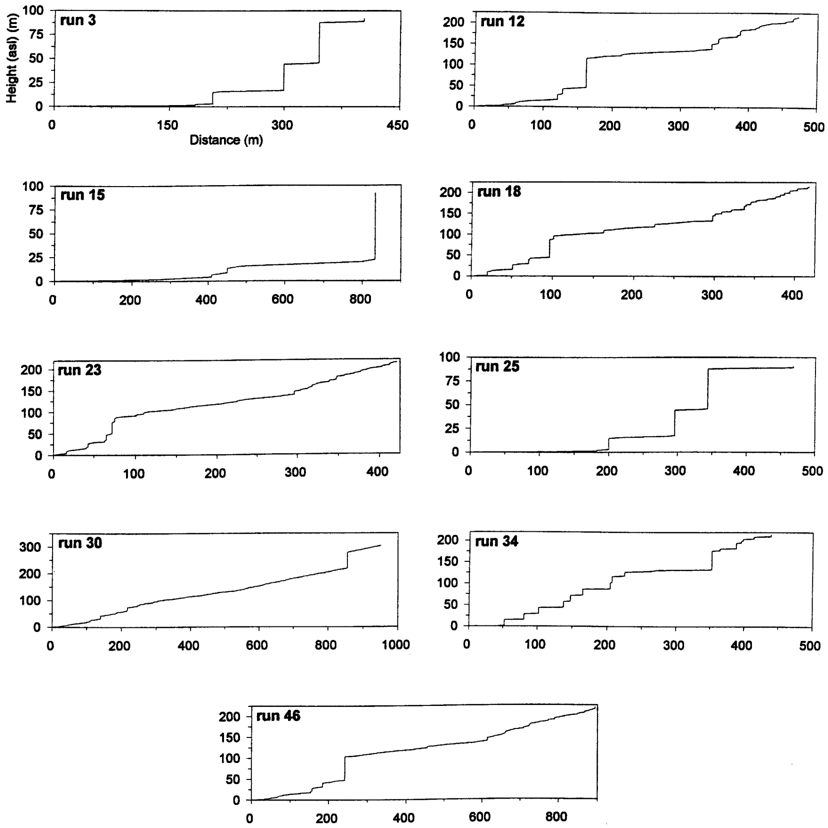

Fig. 9. Subaerial profiles produced in model runs with slow uplift (Table 3). The M and S_{fmin} values used in runs 18, 23 and 30 tend to generate slow rates of erosion ($M=1$, $S_{fmin}=1000$), and fast erosion in runs 3, 15 and 25 ($M=10$, 50 and 10, respectively, $S_{fmin}=20$). Values in other runs tend to produce intermediate rates of erosion. In all runs, however, the effect of other variables also needs to be considered.

lift. Large numbers of terraces were usually produced on rapidly rising land. The early Quaternary terraces in the upper portion of the profile were narrower than the fewer but wider middle and late Quaternary terraces at lower elevations (Fig. 8). These profiles were quite different from those produced under conditions of slow uplift, which were often dominated by a very wide surface in the middle part of the subaerial profile consisting of several individual terraces separated by very low cliffs (Fig. 9). This wide terrace developed at the transition between the short interglacial sea level stillstands of the early Quaternary and the longer stillstands of the middle Quaternary, however, and it would have been less prominent if the transition in the model had been more gradual.

The widest coastal terraces in California occur where there has been the least tectonic uplift (Bradley and Griggs, 1976), but, with the notable exception in some runs of the wide mid-profile surface, the terraces in the present study were generally narrower when uplift was slow than when it was fast. Although there was no relationship between terrace gradient and elevation in runs with either fast or slow uplift, a comparison of six runs, which differed only according to the rate of uplift, demonstrated that in each case, terrace gradient was 0.5–1.5° higher in the runs with fast uplift.

Other factors contributed to the more detailed form of the two main types of profile. Although there was little evidence of any relationship between tidal range and the number of terraces

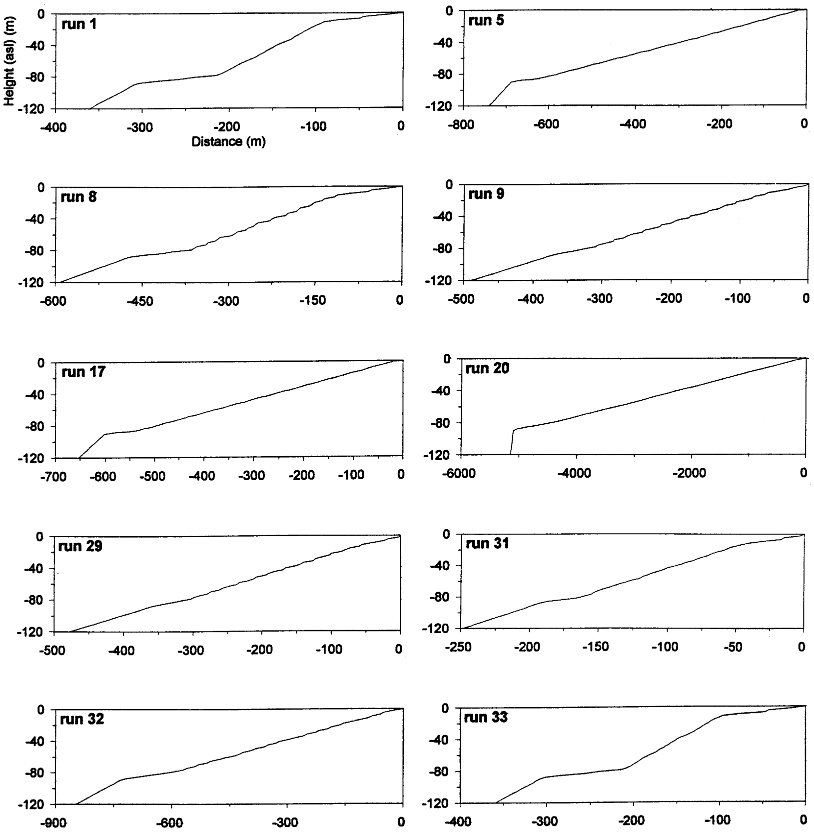

Fig. 10. Shelf profiles created in model runs with fast uplift (Table 3). The M and S_{fmin} values used in runs 9, 29 and 31 tend to generate slow rates of erosion ($M = 1$, $S_{fmin} = 1000$), and fast erosion in run 20 ($M = 50$, $S_{fmin} = 20$). The values used in the other runs tend to generate intermediate rates of erosion. In all runs, however, the effect of other variables also needs to be considered.

that developed, the gradient of the terraces was lower in runs with meso- rather than macrotidal range, a conclusion which was consistent with the relationship between tidal range and the slope of intertidal shore platforms found in previous model runs and in the field (Trenhaile, 1978, 1987, 2000, 2001a). In general, the terraces were wider and more gently sloping and the cliffs were higher when the resistance of the rock was low. Weak rocks and rapid erosion assisted the seaward truncation of erosional surfaces through the formation and extension of terraces at lower elevations, however, and consequently, in a few runs, because of reduced truncation, the terraces were wider in resistant than weak rocks in the lower portion of the profile. In a previous application of the wave erosional model to tectonically stable landmasses, it was found that because of low wave attenuation and the rapid erosion and modification of steep slopes, the morphology of the intertidal shore platforms became independent of the initial surface gradient after only one or two glacial cycles (Trenhaile, 2001a). The initial gradient retained its importance on tectonically mobile coasts, however, because the waves were repeatedly attacking previously uneroded sections of the initial profile during each stillstand. Although the effect of the initial slope was somewhat inconsistent in the model runs, more terraces were generally produced when the slope of the hinterland was steep than when it was gentle, and they tended to be narrower and more steeply sloping.

Terraces produced during interglacial stages on rapidly rising coasts are raised above the elevation of subsequent interglacial sea levels, thereby preserving their original form. The elevation of the

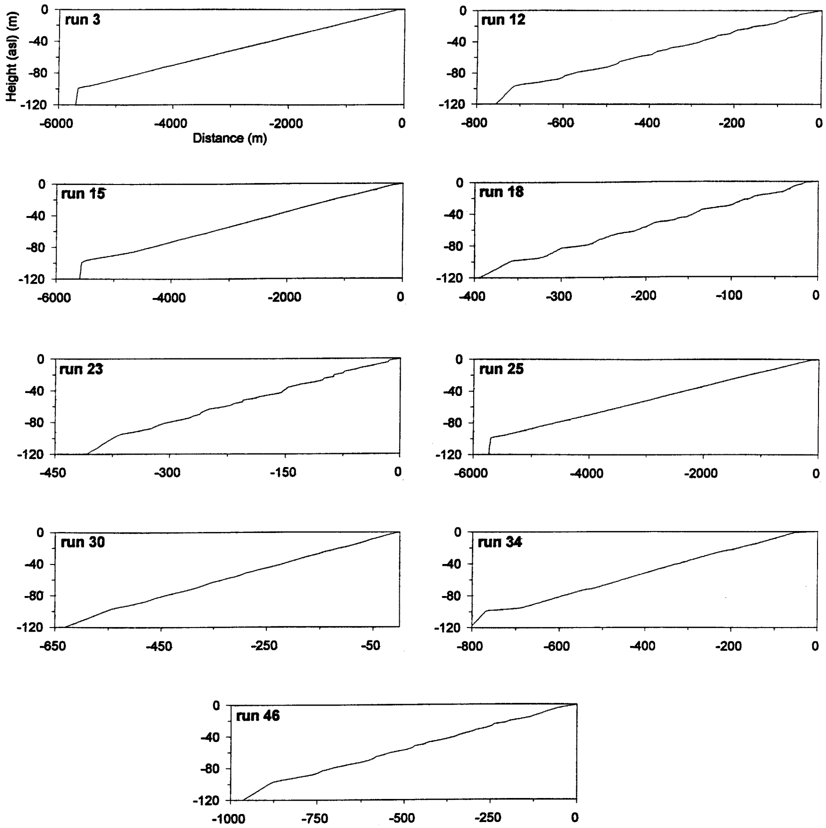

Fig. 11. Shelf profiles created in model runs with slow uplift (Table 3). The M and S_{fmin} values used in runs 18, 23 and 30 tend to generate slow rates of erosion ($M=1$, $S_{\mathrm{fmin}}=1000$), and fast erosion in runs 3, 15 and 25 ($M=10$, 50 and 10, respectively, $S_{\mathrm{fmin}}=20$). Other runs used values that tend to generate intermediate rates of erosion. In all runs, however, the effect of other variables also needs to be considered.

junction of each terrace with the former cliff, which will be referred to as the inner edge, was therefore usually within the tidal range during interglacial stages. Terraces on rapidly rising coasts provided an excellent record of Quaternary interglacial sea levels, extending up to about 1460 m above present sea level, and representing a period from the early Pleistocene to the Holocene. The interglacial record was complete in most of these runs, except when conditions were particularly conducive to rapid erosion. In these runs, cliff retreat, particularly during the longer interglacial stages of the middle and upper Quaternary, destroyed all but a few of the lower and more recent terraces (Fig. 8, run 20).

In runs with slow uplift, there was frequently overlap between the intertidal zones during interglacial and much older glacial stages, and sometimes between the intertidal zones of consecutive interglacial, and/or glacial, stillstands. This made it difficult to determine the relationship between sea levels and the occurrence of terraces at the end of model runs, particularly in runs with a high tidal range, and in the upper part of the profiles where there was low vertical separation between consecutive palaeo-sea levels in the early Quaternary. Nevertheless, the inner edge of the younger terraces in the lower portion of the profiles generally indicated that they were cut during interglacial stages. Because of overlap between palaeotidal ranges and the elimination of older terraces by younger terraces developing at slightly lower elevations, the terraces on slowly rising coasts generally provided only a very incomplete record

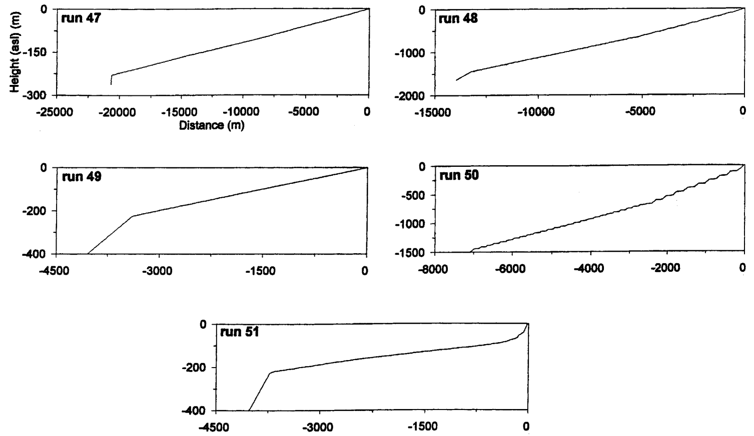

Fig. 12. Shelf profiles produced in model runs with subsiding land (Table 3). The M and S_{fmin} values used in run 47 tend to generate fairly fast erosion ($M = 10$, $S_{fmin} = 20$), and intermediate rates of erosion in the other runs. In all runs, however, the effect of other variables also needs to be considered.

of former sea levels, particularly of those in the early Quaternary. In most runs, no more than one-third of the Quaternary interglacial sea levels were represented by terraces, and in runs that were particularly conducive to rapid erosion, there were only a few late Quaternary terraces in the lower portion of the profile (Fig. 9, runs 3, 15, and 25).

The characteristics of the submarine surfaces produced in model runs with rising landmasses varied according to the rate of uplift. The positions of the last two glacial sea levels were below modern sea level on rapidly rising land, at depths of about 80–90 m (late Wisconsin) and 6–7 m. Although there was generally some flattening of the profile at both glacial levels, however, fast uplift otherwise produced essentially linear submarine profiles (Fig. 10). Well defined late Wisconsin terraces developed in runs where there was fairly rapid erosion and modification of the submarine surface (Fig. 10, runs 1 and 33), whereas only slight flattening was evident in runs with slow erosion (Fig. 10, runs 9, 29 and 31).

The last seven glacial palaeo-sea levels were within the subtidal zone in runs with slow uplift. The submarine profiles in all these runs were again essentially linear, however, except for some flattening at depths of −80 to −90 m (late Wisconsin) (Fig. 11, runs 15 and 34), and in a few cases, slight undulations in the profile corresponding to the positions of older glacial stillstands (Fig. 11, runs 12, 18 and 23). The general lack of well defined terraces and steep cliffs on the submarine shelves of rising landmasses can be attributed to post-formational modification by intertidal and subtidal wave erosion, associated with rising and falling sea level during subsequent glacial cycles. This explanation is consistent with the fact that the late Wisconsin terrace was the most persistent and best developed, because, as the deepest and most recent terrace, it was only subjected to a fairly brief period of post-formational erosion as the sea rose to its present position.

Although all the submarine profiles experienced erosion during model runs, unless conditions were particularly favourable for rapid wave erosion, the gradient of the eroded profiles in runs with rapid uplift was generally similar to that of the initial surface. In contrast, the submarine gradients in all the runs with slow rates of uplift were much lower than the initial profile gradient.

3.2.2. *Subsiding land*

The eroded profile extended from modern sea level to a depth of almost 225 m in runs with slow land subsidence, and to a depth of almost 1460 m in runs with rapid subsidence (Fig. 12). There were some similarities, as well as some differences, between the erosional profiles that developed on

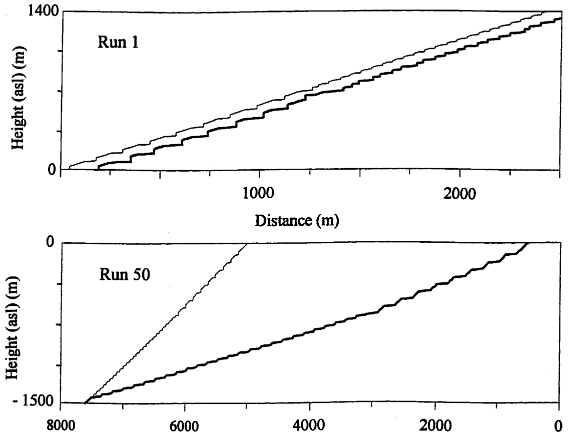

Fig. 13. Comparison of normal (thicker lines) and alternate (thinner lines) runs for rapidly rising (run 1) and rapidly sinking (run 50) landmasses. The alternate run 1 only allowed erosion during interglacial standstills, whereas the alternate run 50 only allowed erosion during glacial stillstands.

subsiding and emerging landmasses. The initial slope in all runs with subsiding land experienced erosion and significant slope reduction. On rapidly subsiding landmasses, terraces that formed during low glacial stillstands were preserved as they were quickly carried below the level of effective wave action. Therefore, whereas terraces on rapidly rising landmasses were produced during interglacial stages, terraces on rapidly subsiding landmasses were primarily glacial in origin. Furthermore, in contrast to interglacial terraces on rising coasts, glacial terraces on subsiding coasts experienced post-formational modification by intertidal and subtidal erosion as sea level began to rise to the next interglacial level; consequently, these terraces were more subdued in form, and were backed by sloping, degraded cliffs, rather than by the vertical cliffs of interglacial terraces (Fig. 7, run 48 and 50).

Terraces formed in hard rocks tended to have a shallow basin shape (Fig. 7, run 48). Their floors sloped more steeply seawards than terraces formed in weaker rocks, and the degraded cliff was more gently sloping. The lack of a distinct inner edge made it difficult to determine the palaeo-sea levels responsible for the formation of these terraces. Rapid erosion produced fairly well developed terraces backed by steep slopes (Figs. 7 and 13, run 50). The inner edges of these terraces were within the intertidal zones of former glacial stages, and the higher terraces, which formed during the latter part of the Quaternary, were wider than the lower terraces formed during the shorter and smaller magnitude glacial cycles of the early Quaternary. Despite post-formational modification, these runs provided an essentially complete record of glacial sea levels. Because the degree of modification increased with the depth and age of the terraces, however, the last few interglacial sea levels were also preserved in some runs as narrow terraces, positioned between the wider glacial terraces in the upper portions of the submarine slope. At greater depths, post-formational modification reduced the interglacial terraces to slight knicks in the steep slopes between the glacial terraces, and below that, all evidence of older interglacial sea levels was completely eliminated.

Post-formational modification by intertidal and submarine erosion, as sea level repeatedly swept up and down the profile, effectively removed the terraces on slowly subsiding landmasses and produced a wide, gently sloping shelf with an essentially linear profile (Fig. 12, runs 47, 49, and 51). Nevertheless, there were sometimes a few narrow interglacial terraces, of middle and late Quaternary age, in the upper portion of the profile, and slight breaks of slope related to glacial sea levels in the lower part.

3.2.3. Isolating the role of interglacial and glacial sea levels

The origin of the inner edge of most of the terraces that developed on rapidly rising and sinking landmasses was unambiguous, as they were within the tidal ranges of only interglacial or glacial stillstands, respectively. In some runs, however, especially with slow changes in the elevation of the land, the inner edge sometimes fell within the tidal range of several intergacial and glacial stillstands. The model was modified to allow erosion only during interglacial stages on emerging coasts, and during only glacial stages on subsiding coasts in order to:

(a) determine if any terraces on rising landmasses are produced during glacial stages, or on subsiding landmasses during interglacial stages; and

(b) to assess the effect of rising and falling sea levels on terrace width and gradient, and on the relationship between former sea levels and the elevation of the inner edge.

For convenience, the term 'normal' was used to refer to the runs with erosion by glacial, interglacial and rising and falling sea levels, and 'alternate' for runs that allowed erosion only during glacial or interglacial stillstands. The total amount of erosion accomplished in each alternate run was much less than in the corresponding normal run. Comparison of the erosional profiles produced in run 1 by the normal and alternate runs demonstrated that interglacial erosion could account for the occurrence of all the subaerial terraces on rapidly emerging landmasses (Fig. 13). In the normal runs, erosion at sea levels slightly below the interglacial maximum reduced the height of the inner edge, whereas erosion on cliffs and other steep slopes reduced the width of the terraces at higher elevations (Fig. 14). Therefore, terraces produced entirely by erosion during interglacial stillstands extended to lower elevations and they had higher inner edges than those produced under normal erosive conditions. For example, inner edges in run 1 were generally at the MLWN or MLWS palaeo-tidal levels in the normal run, and at the MT or MLWN levels in the alternate runs. Many of the terraces produced by normal erosion on rapidly rising land consisted of a series of gently sloping surfaces separated by low cliffs, in contrast to the more continuous and smoother surfaces produced entirely by erosion during interglacial stillstands.

There was also a good correlation between the occurrence of terraces created by the normal and alternate versions of run 12, which demonstrated that terraces on a slowly emerging coast are also the result of erosion during interglacial stillstands. The terraces produced entirely by erosion during interglacial stages generally consisted of a series of gently sloping surfaces separated by low cliffs, but elimination of these cliffs and terrace coalescence produced wider surfaces in the normal run.

Good correspondence between the occurrence and morphology of terraces produced in the normal and alternate runs therefore provided confirmation that they were created by erosion during interglacial stages on rising landmasses. On the other hand, the normal profiles were much further landward of the alternate profiles, which consisted essentially of interglacial terraces superimposed on the initial profile. Therefore, erosion on emerging coasts during glacial stillstands and when sea level was rising and falling was largely responsible for the landward position of the coastal slope on which the interglacial terraces developed.

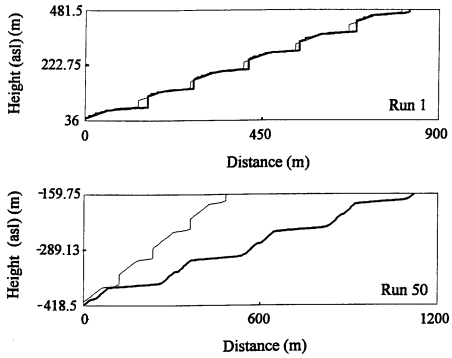

Fig. 14. Comparison of a portion of the runs shown in Fig. 13. For comparative purposes, horizontal values were adjusted so that the normal (thicker lines) and alternate (thinner lines) had a common point of origin.

Alternate runs allowing erosion only at glacial stillstands also accounted for almost all the terraces produced on rapidly and slowly subsiding landmasses (Fig. 13). The terraces produced entirely by glacial stillstand erosion on rapidly subsiding land were narrower than those formed in the normal runs, however, and, because of the absence of post-formational modification by rising sea level following glacial minima, their inner edges were generally lower (Fig. 14). For example, in run 50, inner edges were usually at the MHWN palaeo-tidal level in the normal run, and at the MT level in the alternate run. With slowly subsiding land (run 49), the alternate profile consisted of a series of terraces separated by low cliffs, whereas in the normal run, erosion over a much greater variety of elevations produced a smooth, continuous surface. The main difference between profiles produced under normal and alternate conditions was in the upper part of the profile, above the level of wave erosion during the most recent, and highest, glacial stage. Consequently, the highest terrace (at a depth of about −70 m) produced by interglacial erosion in the normal run with rapid subsidence was absent in the alternate run. Similarly, in the run with slow subsidence, the sloping surface produced under normal erosional conditions in the upper portion of the profile was replaced by a vertical cliff in the alternate run, which permitted erosion only during glacial stages.

3.2.4. The inner edge

The use of wave-cut terraces as palaeo-sea level indicators requires better understanding of the relationship between the morphology of intertidal shore platforms and the present level of the sea. Cliff–platform junctions tend to be close to the spring or neap tidal levels in stormy macrotidal environments, whereas the position of the junction on subhorizontal platforms in mesotidal environments is essentially determined by the mean elevation of the platform, unless there is a well defined ramp at the cliff base (Trenhaile, 1978). Junction elevations vary considerably, however, according to geological influences and exposure to wave attack. The model has previously suggested that cliff–platform junctions tended to be higher, relative to the tide, in high than in low tidal range environments, with resistant than with weak rocks, and with low rather than high initial slope gradients (Trenhaile, 2000, 2001c).

Although there was considerable variation, the elevation of most of the inner edges at the end of model runs corresponded to the mid-tide or mean low water spring tidal levels of the palaeo-sea levels that formed the terraces. Because of post-formational modification, it was more difficult to determine the precise elevation of the inner edge on subsiding landmasses, and most analysis was therefore conducted on rising land. Nevertheless, where it was possible to identify the position of the inner edge on subsiding landmasses, they appeared to be higher (generally near MHWN) than in runs when the land was rising. This is because on subsiding landmasses inner edges were raised by intertidal and subtidal erosion during periods of rising sea level following glacial minima, whereas on rising land, inner edges were lowered by high tidal erosion during periods of falling sea level following interglacial maxima. Consequently, abandoned inner edges on rising landmasses tended to be lower, relative to the tide, than contemporary cliff–platform junctions at the rear of shore platforms. The general occurrence of inner edges that were higher in resistant rocks (MT or MLWN) than those in weak rocks (MLWN to MLWS) is consistent with the field evidence in southern Britain (Wright, 1970; Trenhaile, 1972).

4. Discussion

It has been suggested that seaward tilting might explain why the lower, younger terraces are more gently sloping than the higher, older terraces in parts of California and New Zealand (Ridlon, 1972; Bradley and Griggs, 1976; Pillans, 1983). The model suggested that there are other possible explanations, including an increase in tidal range through time, the presence of more resistant rocks at lower elevations, variations in rates of tectonic uplift, and the occurrence of longer interglacial stages in the middle and late Quaternary than in the early Quaternary. Although there was a general tendency in the present study for the lower and younger terraces to be wider than the older terraces at higher elevation, there was no relationship between terrace gradient and elevation. This partly reflects the fact that only very slight changes in gradient are required to accommodate large increases in the width of gently inclined shore platforms and terraces (Trenhaile, 2001a). The lack of a width–gradient relationship was also the result of the greater lowering that took place on the seaward edge of widening terraces by subtidal erosion during middle and late Quaternary interglacials, than in the shorter interglacials of the early Quaternary.

The model supports the general assumption that dates derived from marine terraces on rising coasts represent interglacial stillstands (Bloom and Yonekura, 1985; Bull, 1985; Lajoie, 1986; Lajoie et al., 1991). Nevertheless, whereas interglacial terraces on rapidly rising coasts were above the level of the sea during later interglacial stages, the lower portions of terraces on slowly rising coasts were often exposed to intertidal marine processes during one or more subsequent interglacials, particularly in macrotidal environments and on terraces formed at higher elevations during the early Quaternary. Dates obtained from material in the lower portions of a terrace might therefore be significantly younger than those derived from near the back of the terrace. Similarly, whereas submarine terraces on rapidly subsiding coasts were carried below the level of wave action during subsequent glacial stillstands, the upper portions were still intertidal during one or more later glacial stages, and, in contrast to terraces on rising coasts, sediments may also be deposited on submarine terraces long after their formation.

On the assumption that successive glacial and interglacial sea levels became progressively lower through the Quaternary, Fairbridge (1977) proposed that shore platforms were cut by frost and floating ice during glacial stages when sea level was similar to today's. Although Fairbridge's proposal cannot be reconciled with more recent data on Quaternary sea levels, the present model suggests that wave action during interglacial stages on rising landmasses often occurred at elevations that had been previously occupied during glacial stillstands, whereas on subsiding landmasses, erosion during interglacial stages preceded erosion during glacial stages. Therefore, on rising landmasses in cold regions, frost and sea ice during glacial stages may have initiated terraces which were later modified during periods of rising and falling sea levels, and then, in some cases, rejuvenated by wave action during interglacial stages.

Although the sea level model helped to determine the factors that influence the formation and morphology of marine terraces, it introduced a degree of uniformity, reflected in the morphology and vertical spacing of the terraces, that does not exist in the field. The simplified model did not consider cycle to cycle variations in amplitude and wavelength, and the possibility that the transition between the more subdued glacial cycles of the early Quaternary and the more severe cycles of the middle and late Quaternary may have been gradual rather than abrupt. Nevertheless, it is possible, based on the results of this study, to predict the general effects of a more irregular sea level history on terrace formation. Highly variable glacial cycles would produce differences in the vertical spacing, width and gradient of terraces, and result in less preservation of the palaeo-sea level record. For example, unless a terrace had reached equilibrium, increasing the duration of an interglacial stage would result in greater cliff recession and the progressive elimination of terraces at higher elevations. Variable rock resistance would have a similar effect, with more rapid erosion in weaker rocks resulting in the truncation, or in more extreme situations the elimination, of terraces at higher elevation. Differences in the elevation of glacial and interglacial sea levels, which could reflect tectonic as well as climatic factors, would also produce greater variation in the vertical separation of adjacent terraces, and thereby influence whether terrace coalescence, truncation, or elimination takes place.

5. Conclusions

The main conclusions of this study include the following:

(a) On an initially steep, linear slope, more erosion generally occurs with rising than falling relative sea level. The rate of erosion increases with the rate of rising and falling sea level, although it may eventually decrease with very rapid changes in relative sea level.

(b) The intertidal width of the simulated platforms was greatest when sea level was falling, but the shelf, or subtidal surface, was much wider when sea level was rising.

(c) Terraces are formed during interglacial stages on rising landmasses, and during glacial stages on subsiding landmasses.

(d) Terraces preserve a more complete palaeo-sea level record on coasts that are rapidly rising or subsiding, than on coasts where changes in elevation are slow.

(e) The gradient of the terraces increased with tidal range, and they tended to be more gently sloping and wider in weaker rocks. The number of terraces increased with the rate of uplift and subsidence, and with the slope of the hinterland. Terraces tended to be narrower and more steeply sloping on steeply sloping landmasses.

(f) Inner edges are higher, relative to the tide, when the land is subsiding than when it is rising, and lower, relative to the tide, than contemporary cliff–platform junctions.

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