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Sea-level Change and Coastal Erosion
海水面 變化와 海岸 浸蝕

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Abstract : Time series of the relative sea levels at the selected tide-gauge stations in the North Pacific and historical aerial photographs in the Hawaiian Islands are analyzed. Long-term rising trend of sea level ranges from +1 to +5 mm/yr at most of the stations, which is primarily due to global warming and tectonic motion of the plates. The annual and interannual fluctuations of sea level result from the thermal expansion/contraction of sea-surface layer due to the annual change of the solar radiation and possibly from a coupled ocean-atmosphere phenomenon associated with an ENSO event, respectively. Sea-level changes in three different time-scales (linear trend, annual oscillation, and interannual fluctuation) and their quantitative contribution to the shoreline changes as a result of long-term cross-shore sediment transport are hypothesized.

要 旨 : 北太平洋에서 선택한 朝夕 定點에서 相對 海水面의 時系列 資料와 하와이 제도에서 海岸線 變化의 航空 寫眞을 分析하였다. 대부분의 定點에서 海面의 長期的 上昇 推移는 +1 내지 +5 mm/yr의 範圍를 보이는데, 주로 地球 溫暖化 및 地質學的 판(plate)의 移動에 의해 나타나고 있다. 海面의 年變化 및 수년 週期의 變化는 각각 太陽 輻射의 年變化에 의한 表層水의 膨脹 및 收縮과, ENSO 週期와 관계된 大氣-海洋의 相互作用으로부터 起因한다. 이러한 세 가지의 다른 時間 規模로 發生하는 海面 變化(장기적 해면상승 추이, 연변화, 수년주기 변화)가 長期的으로 離岸 堆積物 輸送의 결과로서 나타나는 海岸線 變化에 어떻게 定量的으로 寄與하는지 推定하는 假說이 提示된다.

1. INTRODUCTION

Relative sea-level changes, both in long- and short-term scales, play a significant role in producing beach erosion and in changing other coastal processes such as longshore and cross-shore sediment transport, nearshore circulation, and wave transformation. The major causes of relative sea-level change can be primarily classified into two categories: eustatic (or geoidal) sea-level changes and isostatic land-level changes.

Climatic variations may cause volume changes in ocean water due to steric effects, melting/freezing of land ice or freshwater input from other sources like lakes and groundwater. Relative sea-level changes have occurred with different time scales.

During the past several hundred years, the Little Ice Age (between about 1,500 and 1,800 A.D.) created a dramatic fluctuation in global climate by lowering temperatures (IPCC, 1990). The earth is still recovering from this Little Ice Age. The prominant climatic fluctuation today is the El Niño Southern Oscillation (ENSO) a coupled ocean-and-atmospheric global event with a period of about 2 to 7 years, which shows its strongest effects in the Pacific Ocean. El Niño, originally named from an abnormal warming phenomenon in the southeastern Pacific Ocean around the coast of Peru, is now recognized as a large-scale global teleconnection. The Southern Oscillation relates the pressure fluctuation in the southeastern Pacific to alternating high and low pressure fields over the southern Asia (Fig. 1). The interannual variability of sea levels in response to ENSO may cause even greater impacts on coastal

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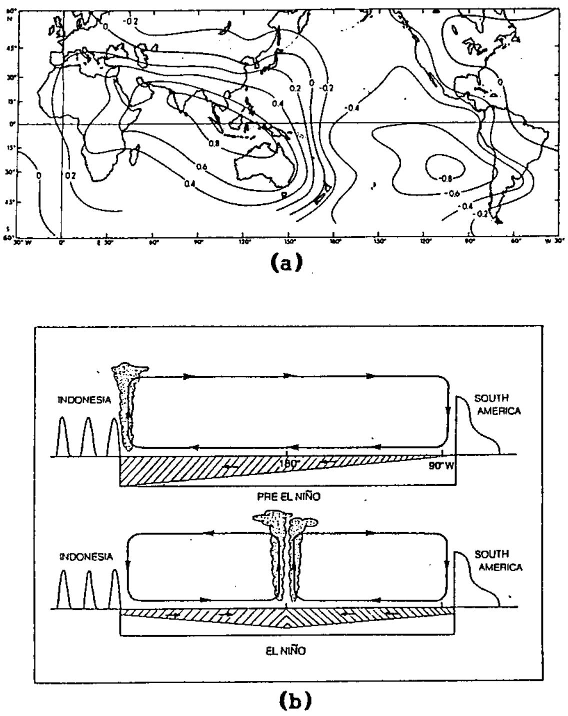


Fig. 1. (a) Southern Oscillation expressed as the correlation of the annual-mean atmospheric pressure with Jakarta, and (b) schematic diagrams of the air-sea coupling before and during El Niño (after Wyrtki, 1975).

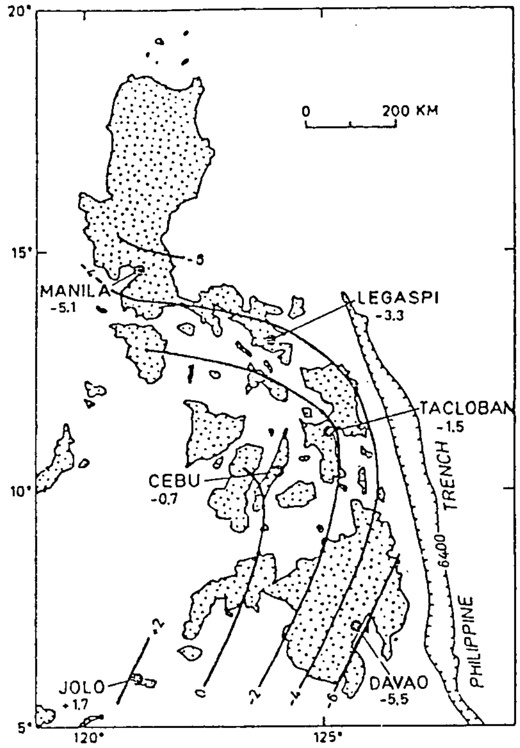
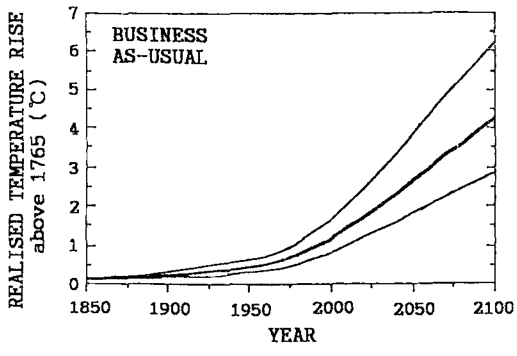


Fig. 2. Tilting of the Philippine Islands indicated by relative land movement at 6 tide-gauge stations with 2 mm/yr intervals. (−) represents land subsidence, and the data are from PSMSL (after Emery and Aubrey, 1991).

erosion (see Komar and Enfield, 1987) than long-term trend of sea-level rise. For instance, the 1982-83 El Niño raised the water level on the coast of Oregon by 60 cm within 12 months, which changed the shape of an inlet there due to erosion (Komar, 1986).

Atmospheric pressure and wind stress may be the major factors for changes in water on time scales of days to months. In practice, the inverse barometric response of sea level to the atmospheric pressure change is found to be a significant part of seasonal changes in high latitudes (Lisitzin, 1974). An increase of atmospheric pressure by 1 mbar ($=100$ Nm^{-2}) causes the depression of sea level approximately by 1 cm. Tropical cyclones or hurricanes result in peaks of mean sea level (MSL) by the combination of inverse barometric response and friction. Since hurricanes or typhoons are maintained by the energy that they extract from the heat in the ocean, thermal effects are very important and the input energy is redistributed within the ocean by stirring action of the storm in addition to the advective effects (Gill, 1982).

Isostatic land levels are changed by tectonic activities such as volcanism, sea-floor spreading, faults and subduction at the plate boundaries, glacial rebound, sediment compaction, land subsidence, and hydroisostacy on islands and continental margins. At active plate margins such as East Pacific Rise or Mid-Atlantic Ridge, volcanic eruption from below the lithosphere, and the transform faults, may change the volume or shape of the local land masses. Due to the changes in spreading rate of the sea floor or different spreading rates of the plates, sea levels may be affected, although the effect is small over time scales of less than a million years. At the active margins of the plates such as east of Japan and the Philippines, two plates collide and the oceanic plate is subducted beneath the contine-



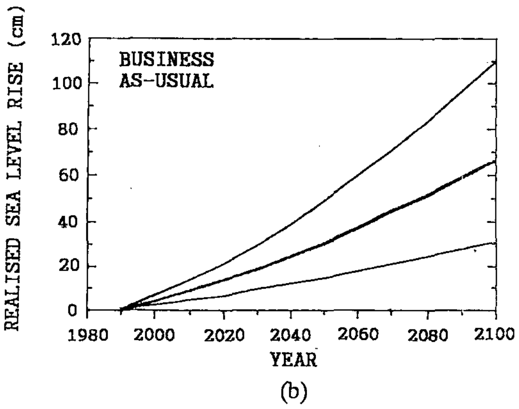


Fig. 3. (a) Simulation of the increase in global mean temperature from 1850 to 1990 due to observed increases in greenhouse gases and predictions between 1990 and 2100, and (b) the predicted sea-level rise showing the best estimate and range, resulting from the Business-as-Usual emmisions (after IPCC, 1990).

ntal plate, resulting in earthquakes and vertical movements from deformation of the land masses. Land tilting by different subsidence rates between eastern and western sides of the Philippines is shown in Fig. 2.

There is little doubt that global MSL has risen by 10 to 30 cm during the past century. Some local tide-gauge records nonetheless have deviated from the global mean from about 100 cm rise in 20 years in Tribeni, India, to about 160 cm fall in 14 years in Rorvik, Norway (see Appendix I in Emery and Aubrey, 1991). One of the most important factors in determining global warming during the past century is the 'additional' greenhouse effect: increasing the surface air temperature by enhanced trapping of longwaves radiated back from the earth's surface due to the increase of greenhouse gases such as CO_2, CH_4, N_2O, and CFCs (chlorofluorocarbons) in the atmosphere. The most important factors causing the greenhouse effect in the atmosphere is water vapor, of which the concentration in the troposphere is determined internally within the climate system and is not affected by human sources and sinks on a global scale. The concentration of ozone (O_3) as a second major greenhouse gas is changing in the atmosphere due to human activities, but the lack of adequate observations prevents us from accurately quantifying the climatic effect of the changes in tropospheric ozone (IPCC, 1990).

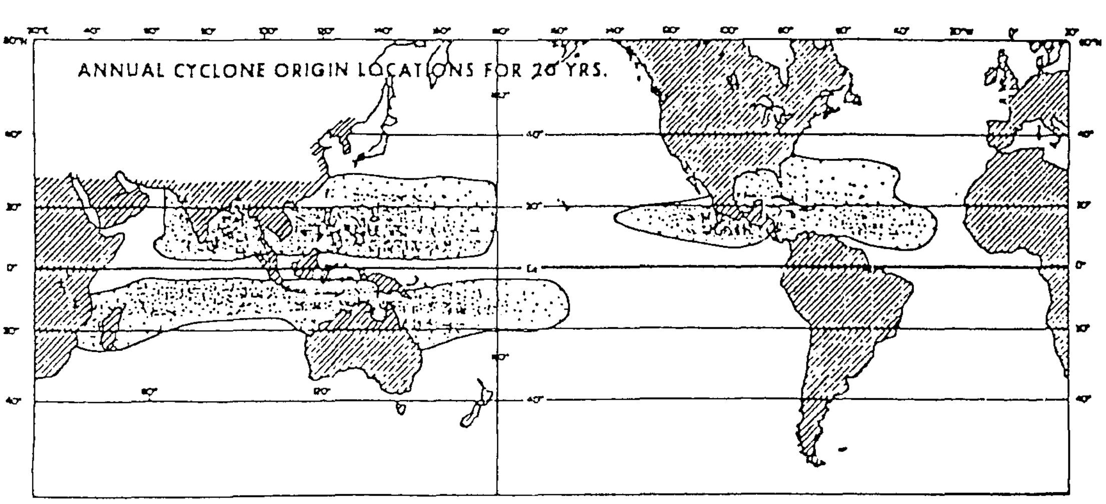


Fig. 4. The location of the points of origin of tropical cyclones during a 20-year period (1958-1977)(after Gray, 1979).

Recently, the IPCC suggested that the average rate of the increase of global mean temperature and the corresponding global MSL rise are estimated to be about 3℃ (from 2℃ to 5℃) and about 60 cm, respectively, before the end of the next century, under the 'Business-as-Usual Scenario' (Fig. 3). There will be significant regional deviations due to different contributing factors to sea-level rise.

Shoreline changes due to cross-shore and longshore sediment transports are the response of the land to the oceanic forcings such as sea-level rise and fluctuations on different time scales, storm surges and wind waves on short time-scales, etc.

Disastrous damages as well as coastal erosion may come from the seismic sea waves (or tsunamis) but the frequency of them is relatively low when comparing with tropical storms or typhoons. The frequency of tropical cyclones was found a world average of 80 events per year during a 20 year period (Fig. 4). When tropical storms over very warm surface water (>26℃) are rapidly amplified into intense vortical storms by a certain mechanism[1] under favorable conditions of moisture supply, they become tropical cyclones such as hurricanes or typhoons (Holton, 1979). The hurricanes occur more likely during summer season. In fact, more than 70% of the total tropical cyclones of the Northern Hemisphere during 1958 to 1977 were reported between July and October (Gray, 1979).

Wave-induced currents, longshore current and rip current, form a nearshore circulation system and contribute to coastal erosion or accretion. Longshore currents are induced by the radiation stress[2], which result in longshore transport of sediments. The quantitative estimate of sediment transport by longshore currents was studied by several arthors (see Bijker, 1967; Komar, 1976; Bailard, 1981, CERC, 1984). Rip currents play a role of offshore transport of the sediments out of the breaker zone, which forms a part of the nearshore circulation. They often occur more clearly when wave heights are high and where the coastline or bottom topography is severely undulated or where the submarine channels are developed offshorewards. The cross-shore sediment transport is associated with the changes in beach profile, which is usually affected by the changes in wave conditions. During high and steep wave conditions, sediments move from onshore to offshore side due to relatively stronger return flow (or undertow), and longshore bars are often formed. During low and less steep wave conditions, sediments move in a reverse direction. Asymmetry of wave form in shallow water (or Stokes' wave) induces shoreward transport, i.e., materials move from the longshore bar(s) to onshore side, and thus the foreshore steepens. A berm often develops at the uprush level. Return flow occurs below wave trough forming a vertical circulation cell due to the conservation of mass, while rip currents form a part of a horizontal circulation cell. Thieke and Sobey (1990) considered the conservation of mass, momentum, and energy to describe the cross-shore mean-flow circulation with the horizontal flow velocity in terms of three components: a time-averaged quantity, turbulence and waves. They show that the return flow is rather less intense and directed offshorewards while wave-induced mass transport velocity between wave crest and trough is relatively intense and directed to the beach.

2. SEA-LEVEL CHANGES IN THE PACIFIC OCEAN

Sea-level changes in the ocean may be decomposed into several signals: long-term trend, interdecadal and interannual fluctuations, and annual, fortnight, and daily oscillations. Daily oscillations may be classified into three different types (diurnal, mixed, and semidiurnal), based on the relative magni-

[1]Conditional Instability of the Second Kind (CISK): the process of the cooperative interaction between the cumulus convection and a large-scale perturbation leading to unstable growth of the large-scale system.

[2]The excessive stress integrated over the water depth under the existence of waves, which has the same unit as that of wave-energy flux [Nm^{-1}].

[3]the form ratio of the sum of the amplitudes of the two main diurnal constituents (K_1+O_1) of the actual tide to that of the two main semidiurnal amplitudes (M_2+S_2).

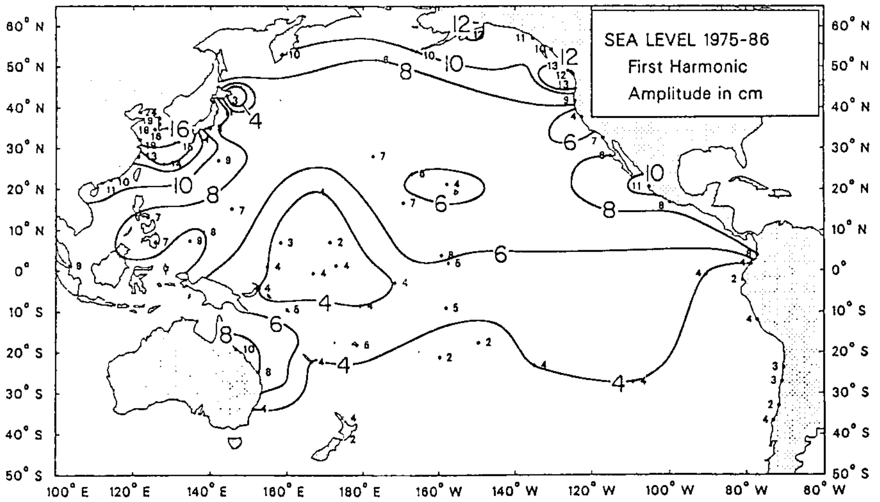


Fig. 5. Amplitude of the first harmonic of the mean annual variation of sea level in cm. (after Wyrtki *et al.*, 1988).

tudes of four major tidal constituents[3]. Fortnight oscillation of sea levels results from the revolution of the moon: spring tides occur when the earth, the moon and the sun stand along a line, and neap tides are when the three celestial bodies make the right angle. The direct effect of these short time-scale oscilations on coastal erosion seems to be insignificant.

One of the most obvious and consistent features of sea-level changes in the world oceans is an annual oscillation, which is primarily related to the expansion and contraction of the sea-surface layer due to the annual variation of the solar radiation by the change of the sun's altitude. The annual difference of mean monthly sea-level values is approximately 10 cm on the island stations of the Pacific, and increases up to about 40 cm on the coastal stations of the Yellow Sea (Fig. 5). It is because they may include the annual variation of the monsoons as well as the steric effect. In addition, the annual and interannual variations of the atmospheric pressure (or wind) system and of the current system are certainly associated with the annual and

[3]Photographic Scale (Q) = f/h:
where f = the focal length of the camera lens, and
h = the flight altitude of the camera above the mean elevation

Table 1. Mean monthly sea-level values and the standard deviations of the annual cycles at the four major stations in the Hawaiian Islands. Units are in cm, and the mean values are relative to the minimum values

month	Nawiliwili Kauai	Honolulu Oahu	Kahului Maui	Hilo Hawaii
Jan	4.9± 6.3	3.0± 6.8	4.1± 5.6	4.3± 7.6
Feb	3.7± 6.4	2.4± 6.5	3.2± 4.7	2.2± 6.9
Mar	2.1± 5.8	1.3± 5.9	1.8± 4.3	1.8± 7.1
Apr	1.1± 5.2	0.0± 6.1	0.3± 4.5	0.9± 6.2
May	0.0± 5.0	0.3± 5.8	0.0± 4.5	0.0± 6.8
Jun	0.7± 5.4	1.2± 6.2	1.8± 6.0	0.7± 7.3
Jul	4.7± 6.4	4.1± 6.8	5.2± 5.8	5.0± 7.1
Aug	8.5± 6.7	7.1± 7.0	8.1± 5.4	9.6± 7.6
Sep	10.7± 5.5	8.6± 6.3	10.5± 5.2	10.4± 7.4
Oct	9.8± 5.0	8.3± 6.2	9.5± 5.1	10.0± 7.4
Nov	7.9± 4.0	6.6± 6.2	7.9± 5.3	8.2± 6.8
Dec	5.9± 4.8	4.6± 6.4	5.1± 5.0	5.3± 7.2

interannual fluctuations of sea levels. On the island stations of the central North Pacific where the steric effect is the only dominant factor of the annual cycle of sea levels, the minimum and maximum values appear in May and in September, respectively (Table 1 and Fig. 6). The annual pattern of the mean monthly MSL height is quite different in the western Pacific: the maximum sea level height (SLH) moves from August on Wake and Eniwetok, July on Guam to April on Truk and Kwaja-

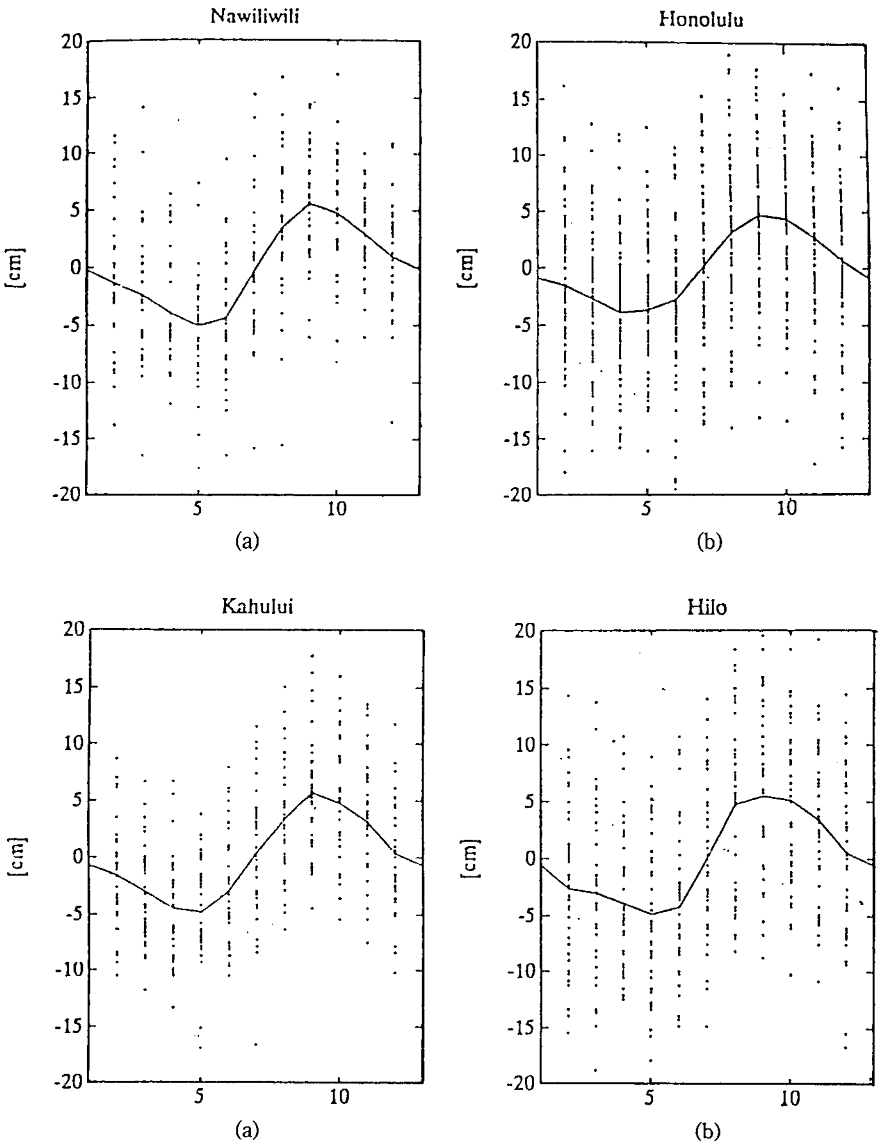


Fig. 6. Mean annual cycles of monthly sea levels at (a) Nawiliwili, Kauai, at (b) Honolulu Horbor, Oahu, at (c) Kahului, Maui, and at (d) Hilo, Hawaii. Data are from PSMSL.

lein. The minimum SLH on these stations in the western Pacific occurs in January except on Wake Island (Fig. 7).

The interannual variation of the water movement may be seen from the time series of the sea-level anomalies deviated from the MSL The linear trend and the interannual fluctuation of sea levels can also be seen from the least-square fitting and the curve with 12 month running means. The linear trend of sea levels could be reliable as long as the

Table 2. The height of sea-level fluctuations and the rate of sea-level rise with their standard deviations on the stations of the Hawaiian Islands

station	annual	interannual	linear trend
Nawiliwili, Kauai	10.7± 5.5 cm	11.7± 3.1 cm	1.8 cm/decade
Honolulu, Oahu	8.6± 6.3 cm	12.4± 4.3 cm	1.6
Kahului, Maui	10.5± 5.2 cm	10.8± 2.0 cm	2.2
Hilo, Hawaii	10.4± 7.4 cm	12.9± 2.1 cm	4.0

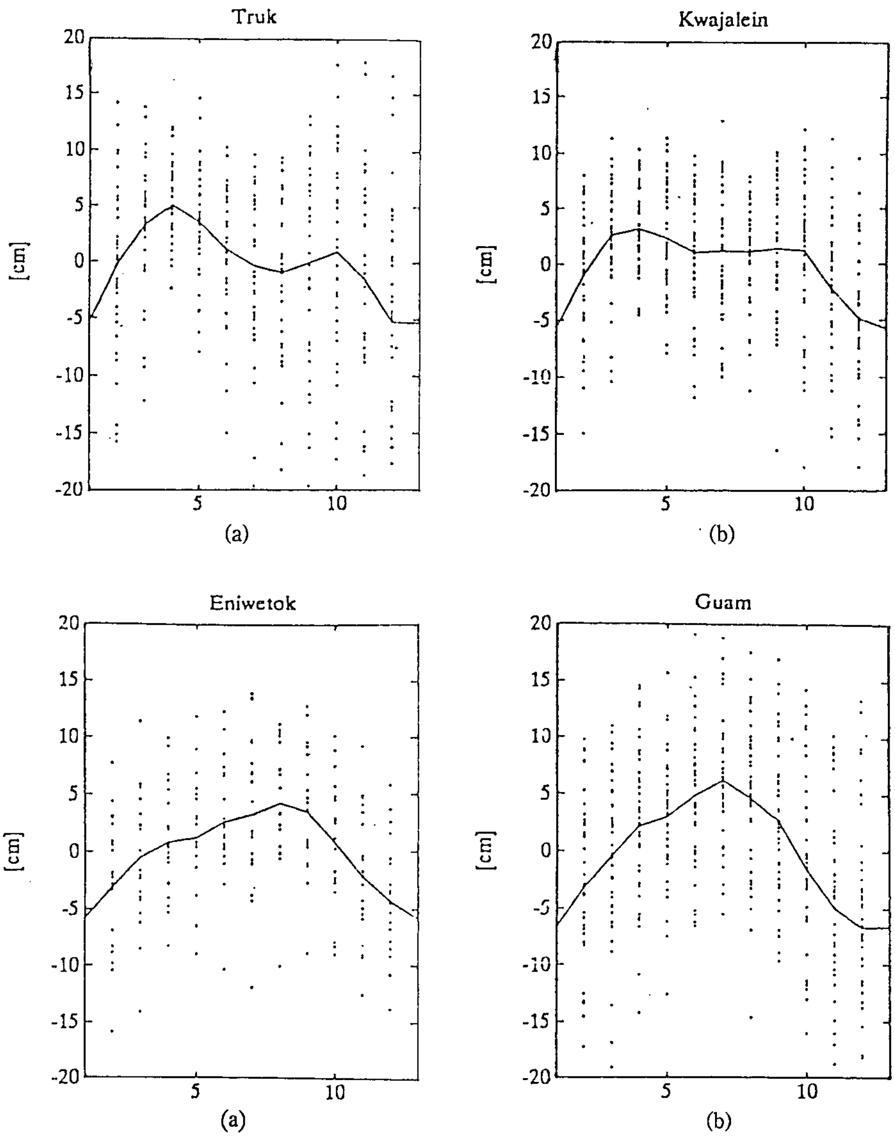


Fig. 7. Mean annual cycles of monthly sea levels at (a) Truk, at (b) Kwajalein, (c) Eniwetok, and at (d) Guam. Data are from PSMSL.

data length is longer than 20 years (personal communication with Wyrtki, 1988). The overall rate of the linear trend of relative MSL rise is about 1 to 5 mm/yr, which includes not only eustatic sea-level rise by global warming but also isostatic land-level fall by volcanic loading or other tectonic movements. In the Hawaiian Islands of the central North Pacific, the islands have been isostatically sinking due to volcanic loading at Kilauea, Hawaii. The difference of sea-level fluctuations and the rate of sea-level rise on the stations of the Hawaiian Islands are shown in Table 2. Hwang and Fletcher (1992) estimated the island subsidence and the projected future submergence rates from the same PSMSL data (Table 3). Their estimates of the linear trend from tide-gauge records are about the same as those analyzed in this study.

In the Hawaiian Islands, the linear trend of sea levels decreases from Hilo (+4.0 mm/yr), Hawaii to Honolulu (+1.6 mm/yr), Oahu or Nawiliwili (+

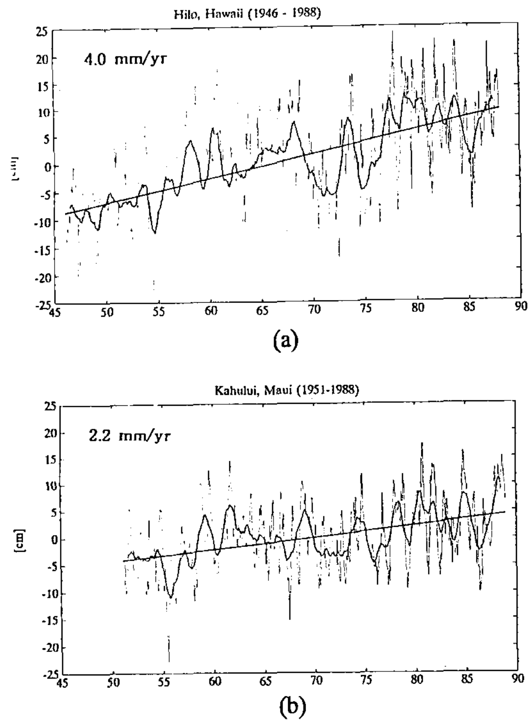


Fig. 8. Long-term linear trend (solid), 12-month running mean (solid curve), and monthly sea-level values (dotted curve) at (a) Hilo, Hawaii, (b) Kahului, Maui, (c) Honolulu, Oahu, (d) Nawiliwili, Kauai, and at (e) Midway.

Table 3. Sea-level trends of rise in the Hawaiian Islands. The numbers after± represent standard deviations. The unit is in cm/decade, based on the scenario by IPCC (1990) (after Hwang and Fletcher, 1992)

station	net submergence	subsidence rate	future submergence
Nawiliwili, Kauai	1.75± 0.30	0.33± 0.61	6.32± 4.17
Honolulu, Oahu	1.57± 0.08	0.15± 0.38	6.15± 3.94
Kahului, Maui	2.46± 0.23	1.04± 0.53	7.04± 4.09
Hilo, Hawaii	3.94± 0.23	2.51± 0.53	8.51± 4.09

glu *et al.*, 1983), which is comparable to the rising trend of relative MSL obtained from the tide-gauge records at Honolulu, Oahu.

Hwang and Fletcher (1992) assume that the subsidence rate of the islands is the net submergence from the records minus the average (=0.42± 0.30 cm/decade) of the different estimates of global MSL rise. The corresponding island-specific subsidence rate then becomes 19% of the total submergence rate at Nawiliwili, Kauai, 10% at Honolulu, Oahu, 42% at Kahului, Maui, and 64% at Hilo, Hawaii.

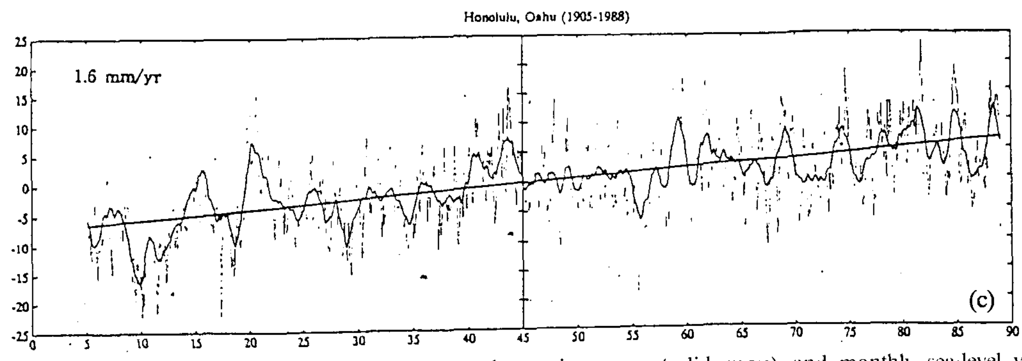
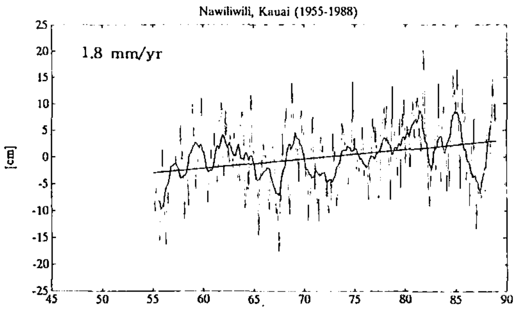


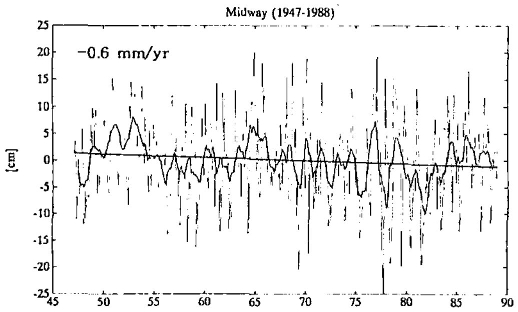
Fig. 8. (continued) Long-term linear trend (solid), 12-month running mean (solid curve), and monthly sea-level values (dotted curve) at (a) Hilo, Hawaii, (b) Kahului, Maui, (c) Honolulu, Oahu, (d) Nawiliwili, Kauai, and at (e) Midway.

1.8 mm/yr), Kauai (Fig. 8). Continual volcanic loading on Kilauea, Hawaii results in the subsidence of not only the island of Hawaii but also the neiboring islands due to isostacy. The subsiding rate of the island of Oahu was estimated as about 1 to 2 mm/yr from te shallow core measurements of a fringing reef crest at Hanauma Bay, Oahu (Nakibo-

And the projected submergence rate is assumed to be the island-specific subsidence rate plus the projected rate of future global sea-level rise, according to the IPCC estimate scenario (=6.0± 3.6 cm decade) (IPCC, 1990). If applying this scenario for the next several decades, the relative sea-level rise would reach up to 30 to 40 cm (± 20) cm) on the major



(d)



(e)

Fig. 8. (continued) Long-term linear trend (solid), 12-month running mean (solid curve), and monthly sea-level values (dotted curve) at (a) Hilo, Hawaii, (b) Kahului, Maui, (c) Honolulu, Oahu, (d) Nawiliwili, Kauai, and at (e) Midway.

Hawaiian Islands in 2040. Therefore, it is encouraged that the engineers constructing coastal structures in the Hawaiian Islands should adjust the design water level due to sea-level rise at least by +50 to +60 cm, in addition to the other factors, for the life-time of projects of about 50 years.

The interannual fluctuation of sea levels often seems to be associated with El Niño events. The onset of an El Niño event, if we define it as the year warm water appeared in the equatorial eastern Pacific, between 1950 to 1985 were reported in 1957, 1963, 1965, 1969, 1972, 1976, 1979, 1982 (Graham and White, 1988). The total period of the interannual rising trends is roughly about 40% of the total sea-level records at Honolulu, Oahu and about 50% of the the total records at Hilo, Hawaii. The vertical range of the interannual fluctuation is from 8 to 20 cm during the rising phases of sea level, and the corresponding rising rate is from 1 to 9 cm/yr (Table 4). Most of the rising phases match with

Table 4. Interannual rising trends from 12 month running means of the monthly MSL records (a) at Hilo (1947-1988), Hawaii, and (b) at Honolulu Harbor (1905-1988), Oahu

(a)

period	total rise [cm]	rising rate [cm/yr]
1955-1958 (4 yrs)	11-15	2.75-3.75
1962-1968 (7 yrs)	8-10	1.14-1.43
1972-1973 (2 yrs)	12.5	6.25
1975-1977 (3 yrs)	10	3.33
1985-1988 (2.5 yrs)	8.5	3.40

(b)

period	total rise [cm]	rising rate [cm/yr]
1910-1916 (6 yrs)	20	3.33
1918-1920 (2 yrs)	17.5	8.75
1929-1930 (2 yrs)	10	5.0
1935-1936 (1.5 yrs)	7	4.67
1939-1944 (3.5 yrs)	11.5	3.29
1955-1959 (4 yrs)	17	4.25
1960-1962 (2 yrs)	9	4.5
1967-1969 (2 yrs)	10	5.0
1973-1974 (2 yrs)	10	5.0
1976-1981 (5.5 yrs)	12	2.18
1986-1988 (3 yrs)	11	3.67

the occurrences of El Niños. But it is rather surprising that the signal of 1982-83 El Ni o, despite its strong intensity, is not clear in the records at Hilo, Hawaii.

3. SHORELINE CHANGE AND BEACH RECOVERY

Long-term (or interdecadal) change of shoreline can be interpreted from aerial photographs of shoreline. The traditional way to interpret aerial photographs is to determine vegetation line or water line by eye, which involves some level of subjectivity and inconsistency in the detection process since the coastline as a transition zone between land and sea forms a complex system of brightness variation in the panchromatic spectrum (Shoshany and Degani, 1992).

In general, errors come from two typical sources; the one is due to photographic processes such as lens distortion, camera tilt, film development, differential scale change from the center to the margins,

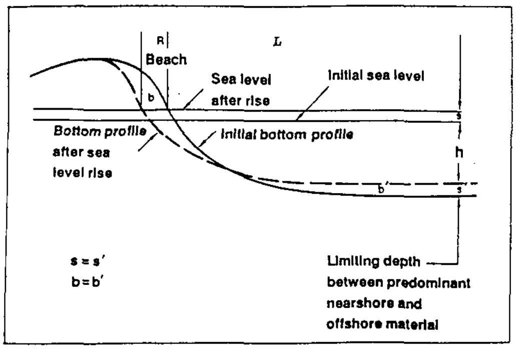


Fig. 9. Schematic diagram of the relationship between sea-level rise and coastal erosion by the Bruun Rule.

and relief displacement, etc. The other is due to geometric adjustment between photographs and base map.

Lens distortion and camera tilt may result when an airplane and camera are not exactly parallel to the mean plane of the earth's surface at the instant of exposure. Scale change may result from the shifts in altitude along the photographic flight line, especially with light aircraft. The possible decrease of the aircraft elevation may significantly increase the scale of the air photos. Thus exact scale should be determined for digitizing data from each air photo. Correct interpretation of the high water line (HWL) and careful annotation are required to avoid large miscalculations. In order to minimize error associated with shoreline annotation, photographic scale[4] should be maximized, that is, measurement altitude must be low and camera lens should be large. Relief displacement changes radially from the center of the photo, which is coincident with the *nadir* point, the point vertically below the camera, for a truly vertical photo. Since most coastal features have low relief, radial distortion due to elevation differences is not serious.

One of the most important factors to reduce errors in shoreline changes from aerial photographs, in addition to the technical difficulties described above, is that the measurement schedule should be set up for the time of HWL of a day and at least for the same month of a year. HWL is now commonly accepted as the most distinguishable boundary between land and water in image processing. In order to carefully analyze shoreline recession from two air photos taken at the same site with several year interval, the two photos should be the ones taken in the same month (or strictly speaking, on the same Julian day) of a year due to the annual cycle of sea levels. If two photos were produced in different season each other, they must be calibrated with the information of the bottom slope of the beach and of the annual cycle measured at the nearest tide-gauge station[5]. That is, the vertical change of sea level between two different months or seasons estimated from the annual cycle can be incorporated with the bottom slope of the beach to calibrate the horizontal shoreline change.

Bruun (1962) proposed a simple geometric relationship between sea-level rise and cross-shore sediment transport (Bruun Rule) (Fig. 9). It is assmued that

1) the upper beach is eroded due to the landward translation of the profile,

2) the material eroded from the upper beach is transported and deposited offshore,

3) the rise in the nearshore bottom as a result of this deposition is equal to the rise in sea level, thus maintaining a constant water depth.

When the gradient of longshore sediment transport is negligible, the eroded volume of sediments must be balanced by the volume to elevate the active width of the profile (L), and a vertical distance, i.e., sea-level rise (S), such that

$$R=\frac{L}{B+h}\times S\approx\frac{S}{\tan\phi} \tag{1}$$

where R = the shoreline retreat, S = sea-level rise, L = cross-shore distance to the water depth (h) of effective, motion, B = berm height, and $\tan\phi=(B+h)/L$ = average slope of the nearshore along the cross-shore width L.

The depth of effective motion (h) is generally meant 'effective vertical dimension over effective horizontal dimension of active profile'. The basic assumption of the Bruun Rule is the existence of an 'equi-

[5]When the tide-gauge station is less than 2 km apart, the accuracy of monthly sea level lies within 2 cm, which can be considered the environmental noise (Wyrtki *et al.*, 1988).

librium beach profile', which is maintained or eventually achieved following a change in water level. The concept of equilibrium profile may be defined as an idealization of conditions which occur in nature for particular sediment characteristics and steady wave conditions (Dean and Maurmeyer, 1983). Bruun proposed an equilibrium beach profile given by

$$h(x)=Ax^{m} \quad (2)$$

where $h(x)$ = water depth, x = horizontal distance from the shoreline, A = dimensional shape parameter.

Dean (1977) showed that the above equation with m = 2/3 was consistent with uniform energy dissipation per unit volume across the surf zone by applying least-square fitting to about 500 beaches along the U.S. east coasts and the Gulf of Mexico. The Bruun Rule is basically a simple equilibrium profile model that assumes the constant profile bounded by a closure depth. The closure depth (or a depth of effective sediment motion) is a critical assumption such that there is no net transport of sediment beyond the depth. The offshore current velocity capable of transporting sediment beyond the closure depth was assumed as an insignificant and probably slow process (Bruun, 1962). But Pilkey *et al.* (1993) claimed that the concept of a closure depth is not valid and they suggest the 'regime profile' concept to fit reality for certain beaches. In reality, the depth of effective sediment motion may depend on the time scale of interest. For longer time-scale of interest, not only waves but also sea-level fluctuations should be considered as forcing functions to the beach erosion. Moreover, the longer the period of sea-level fluctuations the more effective it may be on beach erosion if longshore sediment transport and all the other sediment-budget terms are negligible, that is, if cross-shore transport is the only significant erosion/accretion process. Dean and Maurmeyer (1983) suggested the modified Bruun Rule by considering the third dimension in the longshore direction:

$$P(B+h)R = L \times S + G_B \quad (3)$$

where P = decimal fraction of eroded material that is compatible with the surfzone sediment, and G_B = sediment-budget term including contributions from rivers or offshore, losses due to sediment being blown inland or transport offshore, and the longshore gradient of the littoral drift.

The amount of littoral sediment drived from shoreline recession is balanced by the sum of the quantity required to maintain the equilibrium profile relative to a sea-level rise ($L \times S$) and the sediment-budget term (G_B). As long as we can estimate all the sediment-budget terms quantitatively, their idea seems to be reasonable, but the reality may not allow us to correctly estimate both longshore and cross-shore transports in a long-term scale. Therefore, the relationship between sea-level rise and coastal erosion must be examined at least for an isolated beach or a *littoral cell* where there is little net longshore sediment transport to and from neighboring beaches. Some authors (Edelman, 1968; Dean, 1977; Kriebel and Dean, 1992) applied the Bruun Rule for short time-scale change of beach profiles. Komar *et al.* (1991) argue that, if water level is rapidly changed such as an abnormal sea-level rise during an El Niño, the response of the beach may

Table 5. The location and the number of transects (or number of sites) of the aerial photographs in the four major Hawaiian Islands. Data selected (or averaged) from Makai Ocean Eng. (1991) and Hwang (1981)

island	location	number of photos	number of transects	period
Kauai	KA (southwest)	4	11	1953-1988
	KB (southeast)	4	9	1953-1988
	KC (east)	3-4	36	1950-1988
	KD (northeast)	3	13	1960-1988
	KE (north 1)	4	10	1950-1988
	KF (north 2)	3-4	6	1950-1988
Oahu	north	3-7	82 (12)	1928-1979
	windward	4-7	133 (14)	1928-1980
	south	3-6	52 (11)	1928-1979
	leeward	4-6	41 (10)	1949-1979
Maui	MA (west)	3-4	25	1949-1988
	MB (southwest)	4	19	1949-1988
	MC (southwest)	4-5	23	1949-1988
	MD (north)	4	21	1050-1988
Hawaii	HA (northwest)	4	11	1950-1988
	HB (west)	4-6	18	1950-1988

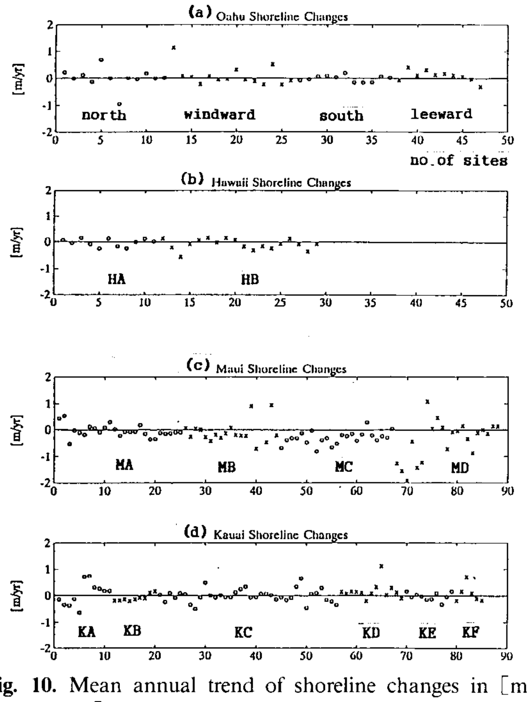


Fig. 10. Mean annual trend of shoreline changes in [m/yr] on (a) Oahu, (b) Hawaii, (c) Maui, and (d) Kauai. Data are selected from Hwang (1981) for (a), and Makai Ocean Eng. and Sea Eng. (1991) for (b), (c), (d).

be too slow to maintain the equilibrium profile. Despite the controversy of the validity of a closure depth in an equilibrium profile model and of the justification for investigating short-term variations, it has been applied for the design of many coastal engineering projects on the continental beaches.

Sea-level rise never occurs in a steady manner but in fluctuations with different time-scales. In fact, interannual fluctuation of sea level seems to be very important to coastal processes and shoreline changes since the fall after the rising phase of sea level with interannual time-scale cannot usually restore the lost amount of the sediments as is the case of the annual oscillation of the cross-shore transports. There was a report that some of the sediments transported offshorewards even in an annual oscillation were permanently lost to the deep offshore region (Campbell and Hwang, 1982). Hence, the computation of only long-term shoreline recession by the Bruun Rule definitely results in underestimate of the real amount of coastal erosion.

The average yearly trends of accretion or erosion of some beaches on the Hawaiian Islands are reanalyzed by taking into account the variation of the shoreline position due to the annual cycle of sea level (Table 5). The calibration of shoreline change by interannual fluctuation of sea level was not carried out in this study because of the limited knowledge about the exact amplitude and period of the occurrence and about the relationship of it with a possibly associated ENSO event. The most serious erosion problems were reported on the Island of Maui, especially on the north shore (Fig. 10). More than 70% of the total number of transects were eroding. On the Island of Oahu, Waimea Bay Beach had the highest recession rate of the shoreline (-0.95 m/yr). Meanwhile, maximum accretion was reported on Kahuku Golf Course Beach ($+1.14$ m/yr) as the result of sand dumping. On the Island of Kauai, erosion was reported at about half of the transects. On the west to northwest beaches of the Island of Hawaii, approximately 60% were reported as eroding.

Beach processes are basically a three-dimensional behavior which includes longshore and cross-shore transport processes. Longshore sediment transports are generally carried by the wave-induced longshore currents, which show maximum speed near a wave-breaking depth (Longuet-Higgins, 1972). In many beaches of the Hawaiian Islands wave breaking often occurs a few times instead of once over the long reef-bottoms, i.e., highly dissipative and complex shallow bottoms. Cross-shore sediment transports result from the combined effect of onshore transport above the wave trough due to the asymmetry (nonlinearity) of the wave form and offshore transport below the wave trough to the bottom due to the return flow. During storm-wave conditions, offshore transport of sediments dominates due to stronger return flow. Low and mild waves cause sediments to move shorewards and often develop a berm at the uprush level (Gerritsen and Jeon, 1991). But, after a storm, subsequent mild waves recover the beach only at shallow depths since small wave heights may not sufficiently affect the sediments at deeper depths to which they are transported by big storm waves. The recovery process is much slower than the erosion process. Therefore, beach profiles in general never reach a full equilibrium state within such a short time-scale due to the continuous variability of the wave-energy level.

Alternate onshore and offshore sediment transports usually occur in an annual time scale, depending on favorable (or mild) and unfavorable (or high) wave conditions during different seasons. When favorable wave conditions coincide with the lowering phase of the annual cycle in sea level, beach recovery (or onshore transport) rates may be faster. This implies that the rising phase of sea level may result in more erosion or more offshore transport of the eroded materials to greater depths when the rising sea level coincides with unfavorable wave conditions. In general, cross-shore sediment transport in an annual time-scale seems to occur alternately between summer winter seasons, and beach erosion and accretion (or recovery) rates seem to be comparable on the selected beaches in the Hawaiian Islands during 1962 to 1963 (Moberly, 1968). In an interannual and long-term scales, many of the Hawaiian beaches do not totally recover. In other words, they are chronically losing sands to deep-water region. Moberly (1968) argues about the mechanisms of sand loss to deep water on the Hawaiian beaches, which are 1) by waves and currents mainly along sand-bottomed channels, 2) by sedimentation to deep water, 3) by abrasion, 4) by cemetation to beachrock, and 5) by storms.

Although waves and currents are the primary mechanisms of littoral sand transports, beach erosion might not have been so serious without continuous rise and fall of sea level in an interannual time scale because an eleveated sea level makes it possible for waves to transport more sediments to deep-water region which can never be fully returned even after sea level falls. As the time scale of sea-level fluctuation becomes longer, the rate of beach recovery seems to be lower. Thus, it is hypothesized that the 'beach recovery rate' (χ) exponentially decreases as a function of the period of sea-level fluctuation such that

$$\chi = B \times \exp\left(-\frac{t}{T_c}\right) \quad (4)$$

where B = beach-erosion coefficient (≈ 1 for sandy beach) as a function of underlying geology and sediment characteristics, t = period of sea-level fluctuation (in month), and T_c = characteristic time-scale of the chronic beach erosion

Table 6. Contributions of sea-level changes and waves to shoreline retreat on (a) the Hapuna Beach, Hawaii, and on (b) the Waimea Bay Beach, Oahu, based on the hypothesized beach recovery rate and historical aerial photographs

(a) Hapuna Beach, Hawaii ($\beta = 1/15$)

	SLR (vertical)	E/A (horizontal)	Net Retreat
long-term trend	0.4 cm/yr	−0.06 m/yr	−0.06 m/yr
interannual fluctuation	14 cm/4 yr (3.5 cm/yr)	−0.53 m/yr +0.32 m/yr	−0.21 m/yr
annual oscillation	10 cm/yr	−1.50 m/yr +1.35 m/yr	−0.15 m/yr
short-term effects			−0.37 m/yr
Total			−0.79 m/yr

(b) Waimea Bay Beach, Oahu ($\beta = 1/20$)

	SLR (vertical)	E/A (horizontal)	Net Retreat
long-term trend	0.2 cm/yr	−0.04 m/yr	−0.04 m/yr
interannual fluctuation	16.8 cm/4yr (4.2 cm/yr)	−0.84 m/yr +0.50 m/yr	−0.34 m/yr
annual oscillation	9 cm/yr	−1.80 m/yr	−0.18 m/yr +1.62 m/yr
short-term effects			−0.39 m/yr
Total			−0.95 m/yr

For the case studies at the separate sandy beaches in the four major Hawaiian Islands (Hwang, 1981; Makai Eng. & Ocean Eng., 1991), T_c was applied as 100 months ($\approx$8.3 years), based on the aerial photographs. This value is approximately twice the average ENSO period. Even in long-term scale, shorelines may fluctuate back and forth depending on the changes in sea level and wave climate. Therefore, it is important in this hypothesis that the time scale of chronic beach erosion should be properly estimated from historical air photos at littoral cells.

Then, the beach recovery rate in an annual oscillation (t = 12 months) and in an interannual fluctuation (t≈48 months) can be calculated as

$$\chi_{1yr}=\exp\left(-\frac{12}{100}\right)\approx 0.89 \quad (5)$$

and

$$\chi_{4yr}=\exp\left(-\frac{48}{100}\right)\approx 0.62 \quad (6)$$

It means that sediment losses by sea-level fluctuation with the periods of 1 year and 4 years are about 10% and 40%, respectively. If the sea-level rise (SLR) can be applied to the beach slope (β), the net change of shoreline by sea-level rise are calculated as follows:

① shoreline retreat$=SLR\times\beta$ (7)

② shoreline accretion$=SLR\times\beta\times\chi$ (8)

①−②; net retreat$=SLR\times\beta\times(1-\chi)$ (9)

When applying long-term linear trend, interannual and annual fluctuations of sea level for beach recovery rates, the contribution of each component to the shoreline retreat on the Waimea Bay Beach, Oahu and Hapuna Beach, Hawaii was estimated in Table 6. The direct contribution of long-term trend of sea-level rise to the total beach erosion does not seem to be significant on both beaches; 4% on Waimea Bay Beach, Oahu and 8% on Hapuna Beach, Hawaii. But, if global warming is accelerating in the near future, thermal expansion of seawater would contribute more to long-term rise in sea level. The contributions of sea-level changes, not only long-term linear trend but also interannual and annual components, and (short-term) waves to the shoreline retreat were estimated about half and half, respectively, at both beaches. The shoreline retreat by sea-level change only can be formulated by

$$\begin{aligned} R(t)&=R_o(t)+R_1(t)+R_2(t)\\ &=\frac{L}{H}r\Big\{S_o+S_1\Big(1-B_1\exp\Big(-\frac{t_1}{T_c}\Big)\Big)\\ &\qquad+S_2\Big(1-B_2\exp\Big(-\frac{t_2}{T_c}\Big)\Big)\Big\} \end{aligned} \quad (10)$$

where $R(t)$=total shoreline retreat by sea-level changes, $R_o(t)$=retreat by long-term linear rise, $R_1(t)$=retreat due to annual oscillation, $R_2(t)$=retreat due to interannual fluctuation, S_o=vertical sea-level rise, S_1=mean height difference of the annual oscillation, S_2=mean height difference of the interannual fluctuation, B_1=erosion coefficient for the annual oscillation with $t_1=12$, and B_2=erosion coefficient for the interannual fluctuation with a quasi-period t_2

Here it is assumed that there is no longshore transport of sediments and that the beach slope is always the same during eorsion process. The quasi-periodicity and the irregularity of the height difference of the interannual fluctuation make it difficult to predict beach erosion and shoreline retreat. When the mean period of ENSO events is assumed to be 4 years, the shoreline retreats on an isolated beach for different slopes with sea-level changes can be calculated as follows (Table 7):

In addition, potentially accelerating global warming in the near future will definitely cause the long-term effect to contribute more to the beach erosion. The possible worst scenario of the coastal erosion might be the case when big storm surges (or seismic sea waves) attack low-lying coastal areas during the maximum phases of annual and interannual fluctuations with the rising trend of sea level.

Table 7. Calculated shoreline retreats [cm/yr] for different beach-face slopes with typical sea-level changes (S_o=2 cm/10 yrs, S_1=10 cm/yr, S_2=16 cm/4 yr are assumed here)

beach-face slope	$R_o(t)$ trend	$R_1(t)$ annual	$R_2(t)$ interannual	$R(t)$ total
1:5	1.0	5.7	7.6	14.3
1:6	1.2	6.8	9.1	17.1
1:7	1.4	7.9	10.7	20.0
1:8	1.6	9.0	12.2	22.8
1:10	2.0	11.3	15.2	28.5
1:20	4.0	22.6	30.5	57.1
1:30	6.0	33.9	45.7	85.6
1:40	8.0	45.2	61.0	114.2
1:50	10.0	56.5	76.2	142.7
1:75	15.0	84.8	114.3	214.1
1:100	20.0	113.0	152.4	285.4

4. SUMMARY AND CONCLUSIONS

There is little doubt about the fact that the relative sea-level has risen between 10 and 30 cm during the past century. Relative sea-level rise is a combined effect of eustatic sea-level rise and isostatic land-level fall. Potentially accelerating rate of the increase in the atmospheric concentrations of the greenhouse gases is recognized to cause the corresponding increase of atmospheric temperature, which will in turn result in the global mean-sea-level rise by about 60 cm within the next 100 years, according to the IPCC (1990) scenario.

Sea levels never steadily rise or fall but fluctuate with different time-scales in a long term. One of the most prominant feature of sea-level change in the world oceans is an annual cycle, the height difference of which is about 10 cm in the Pacific Ocean, except in some marginal seas. Time series of sea-level records smoothed with twelve-month-running means show an interannual fluctuation with a quasi-period of 2 to 7 years, which are closely associated with ENSO cycle. The height difference of an interannual fluctuation of sea level ranges from about 10 to 20 cm.

Long-term linear trend of sea-level rise ranges from 1 to 5 cm/decade at most of the tide-gauge stations in the Pacific Ocean. Volcanic loading at Kilauea, Hawaii results in land subsidence in the Hawaiian Islands. The Island of Hawaii subsides with fastest rate (+2.6 mm/yr), the neighboring island Maui with a slower rate (+0.8 mm/yr), and the Islands of Oahu (+0.2 mm/yr) and Kauai (+0.4 mm/yr) with slowest rates. The slightly falling trend of sea level at Midway *atoll* is associated with reef evolution and plate movements. Subduction of the oceanic plate underneath the continental plate makes land subsidence and tilting in the Philippine and Japan Islands. The rate of submergence is the sum of the rate of subsidence and the global rate of sea-level rise (+1.4 mm/yr) for the past century.

The annual cycle of sea level and the alternate wave conditions are the primary causes of the cross-shore oscillation of sediment transport, which results in some permanent loss of sediments to deep water. The interannual fluctuation of sea level may be the most significant factor of coastal erosion problems, which seems to contribute about five to ten times more than the long-term linear trend does. As the time scale of the fluctuation becomes longer, the rate of beach erosion would be higher or beach recovery rate would be lower. Based on historical aerial photographs of shoreline changes in the Hawaiian Islands, it is hypothesized that the beach recovery rate exponentially decreases as the period of sea-level fluctuation increases.

In order to correctly estimate the contributions of each component of sea-level signals (linear trend, annual oscillation, and interannual fluctuation) and of waves to the beach erosion in a long-term scale, a new model including those components should be developed. A modified Bruun Rule, based on the hypothesis of the beach recovery rate, is suggested as Eq. (3). The time scale of the chronic beach erosion must be carefully defined and estimated from the data about shoreline changes such as wave climate, aerial photographs, frequency of tropical and subtropical storms, local tectonism, and human impacts.

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