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# Mangrove Ecosystem Collapse During Predicted Sea-Level Rise: Holocene Analogues and Implications

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## ABSTRACT

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Review of the stratigraphic record of mangrove ecosystems during sea-level changes of the Holocene shows that low islands will be particularly vulnerable to the loss of mangrove ecosystems during the rises of relative sea-level projected for the next 50 years. Mangrove ecosystems in these locations could keep up with a sea-level rise of up to 8–9 cm/100 years, but at rates of over 12 cm/100 years could not persist. This is due to low rates of sediment accumulation, with limited sources from outside the mangrove zone, such as from rivers or soil erosion sources. Other factors contributing to mangrove persistence are the primary production rate of forests, shoreline erosion due to deeper and more turbulent water and the frequency and intensity of tropical storms.

**ADDITIONAL INDEX WORDS:** *Mangrove peat, refuge mode, expansive mode.*

## INTRODUCTION

Rise of mean sea-level (MSL) has an immediate and direct effect on ecosystems of the intertidal zone, with decline in influence of terrestrial processes at all locations, and increase in influence of marine processes. It has been envisaged that species with specific tolerances within the tidal spectrum will migrate landward (COMMITTEE ON ENGINEERING IMPLICATIONS OF CHANGES IN RELATIVE MEAN SEA LEVEL, 1987; HEKSTRA, 1989) as their former habitats become increasingly marine. Retreat in this manner of mangrove ecosystems in the Gulf of Papua has been modeled by PERNETTA and OSBORNE (1988). Disruption of established intertidal forests and relocation on formerly more terrestrial surfaces as these become inundated by the rising sea may not be so simple, with alteration of ecological and sedimentological processes. This issue is examined in this review.

Mangroves are a taxonomically diverse group of mainly arboreal angiosperms that grow in the upper intertidal zone of sheltered shores in the tropics. They show high fidelity to particular habitats, because of special physiological

and morphological adaptations to a stressful environment, and this results in zonation of species according to elevation of the substrate. Mangrove species are rarely found outside the intertidal area, and mostly occur between mean sea level (MSL) and the level of mean high water spring tides (LEAR and TURNER, 1977; BELPERIO, 1979). The importance of mangrove ecosystems is well established, as sediment traps promoting aggradation and maintaining the quality of coastal waters, natural breakwaters protecting coastlines from erosion during storms, a natural resource base for silviculture and a large range of economic products, habitats for rare fauna, and nurseries for commercially valuable fish and crustacean species, as well as for educational and tourism uses. These are described in detail by CHAPMAN (1976), CHRISTENSEN (1983), SALM and CLARK (1984), HAMILTON and SNE-DAKER (1984) and TOMLINSON (1986).

Mangrove ecosystems have been termed “mangals” (MACNAE, 1968) to distinguish them from “mangroves,” a term which refers to the specially adapted component species. Usually in context there is no confusion (TOMLINSON, 1986), but the qualifications of “mangrove ecosystem” or “mangrove species” are used here. While mangrove species can grow

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opportunistically for a generation on any tropical shoreline, their permanent establishment over extensive areas is limited to locations of optimal environmental conditions of calm water, a gently sloping sedimentary intertidal area, and a relatively stable sea-level. Extensive mangrove ecosystems occur in such areas as the west coast of Florida, the Ganges and Brahmaputra deltas in India and Bangladesh, the coasts of Sumatra, Borneo and Papua New Guinea, the deltas and estuaries of Queensland and the Northern Territory in Australia, and similar locations in east and west Africa. They are characterized by zonal organization of constituent species, which has often been interpreted as representing successional change over time (DAVIS, 1940; CHAPMAN 1976), by high rates of organic production, and by accumulation of vegetative detritus *in situ* to form a mangrove peat. Mangrove deposits also contain faunal remains and allochthonous inorganic and organic matter brought in by the tides and by rivers, though the magnitude of these components depends on the tidal range and physiographic location.

The accumulation of peat under mangroves in the upper half of the tidal range renders it a useful sea level indicator (SCHOLL, 1963, 1964). For example, near Townsville, Queensland, where MSL is 1.9 m above tidal datum (lowest low water springs) modern mangrove deposits have a maximum range of +1.5 to +3.6 m above tidal datum, and ancient mangrove deposits therefore provide an indication of mean tide level within  $\pm 1$  m (BELPERIO, 1979; GRINDROD and RHODES, 1984). In Tongatapu, Tonga, modern mangroves occur between 0.4 and 0.9 m above MSL, with a mean tidal range of 1.07 m (ELLISON, 1989). Mangrove deposits have been widely used to establish the Holocene sea-level record for in tropical areas, and such studies can be used to indicate how mangrove ecosystems have responded to varying rates of sea-level change in the past and to predict their response in the future. The latter is of immediate importance in assessing the effects of anticipated sea-level rise resulting from global warming.

## MANGROVES AND SEA-LEVEL CHANGE

Eustatic sea-levels rose from ca-120m to near-present levels as Northern Hemisphere

ice sheets melted following the last glaciation (FAIRBANKS, 1989). Maximum sustained rates of rise were at least 1 m/100 years, and this and subsequent decreased rates can be used to interpret the response of mangroves to sea-level change.

Since 6000 BP vertical variation in sea-level has been substantially reduced, but there have been regional and local differences resulting from isostatic readjustment of the earth's crust to the Holocene redistribution of ice and water (CLARK, *et al.*, 1978). DEVOY (1987) integrates concepts and global patterns of sea-level variation. In the Caribbean (Figure 1), sea-levels have been slowly rising since the mid-Holocene, with no higher level than the present (SCHOLL, *et al.*, 1969; PARKINSON, 1989). In Australia and the south Pacific, dating of different indicators of former higher sea-level has accumulated support for a mid-Holocene highstand of up to +2 m in the interval 4000–3000 BP (CHAPPELL, 1983; HOPLEY, 1987). A similar high stand is described from the western Indian Ocean (BATTISTINI, *et al.*, 1976), though it is apparently absent in the south-central Indian Ocean (MONTAGGIONI, 1979). Given the approximately uniform rates of litter production in mature mangrove forests, and in the absence of major allochthonous input, these different regional Holocene sea-level histories (ranging from differences between Florida and Queensland of 5 m at 5000 BP, decreasing to 2 m at 3000 and 1 m at 2000: Figure 1) can be used to explain varying rates of mangrove ecosystem development and accretion and the asynchrony of mangrove dynamics on a variety of timescales (THOM, 1967; STODDART, 1980).

Since the pioneering study of modern tide-gauge records by GUTENBERG (1941), it has also been recognized that sea-level changes are continuing at the present day. BARNETT (1983) showed that these are consistently positive except in areas of isostatic and tectonic uplift. In an analysis of 155 globally distributed though regionally concentrated tide-gauge records for the last 50 years, BARNETT (1984) showed a trend estimate of sea-level rise of  $22.7 \pm 2.3$  cm/100 years. PIRAZZOLI's (1986) analysis of 229 records initiated between 1700 and 1947 AD indicates a modal rise of 10–15 cm/100 years, but with great variability between stations. Using two different methods

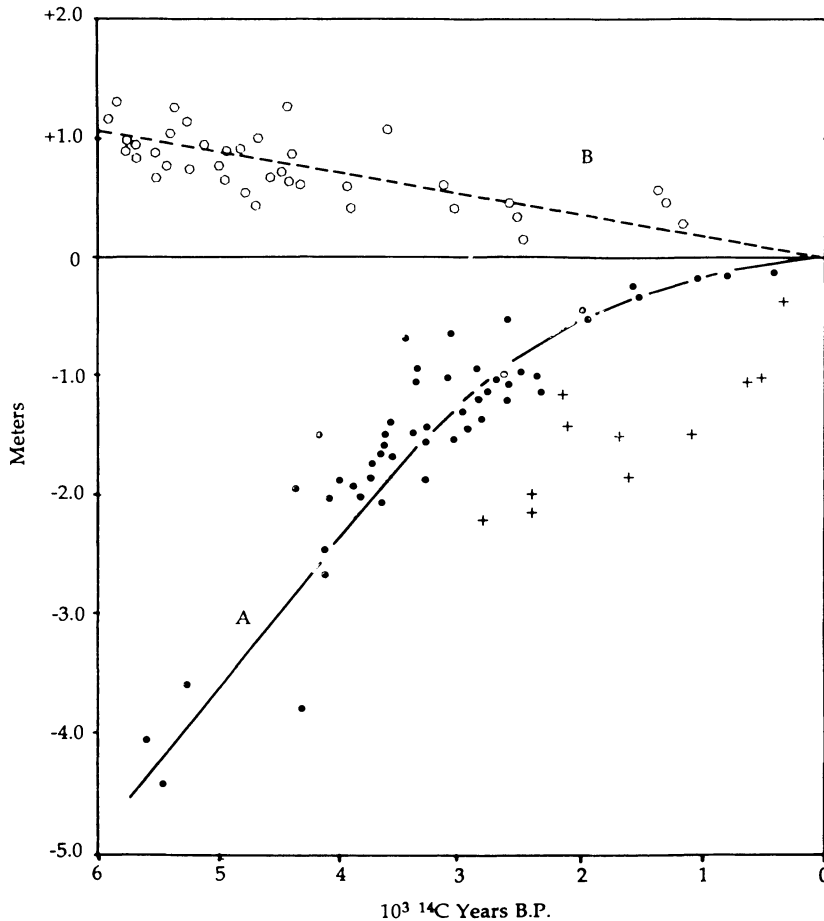


Figure 1. Variation in relative sea level in the late Holocene in (A) Southwest Florida, after Scholl, *et al.* (1969) and (B) North Queensland, after Chappell (1983). In the Florida case, • = samples deposited at or above mean sea level, + = samples deposited at or below mean sea level.

of analysis for 231 stations (excluding Fennoscandia), GORNITZ and LEBEDEFF (1987) find an average rise of 12 and 10 cm/100 years. Most of the stations utilized in these analyses are located in Europe, North America and Japan, however, and thus lie outside mangrove areas. Data for ten stations where mangroves grow are given by PIRAZZOLI (1986); over data sets ranging in length from 40 to 97 years these show a mean rise of 22 cm/100 years. Excluding Bangkok, which is affected by deltaic subsidence, the mean rise is 14 cm/100 years (Table 1). Using slightly different data sets for six of these stations, BARNETT (1984) derives a mean rise of 23.6 cm/100 years. These rates, if

real and persistent, would indicate a sharp reversal of the post-mid-Holocene fall in sea-level of  $-1.7$  cm/100 years identified in Figure 1 for Queensland, and a rapid re-acceleration of sea-level rise in the western Atlantic, to rates equal to or exceeding that of  $+12$  cm/100 years existing in west Florida before 3000 BP.

It is in the context of these Holocene and modern rates of change in sea-level that the consequences for mangrove ecosystems of predicted rises in sea-level resulting from global warming can be evaluated. The WORLD METEOROLOGICAL ORGANIZATION (1986) estimated that increased build-up of carbon dioxide and other greenhouse gases in the atmosphere will,

Table 1. Contemporary rise in sea-level from tide-gauge records.

| Station               | From Pirazzoli (1986) |                 | From Barnett (1984) |                 |
|-----------------------|-----------------------|-----------------|---------------------|-----------------|
|                       | Period of record      | Rate, cm/100 yr | Period of record    | Rate, cm/100 yr |
| <b>Atlantic Ocean</b> |                       |                 |                     |                 |
| Cristobal             | 1909–1969             | +9.0–4.0        | 1909–1969           | +10.97          |
| Bermuda               | 1933–1980             | +28.0–18.0      | 1932–1972           | +23.93          |
| Takoradi              | 1929–1982             | +9.0–39.0       | 1929–1980           | +39.0           |
| <b>Indian Ocean</b>   |                       |                 |                     |                 |
| Calcutta              |                       |                 | 1932–1963           | +34.48          |
| Madras                | 1880–1977             | –5.0 ± 6.0      |                     |                 |
| Vishakhapatam         | 1937–1978             | +8.0 ± 9.0      |                     |                 |
| Rangoon               | 1880–1962             | +18.0 ± 16.0    |                     |                 |
| Andaman Islands       | 1880–1920             | +18.0           |                     |                 |
| Bangkok               | 1940–1981             | +95.0 ± 25.0    |                     |                 |
| <b>Pacific Ocean</b>  |                       |                 |                     |                 |
| Manila                | 1902–1965             | +13.0 ± 8.0     | 1901–1977           | +17.47          |
| Balboa                | 1908–1969             | +18.0 ± 4.0     | 1908–1969           | +15.73          |

among other things, lead to a rise in global mean temperature of 1.5–4.5°C. The magnitude of associated sea-level rise is less predictable, because of both inadequate understanding of the processes of ice-melt and thermal expansion of ocean water and also because of the effects on local relative sea-levels of tectonic and isostatic movements. Projections of future sea-level rise reviewed by the COMMITTEE ON ENGINEERING IMPLICATIONS OF CHANGES IN RELATIVE MEAN SEA LEVEL (1987: 24–28), while uncertain, are summarized in a working expectation of a rise of 0.5–1.5 m by 2100. Widely quoted estimates by Hoffman (1984) over the same time period range from a minimum of 0.5 to a maximum of 3.5 m (*ca* 44–305 cm/100 years) and a probable range of 1.5 to 2.2 m (*ca* 130–190 cm/100 years). In the shorter term, a projected rise of 0.2 m by 2025 was utilized by VICENTE, *et al.* (1988), and 0.17–0.28 m by 2030 by the COMMONWEALTH GROUP OF EXPERTS (1989). These projections give a rate of sea-level rise of 50 cm/100 years for the next half century. The reduced estimate by the American Geophysical Union (MEIER, 1990) of a  $0.34 \pm 0.42$  m rise by 2050 gives a rate of rise of  $57 \pm 76$  cm/100 years.

Modelling of regional sea-level response to ice melting, taking into account structural readjustments, shows considerable variation, but with maximum rates in tropical oceans bounded by mangroves. Table 2 gives estimates from CLARK and PRIMUS (1987) of sea-level rise to the year 2100 for Louisiana, Florida, the

Amazon delta and Bangladesh, based on predictions of melting of the Antarctic ice-sheet (THOMAS, 1985) and the Greenland ice-sheet (BINDSCHADLER, 1985). These are remarkably coherent between localities, ranging from a mean minimum rise of 32.2 cm to a mean maximum of 114.4 cm, equivalent to rates of 26–92 cm/100 years. Interpolation from the results of CLARK and PRIMUS (1987) for other major mangrove areas such as east and west Africa, Indonesia and Melanesia again show coherent rises averaging 110 cm, or 89 cm/100 years.

### STRATIGRAPHIC RECORD OF MANGALS IN THE HOLOCENE

Records of mangal response to sea-level changes in the early Holocene and different regional conditions in the later Holocene are reviewed below, in order to address the question of the reaction of mangrove ecosystems to future accelerating sea-level rise. Emphasis is placed on the stratigraphy of mangal deposits and their rate of accumulation under different rates of change of sea-level.

Detailed work on mangrove stratigraphy of low, limestone islands has been carried out on Grand Cayman, Cayman Islands, and Tongatapu, Tonga. Studies of mangal and deltaic stratigraphy of high Pacific islands have been carried out on Viti Levu (Fiji), Western Samoa and on Pohnpei and Kosrae (Caroline Islands). Mangal stratigraphy on continental margins has been investigated in Florida and northern

Table 2. *Estimates of projected sea-level rise to the year 2100, from Clark and Primus (1987), p. 364 (cm).*

| Area         | Sea-level rise from melting of the Greenland ice-cap, under the assumption of a true eustatic rise of |       | Sea-level rise from melting of the Antarctic ice-cap, under the assumption of a true eustatic rise of |       | Minimum and maximum estimates of sea-level rise from both sources |
|--------------|---|-------|---|-------|---|
|              | 10 cm   | 25 cm | 20 cm   | 80 cm |   |
| Louisiana    | 9.12  | 22.8  | 23.2  | 92.7  | 32.3–115.5  |
| East Florida | 8.33  | 20.8  | 23.4  | 93.4  | 31.4–114.2  |
| Amazon delta | 10.29   | 25.7  | 22.6  | 90.4  | 32.9–116.1  |
| Bangladesh   | 10.05   | 25.1  | 21.7  | 86.7  | 31.8–111.8  |

Table 3. *Critical rates of mangrove peat accretion.*

| Location (source)                  | Rate of accretion (cm/ 100 cal years) |
|------------------------------------|---------------------------------------|
| <b>Low islands</b>                 |                                       |
| Tongatapu (ELLISON, 1989)          | 7.7                                   |
| Grand Cayman (WOODROFFE, 1981)     | 8.8–9.0                               |
| <b>High islands</b>                |                                       |
| <b>Fiji</b>                        |                                       |
| (SOUTHERN, 1986)                   | 11.7                                  |
| (MATSUSHIMA, <i>et al.</i> , 1984) | 7.6                                   |
| (LATHAM, 1979)                     | 13.1                                  |
| <b>Caroline Islands</b>            |                                       |
| (WARD, 1988)                       | 13.4–14.0                             |
| (MATSUMOTO, <i>et al.</i> , 1986)  | 13.7                                  |
| (BLOOM, 1970)                      | 3.0                                   |
| <b>Samoa</b>                       |                                       |
| (MATSUSHIMA, <i>et al.</i> , 1984) | 9.9–10.5                              |
| (BLOOM, 1980)                      | 18.8                                  |

Australia. Data on the reaction of mangroves in these different geomorphological situations is available for the sea-level changes of the Holocene, described below and summarized in Table 3.

Absolute chronology for mangrove stratigraphy is given in this paper by  $^{14}\text{C}$  dating, in radiocarbon years BP (before 1950). Because radiocarbon years have a variable relation over time to sidereal years, the comparison of rates of sea-level rise at different times requires their calibration to sidereal years using the curves of STUIVER & PEARSON (1986), PEARSON and STUIVER (1986), PEARSON *et al.* (1986) and LINICK *et al.* (1986). Corrections from  $\delta^{13}\text{C}$  for fractionation are not made, this being constant over time for mangrove peat hence not affecting rates of change. All rates of accretion or sea-

level change derived from mangal stratigraphy and quoted in this paper are in calibrated radiocarbon years and are hence directly comparable,  $^{14}\text{C}$  ages thus calibrated are denoted in cal years BP.

### South Florida

Extensive mangroves occur in the Ten Thousand Islands and Whitewater Bay areas of the southwest coast of the Peninsula, occupying an area of  $60 \times 25$  km. They have been under the influence of a slowly transgressive sea for the last 4000–5000 years (SCHOLL and STUIVER, 1967; PARKINSON, 1989), with no Holocene sea-level higher than present. Mangroves began colonization of formerly terrestrial surfaces in Whitewater Bay about 3500 BP, and were well established by 3000 BP (SCHOLL, 1964). Analysis of a series of dates from mangrove stratigraphy showed that before 3500 BP sea-level rose at a rate of 9.8 cm/100 cal years (SCHOLL *et al.*, 1969), then slowed to 4.6 cm/100 cal years. Sediment accumulation from biological sources (rooted vegetation, algae and molluscs) and mangrove peat accumulation kept pace with this slower rate (PARKINSON, 1989).

At Rodriguez Bank, southeast Florida, a mangrove woodland of some  $600 \times 300$  m dimensions developed a peat dated at  $5550 \pm 180$  BP, between 3.5 and 4.3 m below present MSL (TURMEL and SWANSON, 1976). This was in a sheltered location as rising sea-level flooded an irregular limestone surface, and on submergence of the entire platform the mangrove swamp ceased to exist. Rate of sea-level rise at this time was 9.8 cm/100 cal years (SCHOLL *et al.*, 1969). In the Ten Thousand Islands area, PARKINSON (1989) finds that mangrove peat production and accumulation

rates were unable to keep pace with rising sea-level at this time.

### Grand Cayman, Cayman Islands

Grand Cayman is a low, limestone island with mangrove swamps extending over 50% of the land area of 197 sq. km. The Holocene stratigraphy of these swamps has been analyzed by WOODROFFE, *et al.* (1980), and WOODROFFE (1981, 1982, 1983). Mangrove-derived peat deposits which underlie most of the mangrove swamps of Grand Cayman have developed during a period of gradual Holocene submergence, since not long before 2000 BP, as shown for Florida. WOODROFFE (1981) has shown how mangroves migrated inland onto formerly terrestrial limestone surfaces during rise of relative sea-level of the Late Holocene. Most cores showed less than 1 m depth of peat, but in places peat up to 4 m thick was found with red woody fragments characteristic of the cortex of root material of *Rhizophora mangle* (WOODROFFE, 1980). Much of the modern mangrove occurs in semi-enclosed embayments with limited marine sediment input and no sediment sources from terrestrial runoff; hence peats show 50–80% organic loss on ignition, demonstrating their autochthonous nature. Radiometric dates on basal mangrove peat samples suggest that sea-level was approximately 185 cm below its present level by at least 2100 BP (WOODROFFE, 1981), 2066 or 2100 cal BP. This gives a rate of peat accumulation from Cayman of 8.8–9.0 cm/100 years.

### Queensland & Northern Territory, Australia

Records of mangrove ecosystems through the Holocene in tropical Australia show that fluvial sedimentation of terrestrially derived material is a factor confusing the reaction of mangroves to sea-level change. Macrotidal ranges of up to 7 m also tend to mask sea-level changes of lesser magnitude in the later Holocene (MCLEAN, *et al.*, 1978; KIDSON, 1982). In north Queensland BELPERIO (1979) and CHAPPELL and GRINDROD (1984) showed shoreline progradation of fluvial/deltaic sediment at rates up to 270 m/100 years, dominating the location and development of mangal. At

Missionary Bay, Queensland, GRINDROD and RHODES (1984) found no unit with sufficient organic material to be called peat in sediment cores from mangrove-occupied estuaries, showing the dominance of these environments by allochthonous sediments.

The largest mangal recorded for the early Holocene is in the Northern Territory, where extensive mangroves flourished between 7000 and 5500 years BP at the time of sea-level stabilization, before fluvial sedimentation at these sites destroyed the habitats of mangroves (WOODROFFE, *et al.*, 1985; WOODROFFE *et al.*, 1987; GRINDROD, 1988; WOODROFFE, *et al.*, 1989). This sequence is well illustrated by the pollen record of mangrove assemblage changes over time in the South Alligator River drillcore 40 (WOODROFFE, *et al.*, 1985). Sea-level conditions were slowly transgressive earlier in the record, with increasing influence of marine processes and corresponding changes in mangrove species present from upper mangrove to sea-fringing types. The pattern then reverses, with marine tolerant species being replaced by landward species, this attributed to sedimentation rather than sea-level fall.

### Fiji

The Holocene mangrove peat record of Fiji is complicated by anthropogenic increases in fluvial sediment yield and hence in deltaic and other coastal sedimentation (SOUTHERN, 1986: p. 215). Consequent accelerated shoreline progradation affects the relationship between mangrove peat formation and the sea-level.

SOUTHERN (1986) found that human-induced factors influenced coastal sedimentation rates on the south coast of Viti Levu more than did sea-level change, as indicated by deltaic accumulation of terrestrial sediments containing carbonized particles. In the deltaic situation it is also sometimes difficult to distinguish freshwater wetland peat from mangrove peat. At Melimeli, Southern found peat to 1.8 m in depth, dated to  $1630 \pm 250$  BP, giving an accumulation rate of c. 11.7 cm/100 years cal. BP. Pollen study of this basal level showed a domination by mangrove taxa such as *Bru-guiera* and *Rhizophora*, but the interpretation was of freshwater peat accumulation in a location partially impounded from marine influ-

ences by discontinuous beach ridges. MATSUSHIMA *et al.* (1984) cored a mangal at Kalokolevu, also on the south coast of Viti Levu, finding silty humus to a depth of 2.25 m. A date of  $2640 \pm 140$  BP at 2.10 m depth in the core gives a subsequent accumulation rate of 7.6 cm/100 cal years. On the same coast at the Rewa delta, BLOOM (1980) dated a peat sample from 4.5 m depth at  $5500 \pm 110$  BP, and attributed stratigraphic records of estuarine muds below fresh-water bog to deltaic progradation.

At Levuka, a restricted swamp on the southwest coast of Lakeba, LATHAM (1979) found organic sediment with 25% organic matter above an organic clay. A date of  $940 \pm 120$  BP at 1.2 m depth gives a subsequent accumulation rate of 13.1 cm/100 cal years. A loose coral sample from beneath the peat gave a date of  $4470 \pm 155$ , taken as evidence for expansion of mangroves over reef flat areas since that time (MCLEAN, 1979).

### Caroline Islands

In 100 cores on the volcanic islands of Truk, Pohnpei (Ponape) and Kosrae (Kusaie) in the Carolines, BLOOM (1970) showed a 6.2 m accumulation of sediments in the last 6500 years. Between 6500 and 4100 years BP these sediments are estuarine muds, accumulating at 25.3 cm/100 cal years. Material dated to 4100 BP then shows widespread conversion of open sandy bays at W. Moen, Truk, to fresh/brackish water swamps, and the subsequent accumulation of peat at 3.0 cm/100 cal years. This peat is formed by *Phragmites* and the mangrove palm *Nypa* as well as by mangrove species *sensu stricto*. At Sapuk, (northeast Moen, Truk), mangroves have extended over drowned slopes, backed landward by freshwater swamps in former estuaries. Typically 1–2 m of red-brown, woody, tough mangrove peat has accumulated above a level bench of carbonate rock. At Neiwe (S.E. Moen, Truk) MATSUMOTO, *et al.* (1986) cored to find peat to 2 m depth, and a date of  $3980 \pm 170$  BP at the base of the peat unit interpreted as the time of conversion of shallow sand flats to reed swamp, with slowing of rates of sea-level rise. On Tol, in the Truk group, 3 m of red-brown mangrove peat has accumulated above shelly estuarine mud (BLOOM, 1970).

At Rohi, S.E. Pohnpei, MATSUMOTO, *et al.* (1986) found mangrove peat to 4 m depth, where a date of  $4790 \pm 220$  BP showed time of conversion of low tide reef flats to mangrove swamps when the rate of sea-level rise was slowing. Mangrove peat accumulation then kept up with rising sea-level, at a maximum rate of 13.7 cm/100 cal years. The same pattern is shown from Kosrae in the Eastern Carolines by WARD (1988), who found rapid accumulation of peat-dominated sediment in the last 4000 years. Figure 2 is based on stratigraphic records given in that paper. At Yewal, peat was found between depths of 5.7 and 2.6 m, giving dates respectively of  $3690 \pm 100$  and  $2280 \pm 130$  BP, an accumulation rate of 14.8 cm/100 cal years. At Yela, peat was found between 5.8 and 3.0 m, giving dates respectively of  $3500 \pm 100$  and  $1820 \pm 100$  BP, and an accumulation rate of 13.4–14.0 cm/100 cal years. This rapid sedimentation rate is attributed to human disturbance and soil erosion inland.

There is little evidence of mangrove swamp development in the earlier Holocene, before 5000 BP. Near Kolonia on Pohnpei, a trace of red woody mangrove peat was found at 4.9 m depth in estuarine mud. Abrupt recent expansion of mangroves on the islands has been attributed to recent introduction, but from the stratigraphic evidence BLOOM (1970) concluded that mangroves were present in earlier refuges in the islands and their expansion delayed by excessively rapid submergence and lack of suitable intertidal locations for colonization.

### Western Samoa

BLOOM (1980) reported dates on peaty mud samples from mangrove swamp cores on the south coast of Upolu Island, Western Samoa. Samples from 5.8–6.1 m and 5.2–5.5 m depth gave radiocarbon dates of  $4845 \pm 95$  and  $4655 \pm 95$  BP, respectively; and samples from 3.6–4.3 m and 2.7–3.0 m depth gave dates of  $3060 \pm 95$  and  $1595 \pm 85$  BP, respectively. For the last 1500 years, mangrove sediments have been accumulating at a rate of 18.8 cm/100 cal years. Little detail of the stratigraphy is given, but results are interpreted as evidence of tectonic submergence of the island during the later Holocene at a rate of 6.6–6.8 cm/100 cal years.



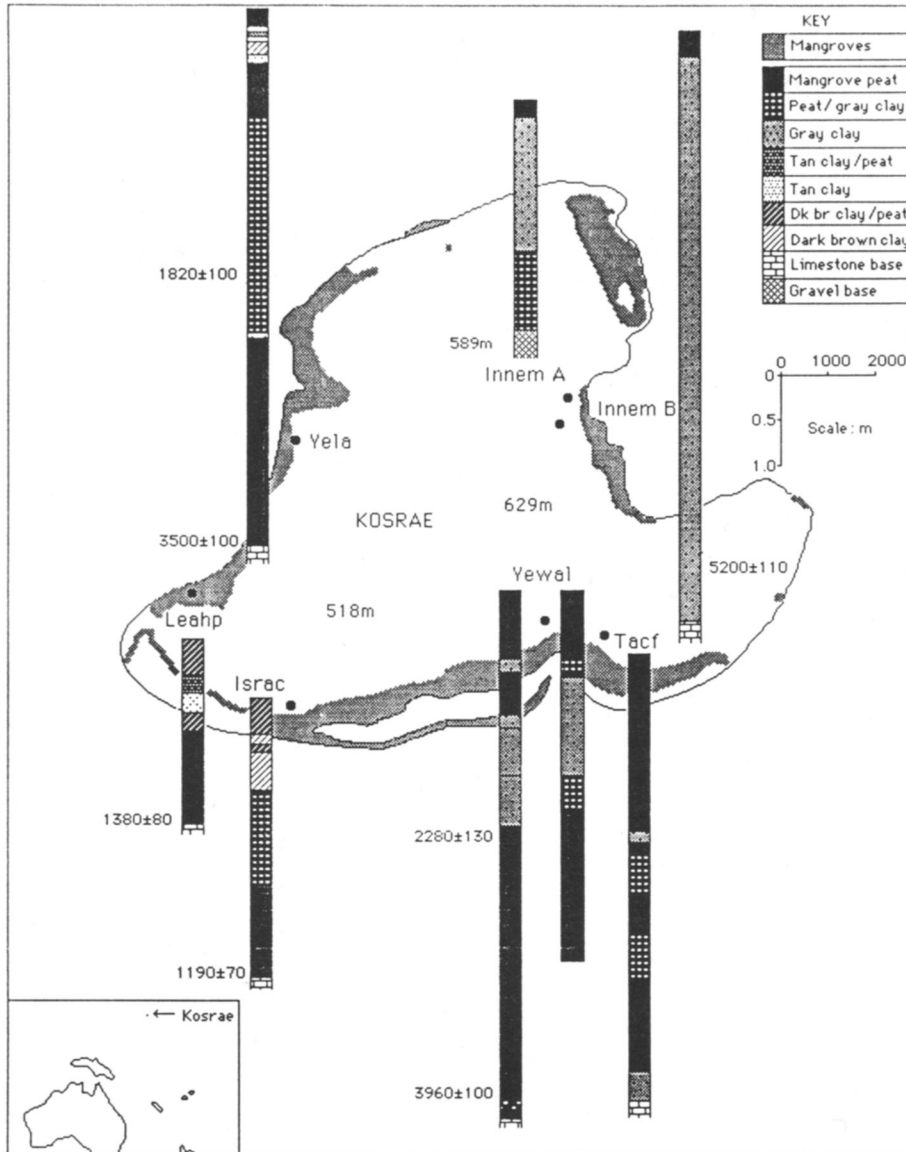


Figure 2. Coastal stratigraphy of Kosrae, Caroline Islands, data from Ward (1988).

MATSUSHIMA *et al.* (1984) cored a mangrove swamp at Lefaga Bay on the south coast of Upolu Island to find silty peat to 1.5 m depth, below which was a shelly silt that gave a date of  $1580 \pm 160$  BP (SUGIMURA *et al.*, 1988). This gives a rate of accumulation of 9.9–10.5 cm/100 cal years.

**Tongatapu, Tonga**

Tongatapu is a low, flat limestone island, possessing mangrove fringes on its sheltered, leeward northern coast, and around the central Fanga 'Uta lagoon. Stratigraphy of these mangrove areas showed sediment depths of 1–4 m (Figure 3, based on ELLISON, 1988).

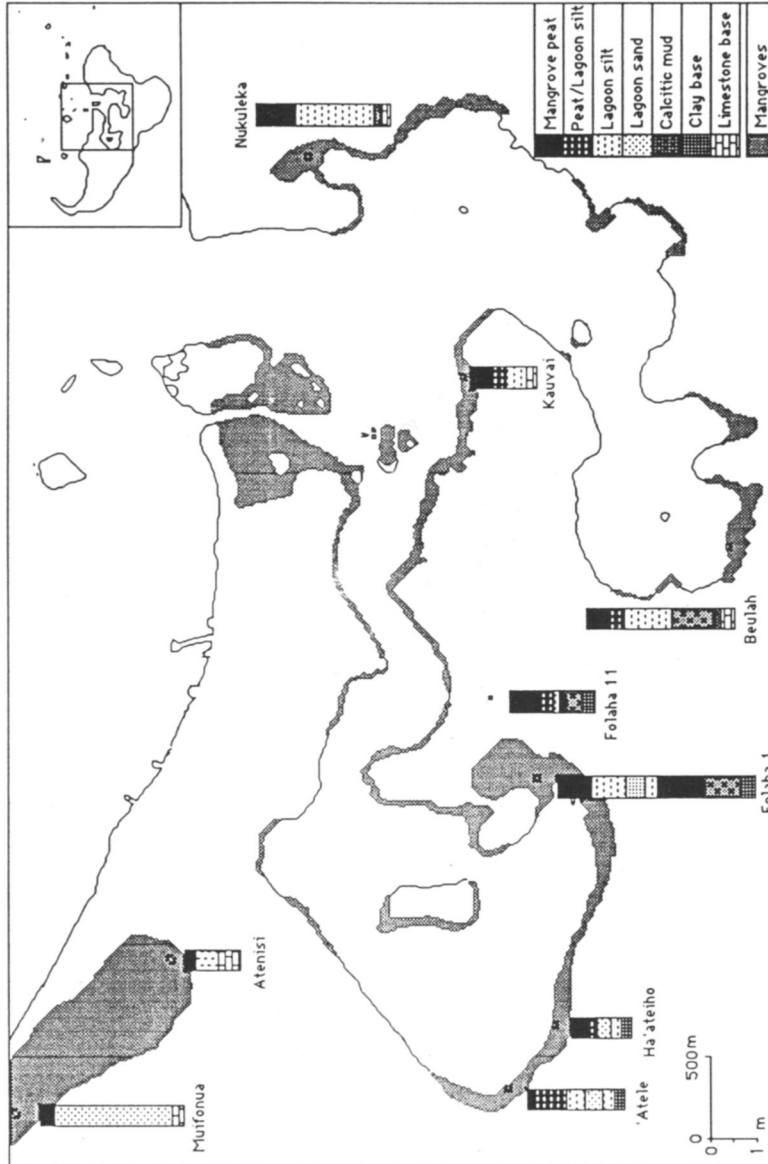


Figure 3. Coastal stratigraphy of central Tongatapu, Tonga.

At Folaha, the largest mangrove area in the western arm of the Fanga 'Uta lagoon, a red, woody mangrove peat was discovered between depths of 1.3–2.3 m below present MSL (ELLISON, 1989; ELLISON, in press). This peat unit underlies all of a 1 km transect across the modern swamp, and continues at lower levels beneath the lagoon. The peat in Folaha core 1

formed between  $6870 \pm 90$  and  $5650 \pm 80$  years BP, giving a rate of peat accumulation in this period of 7.7 cm/100 cal years. Mangrove assemblages contributing to formation of peat at the levels dated have been identified by pollen analysis, and the present elevation of these assemblages has been used to determine past MSL. This shows that MSL was 3.2 m below

present at  $6870 \pm 90$  BP, and 1.7 m below present MSL  $5650 \pm 80$  years BP (ELLISON, 1989). The rate of sea-level rise through the lower peat unit was 12.2 cm/100 cal years.

During this period of slowly rising sea-level, mangrove ecosystems were able to persist by raising substrate levels at a roughly equivalent rate by peat formation. They are drowned out by more rapid sea-level rise, shown stratigraphically by transition from mangrove peat to overlying lagoonal sediment. As shown in Figure 3, no equivalent to this Folaha peat was found elsewhere on Tongatapu, except for peat fragments at 2.6 m depth in the Beulah core in the eastern branch of the lagoon. The Folaha location is very sheltered, and lack of corresponding peat units elsewhere on Tongatapu or other Pacific islands indicate that this rate of sea-level rise is probably the upper limit of rate of sea-level rise with which mangroves can keep up, without significant allochthonous sediment sources.

Deposition of lagoon sediment over areas now occupied by mangroves, as shown in Figure 3, occurred during the regional mid-Holocene sea-level highstand. Upper peat units of all cores show that later sea-level fall allowed mangroves to colonize edges of the lagoon. At the Folaha swamp, pollen analysis of these modern peats between systematic cores showed a sudden change of mangrove species zonation suggestive of a recent 40 cm tectonic raising of the island (ELLISON, 1989), attributed to an earthquake in 1853. This enables detailed peat accumulation rates to be obtained: under higher swamp species such as *Excoecaria* and *Bruguiera*, these average 26 cm/100 cal years, while under the lagoon-fringing *Rhizophora*, they are as high as 53 cm/100 cal years. These rates are higher than those shown associated with stable or rising sea-levels because erosion rates are higher during sea-level rise than during sea-level fall (BRUUN, 1962; SCHWARTZ, 1967).

In more exposed intertidal locations of Tongatapu, such as the north coast and on the offshore islands of Onevai and Pangaimotu, mangroves exist in patches (STODDART, 1975; ELLISON, 1990), but are dominated by marine processes such that they cannot extend to establish zoned forests or accumulate peat. These are the modern analogues of mangroves in refuge mode in conditions of rapid sea-level change, as

envisaged by BLOOM (1970) for the earlier Holocene in the Caroline Islands.

### INFERENCES FROM THE HOLOCENE RECORD

These records of mangrove ecosystems in the Holocene as derived from peat stratigraphy, show that in the earlier Holocene large mangrove swamps did not exist. As sea-level stabilized at various dates in different regions after ca 6500 BP then the first extensive mangrove ecosystems established in sheltered locations.

This is illustrated by the stratigraphic record from the Fanga 'Uta lagoon, Tongatapu, where a large mangrove forest occurred in the middle Holocene only at Folaha in the sheltered western arm of the lagoon. Peat fragments were found at corresponding depths in other sheltered lagoonal areas, but mangroves show no record from this time at other locations where they now occur. A similar situation was described by BLOOM (1970) from Pohnpei. A comparable deposit to that at Folaha is the lower peat at Rodriguez Bank, Florida (TURMEL and SWANSON, 1976). This developed on an irregular limestone surface undergoing marine transgression and ceased to exist when the whole platform was flooded.

Since mangrove peat is not recorded from the period of rapid sea-level rise in the earlier Holocene, we conclude that development of extensive mangal is prohibited in such conditions. Mangroves existed instead in refuge mode, as disorganized, scattered individuals or narrow coastal fringes, with dominant flushing of detritus from root systems. Change in coastal environmental conditions and shoreline location causes successive generations of mangroves to grow in different positions. In some localities, too, lifetimes may be short because of storm damage as the response of biogenic coastal systems lags behind sea-level stabilization (HOPLEY, 1984; NEUMANN and MACINTYRE, 1985). Mangroves are able to persist through habitat migration in refuge mode, but are not able to establish persistent and expansive communities until the sea-level has stabilized.

The response of mangroves to sea-level rise is most clearly shown from the stratigraphic record on low flat islands such as Grand Cayman and Tongatapu where the complicating

factor of terrestrially derived coastal sedimentation is absent. From both locations, peat is shown to accumulate at a rate of 8–9 cm/100 cal years. This indicates the ability of mangrove ecosystems to raise their substrate to keep up with a rising sea-level, in situations where there are no other sources of coastal sediment. On Tongatapu, where modern mangroves occur within a 0.5 m elevational range, the mangrove forest which existed in the Fanga 'Uta lagoon between 7000 and 5500 years BP only partially kept up with a sea level rising at 12 cm/100 cal years, shown by accumulation of a peat of 1 m thick. Increasingly salt tolerant species succeeded over time, until the most specialized mangrove, *Rhizophora*, was drowned out.

At other locations, the mid to late Holocene mangrove peat and sea-level record is obscured by shoreline changes resulting from abundant supply of terrestrially-derived sediment. In locations such as deltas (THOM, 1967) and chenier plains (CHAPPELL and GRINDROD, 1984; GRINDROD and RHODES, 1984), coastal development is controlled primarily by such allochthonous sediment input and mangroves are patterned opportunistically in response to it. On many Pacific islands sediment supply has been accelerated by anthropogenic soil erosion inland. This has influenced the deltaic development and coastal progradation described from Viti Levu and Lakeba, Fiji, and Kosrae in the Caroline Islands. This also leads to opportunistic colonization by mangroves, and makes possible the rapid sediment accumulation rates recorded above. These cases indicate that mangals of high islands and continental coasts can be so dominated by input of terrestrial sediment that the effects of sea-level rise are lessened. Because of the allochthonous component in these sediments, mangrove substrates are accreting at a faster rate than the peats of low limestone islands.

There are certainly problems in the direct inference from the Holocene record of mangal response to future sea-level rise. In particular the major part of the post-glacial rise of relative sea-level was on steep coastlines, especially on reef coasts, until near-present sea-levels were reached. Future sea-level rise, especially in mangrove areas, will occur on topographically low coasts. In such circumstances it may be difficult for mangroves to maintain themselves because of the width of the zone flooded and the

rate of horizontal inundation, and the ability of the mangal to relocate will be too slow both in terms of generation time and accretion rate for communities to maintain themselves. Thus the difference between past Holocene and immediate future conditions does not invalidate the argument presented here.

By analogy with the 'Holocene energy window,' we may also envisage a 21<sup>st</sup> Century energy window as coasts lag in equilibrating to changing sea-level and energy conditions. Further, modelling studies suggest an increase in the destructive energy of tropical hurricanes by 40–60% (EMANUEL, 1987), and their increased frequency (SHAPIRO, 1988), as a result of increasing sea-surface temperatures. This, and the concomitant increase in areas affected by such storms, would increase the already severe stresses on mangals generated by sea-level rise itself.

## DISCUSSION

We conclude from this review that in the absence of significant allochthonous sediment input, mangrove ecosystems appear to be able to keep pace with a rising sea-level of 8–9 cm/100 cal years, are under stress at rates between 9 and 12 cm/100 cal years, and cannot persist in their expansive mode at rates above this. The predicted possible rates of greenhouse-induced sea-level rise of 100–200 cm/100 years make it inevitable that most mangals will collapse as viable coastal ecosystems. This implies that mangrove ecosystems of low islands will be more vulnerable to rising sea-level than those of high islands and continental shores.

However, the environmental diversity of mangals (THOM, 1982; WOODROFFE, 1987) suggests that various factors in addition to sea-level rise contribute towards their ability to maintain extent, location and zonal organization during sea-level rise. In assessing mangal response to the scenario of predicted sea-level rise in ensuing decades we need to consider the sensitivity of the thresholds which govern the transition from expansive to refuge mode in mangals, in different environmental conditions.

The amount of allochthonous sediment available will be a major factor in determining the persistence of mangals during relative sea-level rise. It varies widely along a spectrum

from sediment deficit (coast erosion: SEMEN-IUK, 1980), through sediment absence (in the case of the low limestone islands described here), to various levels of domination by allochthonous sediment input, reaching the extreme case where massive sediment accretion may bury pneumatophores and kill mangroves (STODDART and PETHICK, 1984). This spectrum can be readily related to the physiographic settings of mangals described by THOM (1982) and WOODROFFE (1987). The scenario of ecosystem collapse described in the present paper, and the threshold rates of sea-level rise associated with it, are primarily derived from the neutral case of absence on limestone islands of any allochthonous sedimentation, and the dependence of mangrove substrate response therefore on autochthonous organic accumulation.

This latter is controlled by rates of primary production. LUGO and SNEDAKER (1974), CLOUGH and ATTIWILL (1982) and WOODROFFE (1987) have shown that while many mangals have a litter production of 700–800 g dry weight/m<sup>2</sup>/yr, some may yield up to twice this figure, with an evident inverse correlation between yield and latitude. While the temperature-dependence this implies might suggest a marginal increase in litter production as global warming proceeds, these figures represent input to the forest floor and not rates of accumulation. BOTO and BUNT (1981) in Queensland report the export of 45–70% of litter as a result of tidal flushing, and this figure appears characteristic of mangals where tidal flows are not physiographically inhibited (ODUM and HEALD, 1975). With predicted rates of sea-level rise this rate of tidal flushing will necessarily increase and rates of organic accretion hence decrease, thus increasing the inability of mangals to maintain themselves within tidal limits. It has also long been known that sea-level rise in itself, by allowing higher-energy waves access to the shoreline, will lead to increased mobilization of sediment and accelerated coastal erosion and retreat (BRUUN, 1962; SCHWARTZ, 1967). We therefore conclude that modern mangals will be at least as susceptible to the effects of sea-level rise as those of the Holocene, notwithstanding the differences in conditions between these two periods.

The economic, political and management

implications of the collapse of mangals over the next several decades if the sea-level predictions are correct are considerable, not least because traditional methods of coastal protection (such as construction of sea-walls and similar protection works) are inimical to the ecological requirements of mangroves. Controls of timber exploitation and other potentially disruptive uses can, under the circumstances we have described, have only a marginal effect in offsetting the consequences of rising sea-level. Mangals occur on some one-quarter of tropical coastlines, where their significance in the lives of human populations associated with them, their function in coastal sediment stabilization, as well as their intrinsic scientific importance as ecosystems and recorders of biotic diversity, indicates that particular attention be paid to their survival.

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□ RÉSUMÉ □

Les modifications stratigraphiques des écosystèmes à mangrove intervenus par suite des modifications holocènes du niveau de la mer montrent que les îles basses seront très vulnérables aux pertes de ces types d'écosystèmes qui interviendront au cours de la montée relative du niveau de la mer prévue pour les 50 prochaines années. Les écosystèmes à mangrove peuvent se maintenir avec une montée atteignant 8–9 cm/100 ans, mais à des taux de 12 cm/100 ans, ils ne peuvent persister. Cela est dû à la faible accumulation sédimentaire, au peu de sources d'approvisionnement en dehors des zones de mangrove (rivières, érosion). D'autres facteurs contribuent au maintien des mangroves: la productivité primaire des forêts; l'érosion littorale due à des eaux plus profondes et turbulentes; la fréquence et l'intensité des cyclones tropicaux.—*Catherine Bressolier-Bousquet, Géomorphologie EPHE, Montrouge, France.*

□ ZUSAMMENFASSUNG □

Ein Überblick über die stratigraphische Aufzeichnung von Mangroven-Ökosystemen während Meeresspiegelschwankungen im Holozän zeigt, daß voraussichtlich in den nächsten 50 Jahren besonders flachere Inseln für den Verlust von Mangroven-Ökosystemen während des Anstiegs des relativen Meeresspiegels anfällig werden. Mangroven-Ökosysteme an diesen Standorten können mit einem Meeresspiegelanstieg bis zu 8–9 cm/100 Jahre Schritt halten, aber bei Anstiegsraten über 12 cm/100 Jahre können sie nicht fortbestehen. Dieses ist bedingt durch die geringe Geschwindigkeit der Sedimentakkumulation, aufgrund begrenzter Quellen außerhalb der Mangrovenzone, wie a.B. von Flüssen oder Bodenerosionsquellen. Andere Faktoren, die zum Fortbestehen der Mangroven beitragen, sind die Primärproduktionsraten der Wälder, Küstenlinienerosion bedingt durch tiefere und turbulente Gewässer und die Häufigkeit und Intensität von tropischen Stürmen.—*Gabriele Lischevski, Essen, FRG.*