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ENVIRONMENTAL VARIABILITY AND ENVIRONMENTAL EXTREMES AS FACTORS IN THE ISLAND ECOSYSTEM

BY

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CONTENTS

LIST OF FIGURES

LIST OF TABLES

ENVIRONMENTAL VARIABILITY AND ENVIRONMENTAL EXTREMES AS FACTORS IN THE ISLAND ECOSYSTEM

BY

D. R. STODDART[1] AND R. P. D. WALSH[2]

ABSTRACT

This paper discusses the major environmental hazards affecting island ecosystems, with varying magnitudes and on differing spatial and temporal scales. Vulcanicity and earthquakes are spatially concentrated at present, but can be devastating, especially in continental-marginal areas. Island area and elevation are subject to changes in sea level, notably on Pleistocene time scales but continuing to the present. Tsunamis are episodic sea-level disturbances which may have catastrophic effects. Of climate factors variability in rainfall has greatest ecological consequences: we examine temporal variation in annual rainfall on islands, rainfall seasonality and its variability, the magnitude and frequency of daily rainfalls, and the magnitude and frequency of droughts, using data for tropical island stations. Some perturbations are linked to the El Niño phenomenon and many show regional and temporal coherence. The most extreme climatic disturbances experienced on islands are hurricanes, which vary in frequency both spatially and temporally. This inherent environmental variability and associated extreme conditions has major consequences for the establishment and survival of plants and animals on islands, especially through the control of vegetation growth by rainfall. Ultimately these factors set thresholds for the habitability of islands by man.

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INTRODUCTION

Thirty years ago, in 1961, F. R. Fosberg convened a remarkable symposium on *Man's Place in the Island Ecosystem* at the 10th Pacific Science Congress in Honolulu. At the end of the meeting, O. H. K. Spate (1963) pointed out that much of the discussion had been concerned with scale and with change. During the Symposium the variety of physical environments in the Pacific was comprehensively outlined by William L. Thomas, Jr. (1963). His treatment of the distribution and nature of these environments still stands. In this paper, however, we carry the analysis a stage further by concentrating on spatial and temporal scales of change, and we demonstrate how, far from island environments being merely static backdrops for human activities, they are themselves continuously changing, not only on the scale of recent geological time but also on scales measured in years and decades. Our discussion concentrates on sea-level change and rainfall variability, since we judge these to be of central significance in tropical small-island ecosystems. Nunn (1990, 1991) has recently addressed issues of environmental change on Pacific islands, but since he deals with rainfall only in the context of heavy storms associated with hurricanes our treatments do not overlap. Other authors have recognised the importance of extreme or infrequent events in understanding island ecologies (Waddell 1973, 35; Lea 1973, 56; Whitmore 1974), a subject also judged of importance by McLean (1980), and we consider these phenomena as well as longer-term secular changes. For practical purposes we are concerned mainly with tropical islands, but we take our data from the Caribbean and the Indian Ocean as well as from the Pacific.

VOLCANIC ACTIVITY AND EARTHQUAKES

Many oceanic islands are of volcanic origin, and many coral atolls rest on volcanic foundations. It is not surprising therefore that many islands continue to be affected by continuing volcanicity and seismicity. Such phenomena are strongly concentrated at plate margins, and may have increased in frequency during the Quaternary (Kennett and Thunell 1975). There is also some evidence of temporal periodicities in Pacific volcanism over the past century (McLean 1980, 153).

A basic distinction can be made between the basaltic shield volcanoes of oceanic provinces (such as Hawaii, Tahiti and the Galapagos) and andesitic volcanoes of island arcs and continental borderlands, the latter built primarily by low-angle lava flows and the latter characterised by explosive release of pyroclastic materials. Island volcanoes initiate their growth by submarine activity and finally reach the surface. Of ten active volcanoes in western Tonga, for example, seven are still submarine. The ephemeral creation of new subaerial islands by the other three has frequently been observed, but since they are built of pyroclastics they are easily eroded. Falcon Island (Funuafo'o) was 100 m high in 1885, only 50 m in 1889, down to 13 m in 1895, and had disappeared by 1898. In 1900 it was 3 m high. It appeared again in October 1927 and by spring 1928 was nearly 5 km in diameter and 100 m high; it had disappeared by 1949 (Lister 1890; Hoffmeister et al. 1929). There is a similar history for Metis Shoal (Melson et al. 1970). With continuing activity, however, land may ultimately become more permanent and then become colonised by plants and animals. Surtsey, which was formed on the Mid-Atlantic Ridge south of Iceland in 1963, has been continuously monitored since then, though its biota in such high latitudes is small (Fridriksson 1975).

Temporal and spatial periodicity in volcanicity allows the processes of colonisation, succession and extinction to be studied. Most work has been carried out on the dated individual lava flows of Hawaii (Macdonald et al. 1983), both in terrestrial habitats (e.g. MacCaughey 1917; Skottsberg 1941) and in marine, where the development of coral communities has been studied on a series of flows up to 102 years old (Grigg and Maragos 1974). Such successions can rapidly be terminated by continuing volcanic activity, most dramatically in andesitic areas. The most famous such case is the explosive destruction of Krakatau in August 1883, in an explosion heard up to 4653 km away (Self and Rampino 1981; Simkin and Fiske 1983). In the Lesser Antilles Soufrière on St Vincent exploded on 7 May 1902, followed the next day by Pelée on Martinique, 145 km to the north (Anderson and Flett 1902; Anderson 1908). Ash and mud on the latter generated a massive nuée ardente which destroyed the town of St Pierre. Subsequent activity continues to be monitored, as are the ecological consequences of such events (e.g. Howard 1962).

These can be of ocean-wide extent. Sachet (1955) documented the dispersion of pumice from the Krakatau explosion across the Indian Ocean, where on some islands it formed new beach ridges several meters in thickness. Such non-carbonate inputs are of considerable ecological interest on atolls. Likewise Richards (1958) tracked pumice from the explosion of Barcena in the Islas Revillagigedo in 1952 across the Pacific (Richards and Dietz 1966).

Even comparatively small volcanic events may have considerable ecological consequences. Dickson (1965) points out that in the case of the 1962 eruption on Tristan da Cunha, new lava flows covered 8 ha and ash fields 4 ha, but the vegetated area affected by toxic gases was 32 sq km or one quarter of the area of the island. Tristan well illustrates a conclusion drawn by Brattstrom (1963, 522) following the Barcena eruption: 'The major effect of the catastrophe may not be the event itself. Even though the event may be spectacular and destructive, the side effects or after-effects may be more critical to the survival of both individuals and species ... and to reinvasion and repopulation.' On Tristan da Cunha the explosion occurred in the area most affected by human activities. The population was evacuated from the island, leaving behind uncontrolled cattle, sheep, donkeys, geese, fowls, dogs and

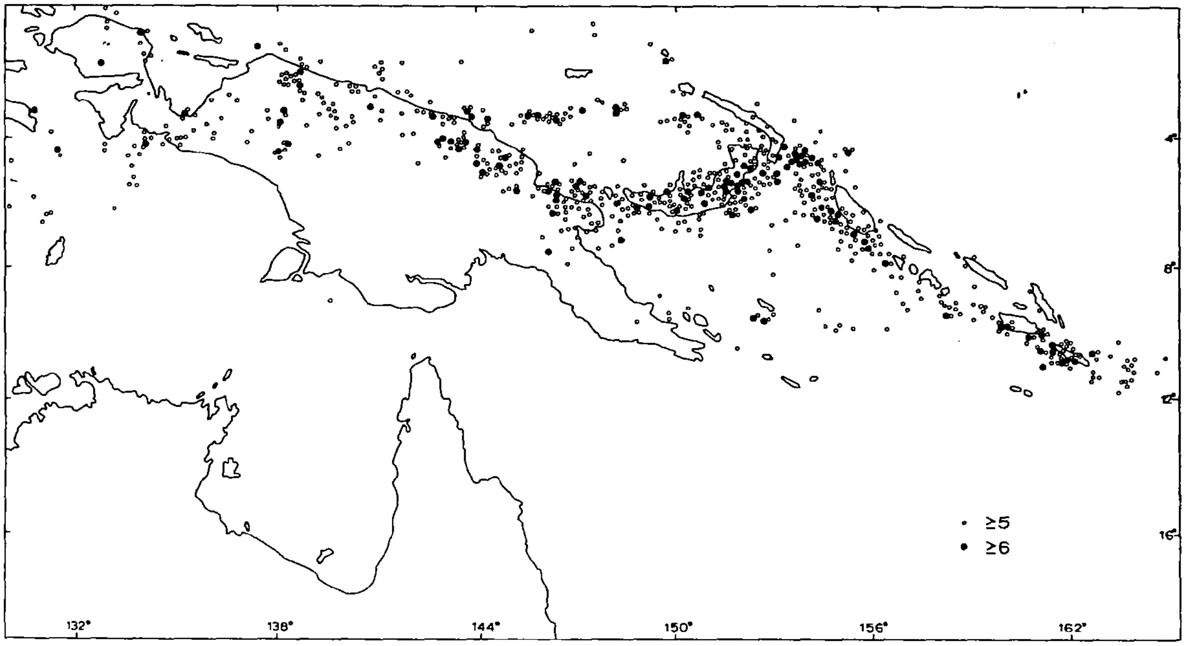

Figure 1. Distribution of earthquakes greater than M = 5 in the New Guinea-Solomon Islands area, 1958-1966 (From Denham 1969).

cats, which perpetuated and transformed the effects of the eruption itself (Dickson 1965; Baird 1965).

Seismically-generated topographic changes may have substantial impact on coastal geomorphology and ecology of islands. Thus at Punta Espinosa on Fernandina Island in the Galapagos the emersion of barnacles and corals indicates upheaval of 0.3-0.5 m between September 1974 and January 1975, and at Urvina Bay on Isabela emersed massive corals indicate uplift of ca 5 m during a few months in 1953-1954 (Glynn and Wellington 1983, 33-36, 121-122).

On the highly seismic islands of Vanuatu [New Hebrides] Taylor and his colleagues have researched episodic reef emergence for over a decade using in part the evidence of microatolls bevelled to former sea-level positions (Scoffin and Stoddart 1978). They have found emergence on North Santo of 1.2 m in 1866 and 0.6 m in 1973; on South Santo of 0.29 m in 1946 and 0.26 m in 1965; on North Malekula of 1.23 m in 1729 and 1.05 m in 1965; and on South Malekula of 0.35 m during 1957-1958 (Taylor et al. 1981, 1987, 1990). Similar upheavals have been documented elsewhere in Melanesia, for example on New Britain and Guadalcanal (Everingham 1974); Figure 1 shows the concentration of earthquakes greater than M = 5 during the short time period 1958-1966. As in the case of volcanic eruptions the effects of earthquakes may be delayed or indirect. Thus the Madang, New Guinea, earthquake of November 1970 (M = 7) had a devastating immediate destructive effect on shallow-water corals in the neighbouring lagoon, but was followed by a lagged sediment input to coastal areas as inland rivers eroded massive landslide deposits (Stoddart 1972; Pain and Bowler 1973).

Over longer time spans episodic tectonic uplift interacting with Quaternary sea-level fluctuations has transformed island topographies in island-arc localities. The suites of uplifted coral reef terraces of eastern New Guinea are perhaps best known (Chappell 1974; Bloom et al. 1974), but similar cases are widespread and illustrate the ephemerality of topographic forms in such situations. The raised terraces of Timor and Atauro have been described by Chappell and Veeh (1978), while Pirazzoli et al. (1991) have identified reef terraces up to over 450 m high and 1 million years old on Sumba in Indonesia. Elevated reefs do exist in intra-plate situations, but there they originate through different mechanisms (McNutt and Menard 1978).

SEA-LEVEL CHANGES

Many of the islands on which volcanic or seismic events have been documented are comparatively large. Fosberg in 1961 pointed out the vulnerability of smaller islands to the effects of external environmental events. Coral atolls and indeed even many oceanic volcanic islands are small. The larger atolls have total areas of about 1000 sq km, and the largest ones (such as Kwajalein in the Marshalls) reach 2500-3000 sq km. But these are total areas bounded by peripheral reefs. The land area at high tide is much smaller. Table 1 gives the total areas and dry-land areas of a number of Pacific and Indian Ocean atolls, and the ratio of one to the other. Clearly quite small shifts in the sea-air interface will have major effects on the area of dry land. Not only will land areas on reefs vary greatly with even quite small changes in sea-level, but many now submerged banks would become land during small (ca 100 m) negative shifts in sea-level such as those associated with the last major continental glaciation. Thus the Bahama Banks at that time formed a flattopped land area totalling 155,000 sq km and the Pedro Bank south of Jamaica 7900 sq km. In the Indian Ocean the Seychelles Bank reached 43,000 sq km, Saya de Malha 40,000, Nazareth Bank 26,000, and the Chagos Banks 13,500 sq km. The southwest Indian Ocean was converted from a largely empty sea to an archipelago of large islands which could serve as stepping-stones for trans- and inter-oceanic dispersing species (Figure 2). Conversely a positive movement of sea-level, of only a few meters, would at least temporarily remove dry-land areas from most coral reefs, with concomitant general extinction of island biotas. The effects of similar sea-level changes on steeper high islands would be less spectacular, except that in continental marginal areas falls in sea-level might unite islands to form larger land masses or indeed terminate their condition of insularity altogether .

Table 1. Areas of atolls

	A Total area sq km	B Land area sq km	A/B
Marshalls:			
Taongi	107.0	3.8	28.2
Bikar	56.5	0.5	113.0
Eniwetok	1023.9	6.4	160.0
Bikini	691.5	7.3	94.7
Rongelap	1104.5	6.4	172.6
Ailinginae	152.2	3.3	46.1
Rongerik	182.2	2.1	86.8
Utirik	92.4	2.7	34.2
Taka	133.9	0.5	267.8
Ujelang	94.1	1.6	58.8
Ujae	216.3	1.6	135.2
Wotho	118.7	4.1	29.0
Lae	26.1	1.6	16.3
Kwajalein	2335.5	14.4	162.2
Likiep	466.4	9.4	49.6
Jemo	3.8	0.2	19.0
Ailuk	232.1	5.7	40.7
Wotje	773.5	8.7	88.9
Erikub	302.1	0.9	335.7
Mejit	9.1	3.4	2.7
Maloelap	1004.9	9.9	101.5
Other Pacific atolls:			
Kapingamarangi	61.6	1.1	56.0
Raroia	398.9	20.7	19.3
Rangiroa	1639.5	43.0	38.1
Indian Ocean atolls:			
Addu	131.8	11.2	11.8
Diego Garcia	170.0	30.0	5.7
Salomon	36.0	3.1	11.6
Peros Banhos	510.0	10.6	48.1
Farquhar	170.0	7.5	22.7
Cosmoledo	152.0	5.2	29.2
Aldabra	365.0	155.0	2.4
Astove	9.5	4.2	2.2

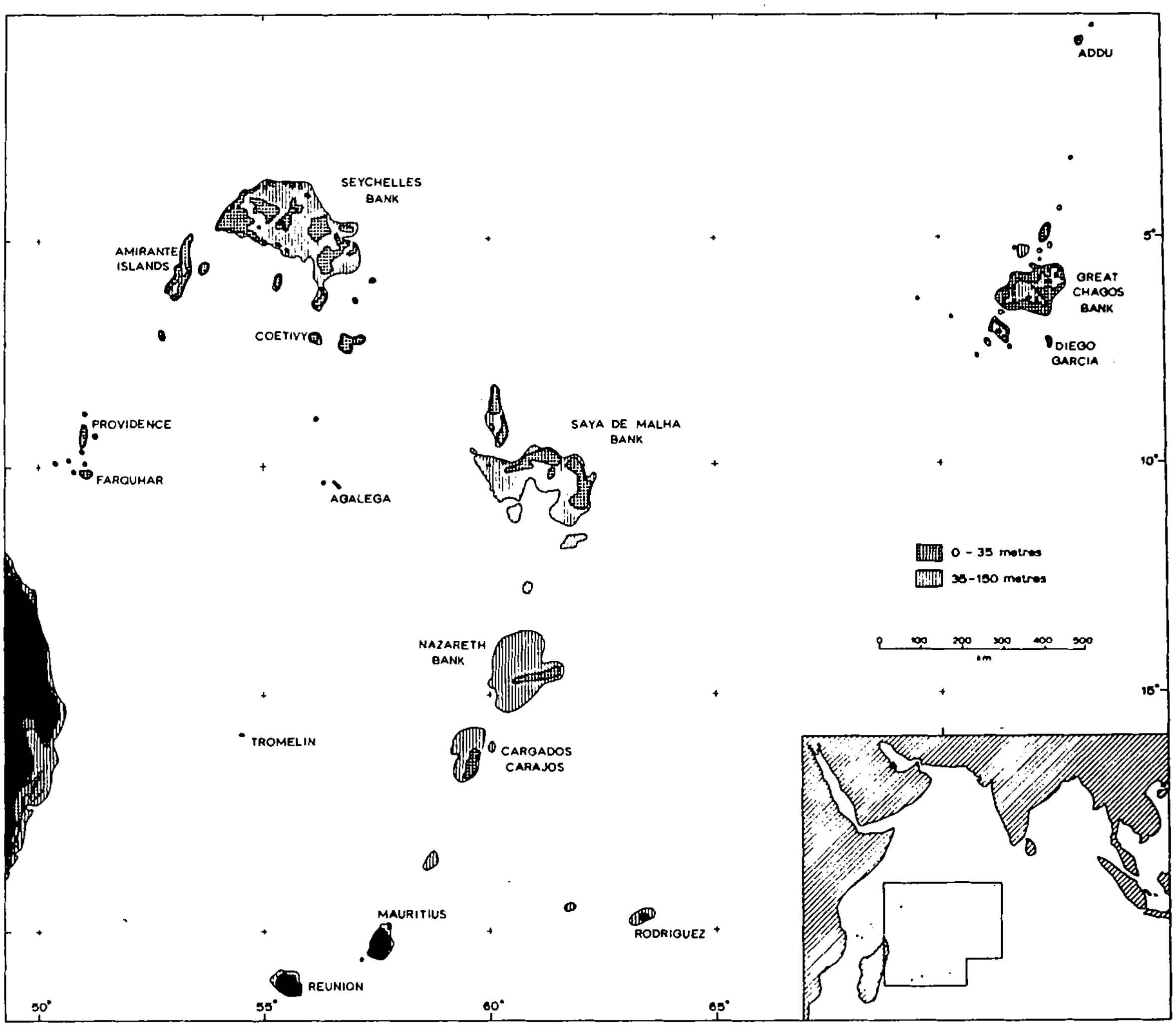

Figure 2. Land areas in the southwest Indian Ocean during the last glacial low stand of the sea (From Stoddart 1971c).

Changes of these magnitudes and consequences are of recent date and have been rapidly achieved. Sea-levels of the last glaciation were more than 100 meters below present sea-level until about 15,000 years ago. The major contribution of water back to the oceans resulted from the collapse of the 4000 km wide Laurentide ice sheet, which disappeared in about 4000 years, its margin retreating by up to 5 meters per week (Andrews 1973). Sea-level rise as a result averaged 1 meter/100 years over several thousand years to 6000 B.P. Because of differential adjustment of the crust to the new water load and also because of decantation of water from isostatically rising land, the transgression in many areas continued after this date, though at different rates in different parts of the world. Figure 3 gives a series of sea-level curves from different areas for the last few thousand years, and shows how recently

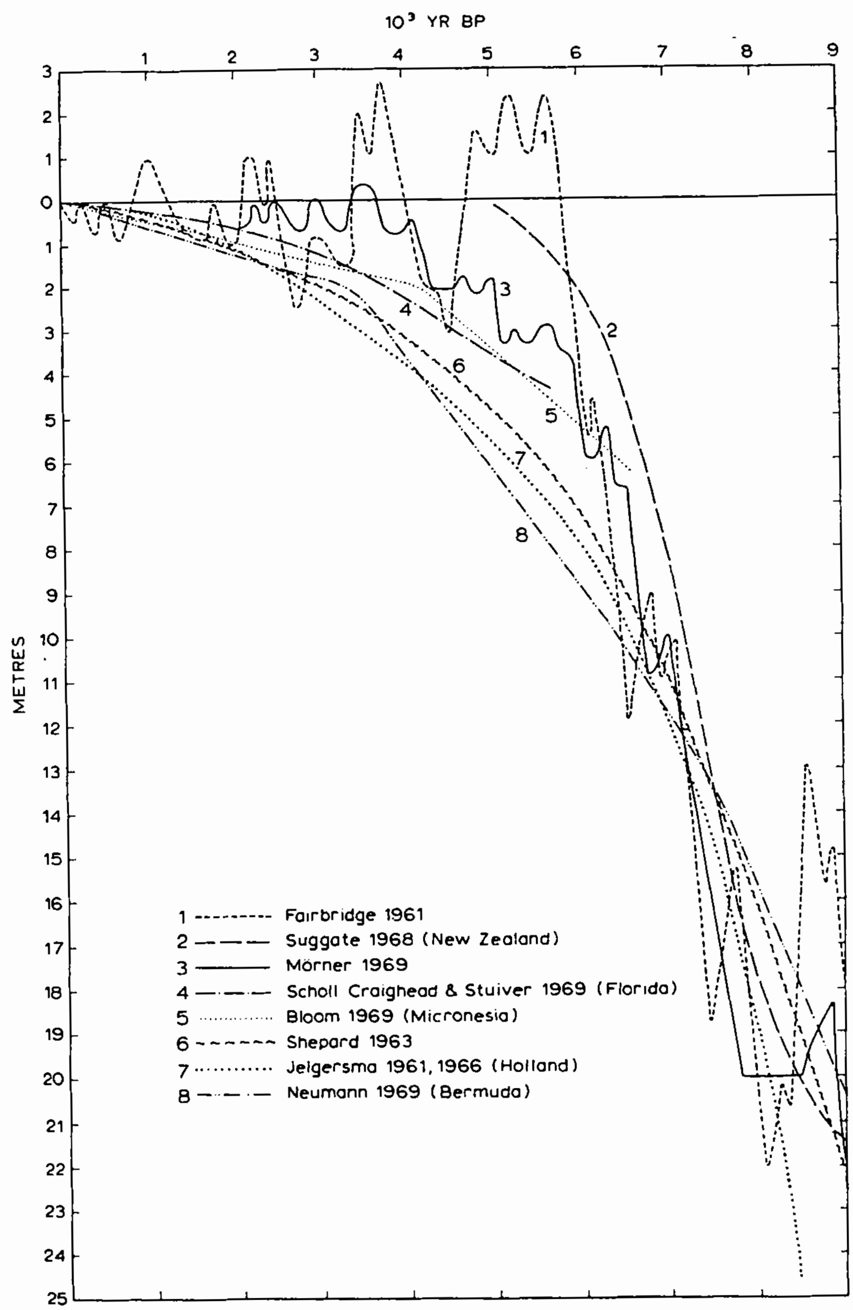

Figure 3. Rise in sea-level over the last 9000 years (From Mörner 1971).

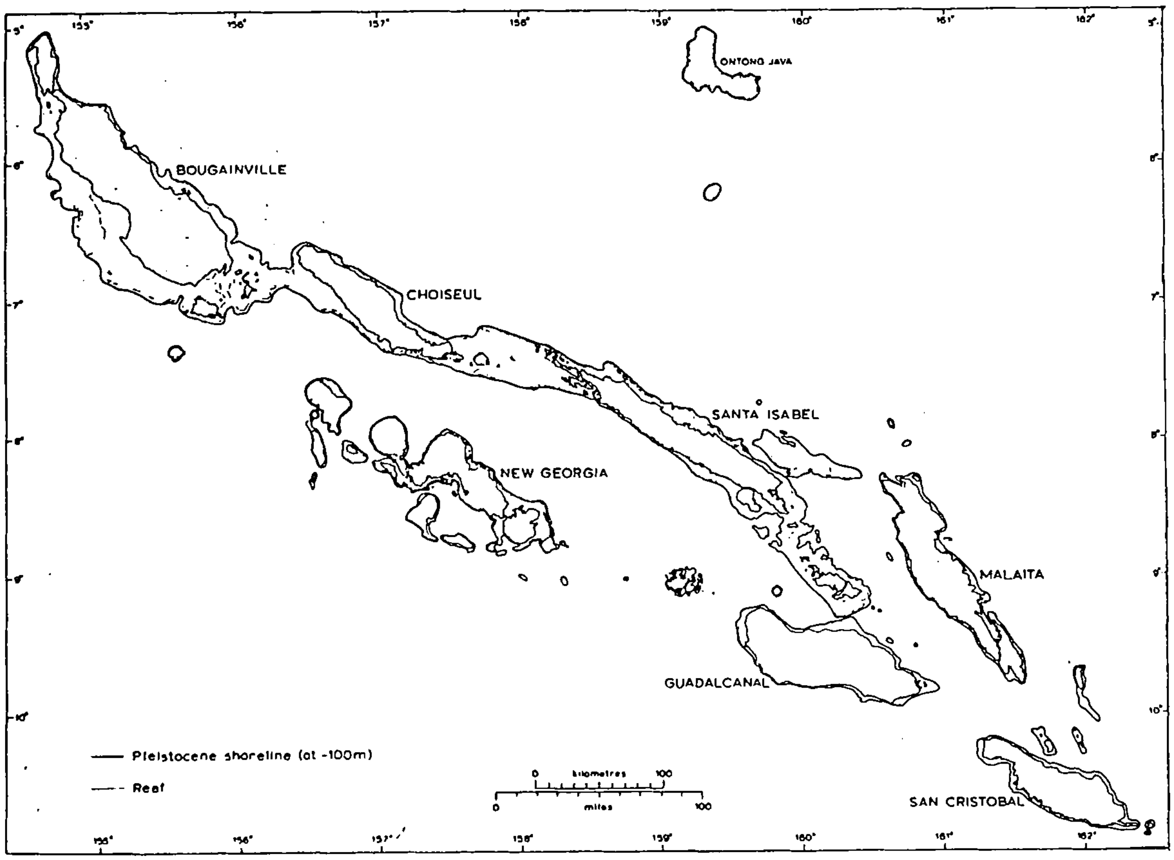

Figure 4. Topography of the Solomon Islands archipelago during the last glacial low stand of the sea and location of modern coral reefs.

the sea has risen from its glacial low level, and for how short a time (2000-3000 years, or 0.1 per cent of the length of the Pleistocene) it has stood at its present level. It is worth noting that while one almost automatically thinks of sea-level change in the vertical dimension, perhaps because of the ubiquitous convention of constructing sea-level curves such as those in Figure 3, in many shelf areas the sea transgressed horizontally across flat-lying lands now shallowly submerged, at rates of up to 10-15 meters per day (Galloway and Löffler 1972). Many modern reefs started growing 6000-7000 years ago when the sea was lower, and have been isolated from present island shores by this process of horizontal transgression; the reefs and islands of the Solomons in the southwest Pacific (Figure 4) form a dramatic example.

Recently it has become clear that since sea-level reached its approximate present level, different areas in the reef seas have had very different histories. This is well illustrated by comparison of the Holocene sea-

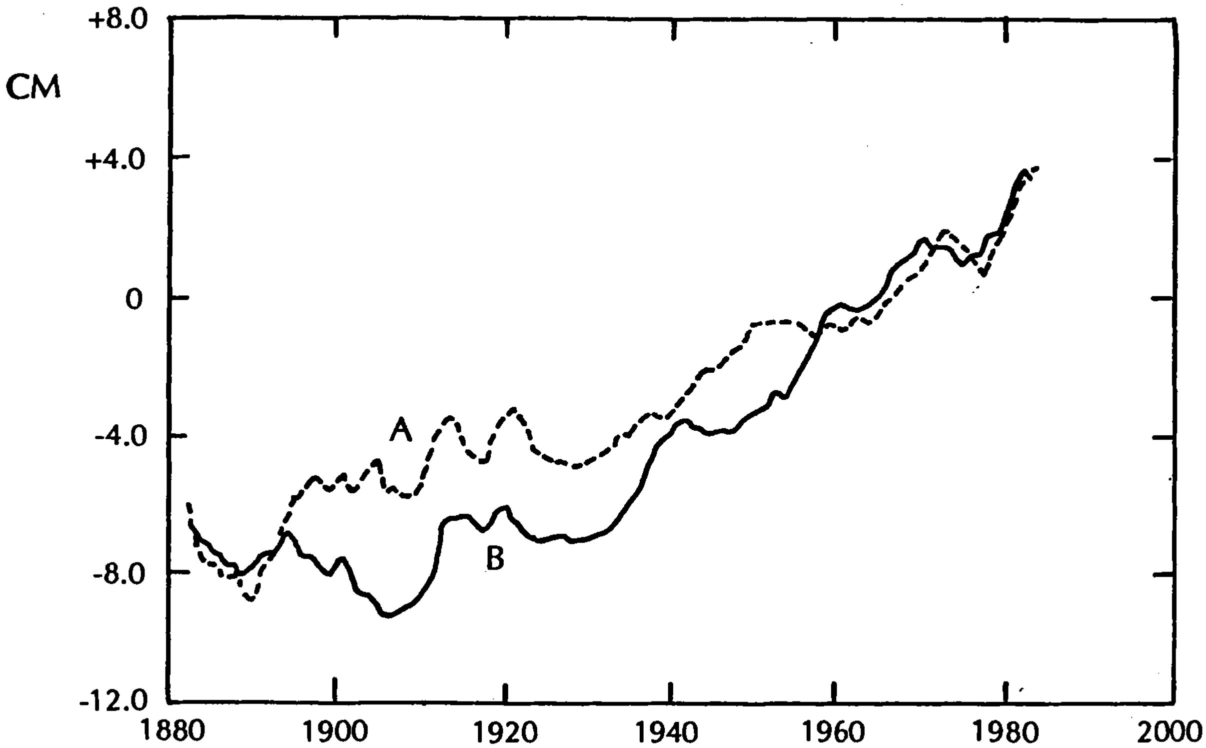

Figure 5. World rise in sea-level since 1880 from tide gauge records. Curve A based on Gornitz and Lebedeff (1987); Curve B based on Barnett (1988). Zero is taken as mean sea-level for the period 1951-1970 (From Gornitz 1991)

level curves of Chappell (1982) for the northern Great Barrier Reef and of Scholl and Stuiver (1967) and Scholl et al. (1969) for the Gulf coast of peninsular Florida. In northern Queensland, the sea reached its present level at least 6000 years ago, rose to a level 1-2 meters above the present, and then monotonically declined to its present position. In Florida, on the other hand, the sea stood 6 meters below its present level at 6000 B.P., and since then has continued to rise at a decelerating rate towards the present level. Pirazzoli and Montaggioni (1986, 1988a, 1988b) and Pirazzoli et al. (1985, 1988) identify a high stand of about 1 meter in French Polynesia between about 4000 and 1500 B.P. Similar evidence has been adduced from the Cook Islands (Woodroffe et al. 1990b), Fiji (Nunn 1990b), and the eastern Indian Ocean (Woodroffe et al. 1990a). Conversely Montaggioni (1979) found no such evidence in the Mascarenes, and this remains the case in the western Atlantic.

Several workers have calculated trends in sea-level rise on a worldwide basis over the period of instrumental tide-gauge records (Gornitz 1991; Figure 5). Thus Barnett' s examination of world tidal records for the period 1881-1980, excluding isostatically anomalous areas, showed a mean rise of 0.14±0.01 meters per 100 years. Gornitz and Lebedeff (1987) calculated a worldwide rate of rise of 0.17 meters per 100 years. Barnett's (1984, 7982-7984) data set includes ten stations in the reef seas, including three atolls. These stations give a mean rate of rise of 0.112 meters per 100 years over the period 1901-1947. Pirazzoli (1986), using the same data sets, derived a mean rise for six reef-sea stations of 0.17 meters per 100 years , which is the same as the worldwide rate calculated by Gornitz and Lebedeff (1987). Inter-station variability is, however, considerable, reaching one-third of a meter in the case of Barnett's (1984) reef stations, and more recently Pirazzoli (1989) has reduced his estimate of sea-level rise over the past century (at least for European stations) by a factor of 2-3, to only 0.04-0.06 meters per 100 years.

Predictions of future sea-level rise over the next several decades, resulting from the 'greenhouse effect', are of the same general magnitude as that of the main postglacial transgression (Stoddart 1990). Among recent estimates Stewart et al. (1990) predict a rise of 0.6 meters by 2050, equivalent to a rate of 1 meter per 100 years, with an uncertainty of a factor of three.

In addition to secular changes, sea-level is also subject to short-period fluctuations, often associated with meteorological conditions or seismic events (see also the discussion of storm surges and tsunamis below). In the Pacific such fluctuations are strikingly correlated with El Niño-Southern Oscillation events. Over the period 1981-1983 these fluctuations had amplitudes of 55-60 cm at Jarvis, Kiritimati [Christmas] and the Galapagos Islands, and 80 cm at Nauru (Lukas et al. 1984). El Niño sea-level fluctuations at Truk have an amplitude of ca 30 cm, compared with 10 cm in non-El Niño years (Cane 1986, 51). During the 1972 event monthly mean sea-level at Guam fell 44.2 cm below mean sea level, in an area with a tidal range at springs of 1 meter. This resulted in extensive mass mortalities of emersed reef-flat organisms, particularly corals, echinoderms and molluscs (Yamaguchi 1975). Substantial fluctuations in populations of algae and sea-grasses related to periods of high and low sea-level have also been documented at Punta Galeta, Caribbean coast of Panama, by Cubit (1985), and used to speculate on the biological consequences of predicted 'greenhouse' sea-level rise. Hopley and Kinsey (1988) have suggested that coral growth could be at least temporarily accelerated as reef-flats deepen from their present frequently low-intertidal levels.

Predicted sea-level rises are also likely to have substantial physical as well as biological consequences. Thus a rising sea-level is likely to lead to erosion and landward retreat of unconsolidated beaches (Bruun 1962, 1983, 1988; Schwartz 1967), a phenomenon which would be exacerbated by increased storminess. The Ghyben-Herzberg freshwater lens on small islands would also be endangered by rising sea-levels (Buddemeier and Holladay 1977; Wheatcraft and Buddemeier 1981; Oberdorfer and Buddemeier 1988; Woodroffe 1989), and this could have profound consequences for terrestrial vegetation, water supply, and even the habitability of islands. Coastal mangrove woodlands are likely to be eroded and may locally face extinction, depending on the rate of sea-level rise (Ellison and Stoddart 1990).

In the year that the proceedings of the Honolulu symposium were published (Fosberg 1963), MacArthur and Wilson (1963, subsequently elaborated in 1967) proposed their equilibrium theory of island biogeography. This theory laid emphasis on the island attributes of distance from source of propagules and of area as factors controlling the probability of successful colonisation and subsequent survival by species in the terrestrial biota of an island. They hypothesised that the relative rates of colonization and extinction resulted in an equilibrium species number independent of the taxa present. They had but two sets of reef island data, both dealing with the distribution of vascular plants. One of these, for the individual islets of Kapingamarangi Atoll in the Carolines (Niering 1963), has been inferred by others to apply to atolls in general. These data do not in fact support the simplistic interpretation placed on them by MacArthur and Wilson (Whitehead and Jones 1969; Stoddart 1975). Their other data set from the Dry Tortugas was deeply flawed and their interpretation of it cannot be sustained (Stoddart and Fosberg 1981). We shall show later that the main controls on biotic diversity on reef islands are ecological, rather than to do with either area or location (Stoddart 1992; Williamson 1989)

Their proposals have greater relevance, however, in the case of higher islands, either elevated reef (makatea) islands or volcanic islands, for both of which area can be interpreted as a surrogate for habitat diversity. This has been well demonstrated for terrestrial arthropods, mollusca and birds in the closely-set islands of the southwest Pacific (Diamond 1972; Greenslade 1968, 1969; Peake 1969, 1971), as well as for oceanic volcanic archipelagoes (Juvik 1979).

Most workers who have considered MacArthur and Wilson's theory have considered the parameters of distance from source and island area to be invariant over time. On geological time-scales this is clearly not the case: the contrast over island life history between the leeward Hawaiian atolls (Kure, Midway), which are moving out of the reef seas, and islands such as Pitcairn, which is moving into them, has previously been noted (Stoddart 1976). It has also been suggested that the presence of endemic palms such as *Pritchardia* and other upland plants, as well as land snails and birds, on reef islands such as Laysan may be relict indicators of a complex plate-tectonic history (Schlanger and Gillett 1976; Rotondo et al. 1981). It is conceivable that the biota of the anomalously old island of Mangaia in the southern Cooks could give similar indications.

On a Pleistocene time-scale, Simpson (1974) found in the case of the Galapagos Islands that island areas and inter-island distances at the time of glacial low sea-levels gave better explanations of biotic diversity than did present conditions. Potts (1983, 1984, 1985) has made a similar case for the marine biotas of continental shelf reef localities. In the case of Aldabra Atoll, western Indian Ocean, now slightly emergent, the record of raised reef limestones shows a sequence of submergence and emergence of the atoll during the late Pleistocene, with at least two periods of total submergence and periods of low sea-level when the exposed land area reached ca 360 sq km

(Braithwaite et al. 1973). The record of one Holocene and three Pleistocene marine faunas shows repetitive colonisation and extinction consequent on changes in area, elevation and topography resulting from sea-level change, as demonstrated by Taylor's (1978) analysis of molluscan faunas. Similar effects are shown by terrestrial faunas, including crocodiles, tortoises, lizards, land birds, and molluscs (Taylor et al. 1979). Likewise in the central Pacific, Paulay (1990) has examined islands with varying degrees of tectonic uplift as a surrogate for Pleistocene sea-level change, and has shown how habitat changes resulting from differential emergence have affected marine invertebrate faunas. Such processes of extinction and colonisation in both terrestrial and marine habitats must have been universal on islands throughout the Pleistocene.

TSUNAMIS

Tsunamis are long-period (15-20 minute), long wave-length (ca 270 km) waves with small amplitude (1-2 meters) in deep water, generated seismically or volcanically principally at subduction zones and travelling across the ocean with propagation velocities of the order of 800 km/h. They are most frequent in the Pacific where most have their origin in the subduction zones off Chile and Alaska. Such long-period waves have little impact on atolls rising steeply from the deep ocean, and their effects are severely dampened by barrier reefs and within lagoons. But on shield-volcano islands such as Hawaii and the Marquesas, especially with deep embayments, tsunamis may cause serious coastal inundation. 14 tsunamis have been recorded in the Hawaiian Islands since 1830, of which six have been serious and one (that of 1 April 1946) disastrous. This latter locally raised sea-level at Hilo, Hawaii, 16.8 meters (Shepard et al. 1950) and by 10 meters in the Marquesas. By contrast the same event raised sea-level at Rangiroa and Hao Atolls in the Tuamotus by 2.2-2.3 meters above mean sea-level (the reef flats stand at ±0.3 meters). Another major tsunami occurred on 22 May 1960, giving similar elevations in Hawaii (Eaton et al. 1961; Cox and Mink 1963), much reduced inundations at Rarotonga, Samoa and in French Polynesia (Keys 1963; Vitousek 1963), and negligible effects on atolls. 121 tsunamis have been recorded at Hawaii since 1837. These included one of 6 meters in 1837, 9.1 meters in 1869 and 1896, 6.1 meters in 1923, 6.5 meters in 1933, 16.8 meters in 1946, 10.4 meters in 1952, 16 meters in 1957, and 10.5 meters in 1960 (these are maximum elevations and usually local; island-wide inundations were much less). The effects of such tsunamis are primarily economic and social, though Bourrouilh-Le Jan and Talandier (1985) have suggested that the mega-blocks on the southwestern reef flats at Rangiroa Atoll, Tuamotus, which are usually attributed to hurricanes, may be tsunami-generated.

The most extreme tsunami-type effects appear to be locally generated rather than trans-oceanic in origin (Latter 1981). The best documented is that which impacted Ishigaki in the southern Ryukyus in 1771, following a seismic event 40 kilometers to the south-south-east. This produced the world's highest tsunami wave, reaching a maximum of 85.4 meters, and lodging large coral blocks at Ohama on the southeast side of the island (Miyagi et al. 1980, 84, 88). Moore and Moore (1984) have proposed that limestone blocks at up to 325 meters on the island of Lanai, Hawaii, have a similar origin. The tsunami generated by the 1883 explosion of Krakatau caused local inundation up to 50 meters on Java and Sumatra and killed 36,000 people (Bullard 1962; Latter 1981).

PATTERNS IN RAINFALL

After location and morphology, climate is probably the single most important factor influencing island ecology and habitability. Of the climatic elements rainfall is the most significant, the most variable, and the best documented (Brookfield and Hart 1966; Taylor 1973; Stoddart 1971a), though one should beware of inferring the uniformity of even low atoll rainfall regimes from single-station records (Stoddart 1983). Thomas (1963) outlined the spatial variations in mean annual rainfall over the Pacific Ocean (Figure 6 gives a revised rainfall map of part of the central south Pacific) and similar maps are available for the Indian Ocean (Figure 7).

The wettest atolls, such as Jaluit, Palmyra and Funafuti in the Pacific and Peros Banhos in the Indian Ocean have mean annual rainfalls of about 4 meters: the heaviest recorded atoll rainfall occurred at Funafuti, with 6.7 meters in 1940. The driest atolls, such as Wake and Canton, may have 0.5 meters or less (Table 2). High-island stations, such as Vanikoro, Pohnpei [Ponape] and Kosrae [Kusaie], have annual means over 5 meters, though in mountainous situations the rainfall may be considerably more. Many island groups, such as the Line Islands, the Marshalls and the Gilbert Islands, and the Chagos-Maldives-Laccadives chain, show pronounced latitudinal gradients in rainfall from very wet to very dry over quite short distances.

The gradient from Wake through the Marshalls and Gilberts to Tuvalu [the Ellice Islands] at 165-175° East longitude, is among the best known (Figure 8), and that from Johnston through the Lines, Phoenix and Cook Islands is comparable (Figure 9). This situation is made more interesting because here Fosberg (1954) has defined a series of latitudinal vegetation zones ('Fosberg zones') delimited by magnitudes of mean annual rainfall. Annual means form a crude index of rainfall input, however, and hence we concentrate on identifying some of the more important components of variability in rainfall patterns, using data from those atolls and high islands in the tropical seas where long records are available.

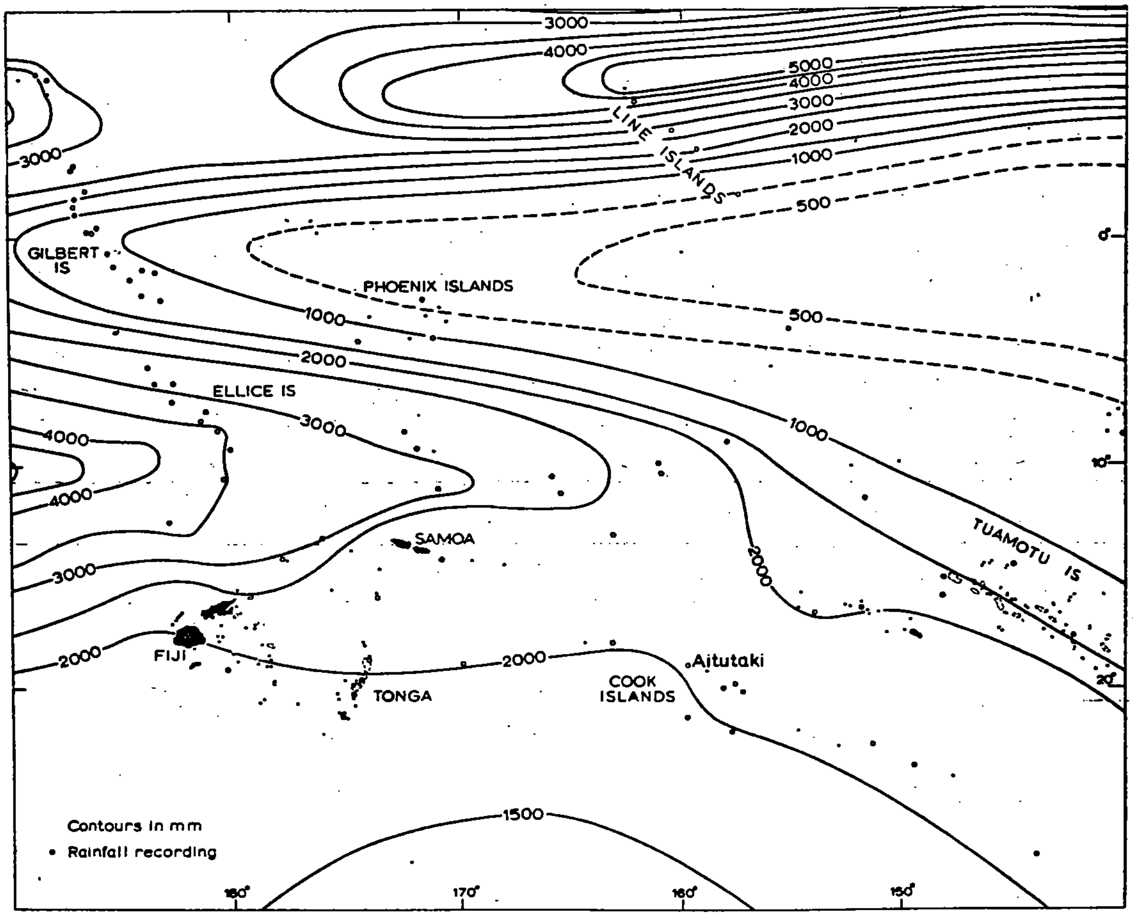

Figure 6. Distribution of mean annual rainfall (mm) in the central South Pacific Ocean.

<u>Temporal variation in annual rainfall</u>

Ten- and twenty-year moving averages reveal significant periodicities in many island records. These have been calculated for 7 Pacific, 6 Indian Ocean, and 7 West Indian stations. Figures 10, 11 and 12 give sample time series for Apia, Samoa; Mahe, Seychelles; and stations on Dominica, Barbados and Trinidad. Tables 3, 4 and 5 quantify the main temporal variations for 3 Pacific, 6 Indian Ocean, and 4 West Indian stations.

In the West Indies the analyses suggest the existence of four rainfall 'epochs' during the last 100 years:

Late 19th century	Pre-1898	Very wet
Early 20th century	1899-1928	Relatively dry
Mid 20th century	1929-1958	Relatively wet
Recent	1959-1989	Very dry

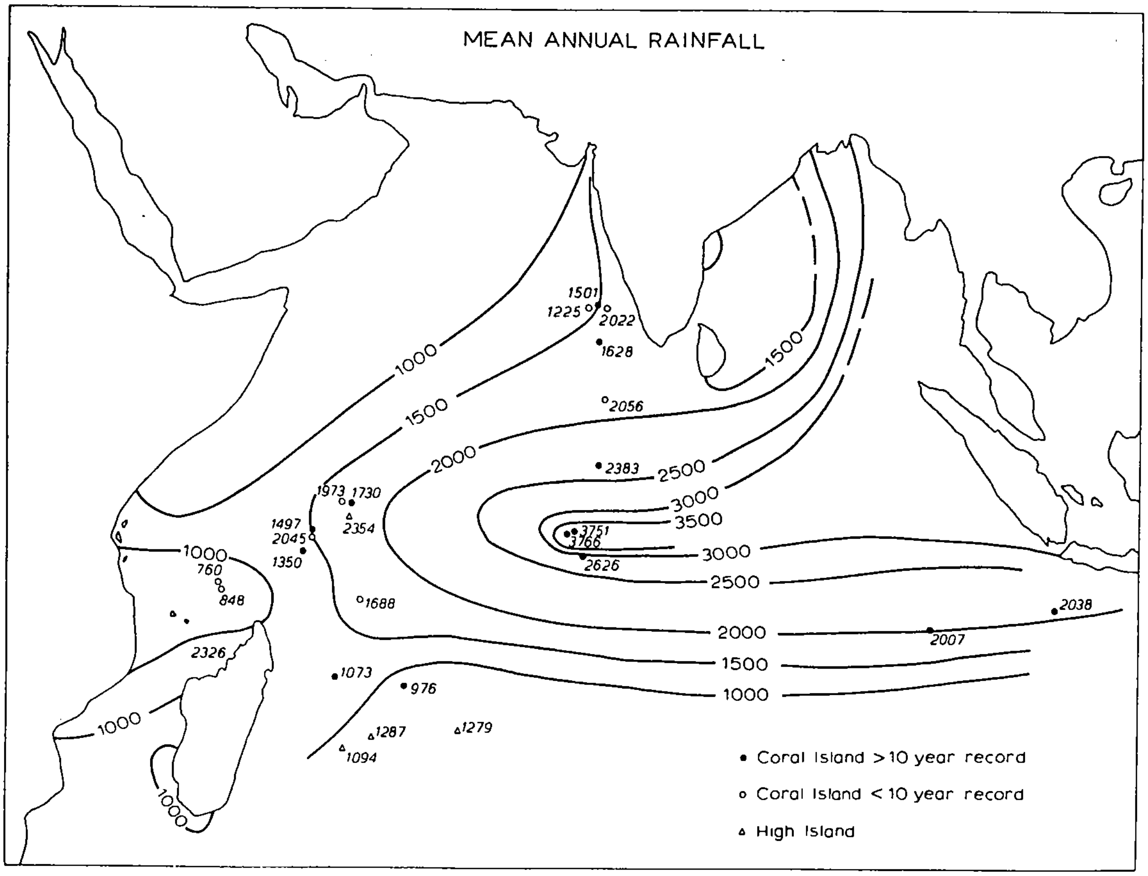

Figure 7. Distribution of mean annual rainfall over the Indian Ocean from coral island data (From Stoddart 1971a).

The magnitude of changes between these periods is substantial: in Barbados, Dominica and Guyana, stations have present-day mean annual rainfalls of 384, 296 and 610 mm (15, 12 and 24 inches) less than means in the late nineteenth century. Kraus (1955) has documented similar periodicities in eastern Australian and eastern North American stations It is interesting to note that Trinidad has recently shown a return to wetter conditions in contrast to the continued very dry conditions in the more northerly islands of Barbados and Dominica.

In the Pacific records are generally too short for all the epochs to be demonstrated, but good records for Fiji, Western Samoa and Rarotonga suggest a pattern of:

Early 20th century	Very dry
1920-1930	High rainfall
1940-present	Intermediate rainfall

Table 2. Mean annual rainfalls of sample atolls

Jaluit	4033 mm
Peros Banhos	3999
Palmyra	3810
Majuro	3048
Lamotrek	2645
Diego Garcia	2599
Kwajalein	2032
Mopelia	1854
Tarawa	1626
Rangiroa	1473
Eniwetok	1346
Aldabra	1000
Onotoa	980
Hull	838
Wake	610
Canton	432

Again the magnitudes of the differences between these periods are substantial. The 1906-1941 mean for Suva, Viti Levu, was 711 mm (28 inches) greater than that for 1883-1905; for Apia, Western Samoa, the 1920-1939 mean was 406 mm (16 inches) greater than that before 1920. Similarly in the Indian Ocean the record for Mahe, Seychelles, shows striking changes over the past century:

1891-1904	Very wet
1905-1922	Less wet
1923-1937	Very wet
1938-1954	Considerably less wet
1955-1968	Very wet
1969-1989	Less wet

The annual rainfalls in less wet periods were 500-700 mm lower than in the wetter periods (Walsh 1984).The ten- and twenty-year moving averages thus identify substantial secular changes. For present purposes it is sufficient to establish that they exist: we are not concerned to determine causes or even to establish synchroneity. Indeed, not all islands show variations of such magnitude and phase (Honolulu, Hawaii, for example, does not), though they

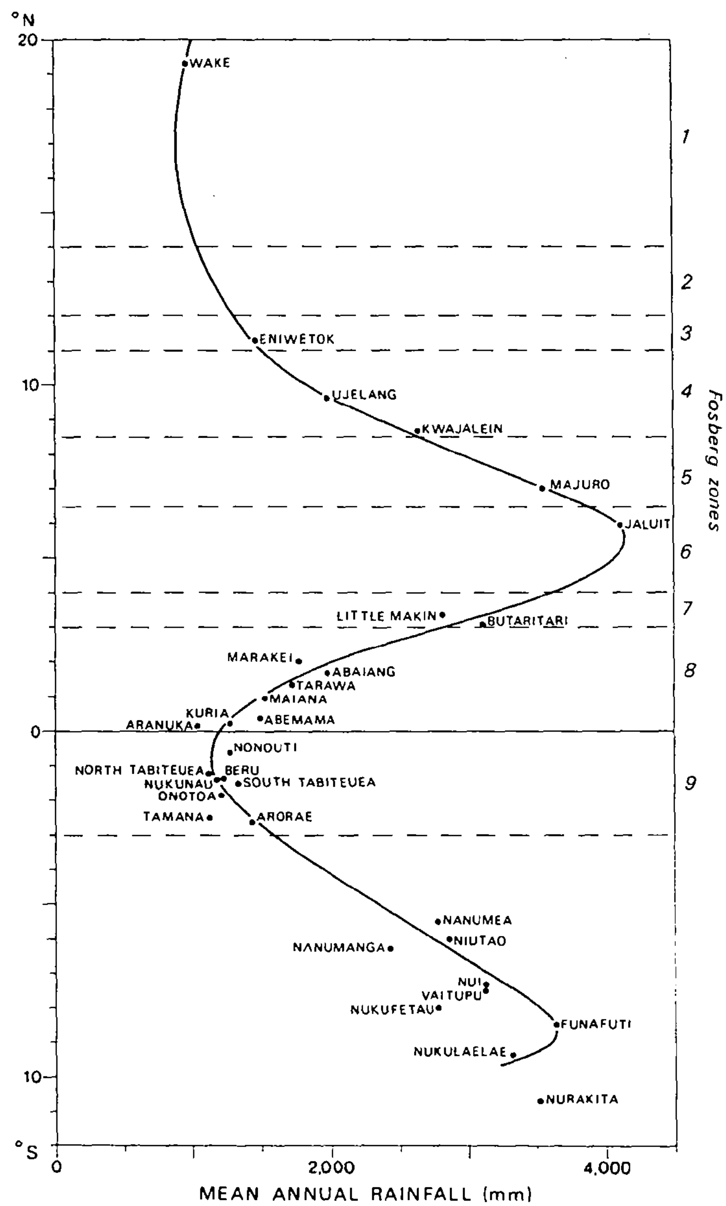

Figure 8. Latitudinal variation in mean annual rainfall in the Marshall Islands, Gilbert Islands and Tuvalu (revised from Fosberg 1956 using data in Taylor 1973).

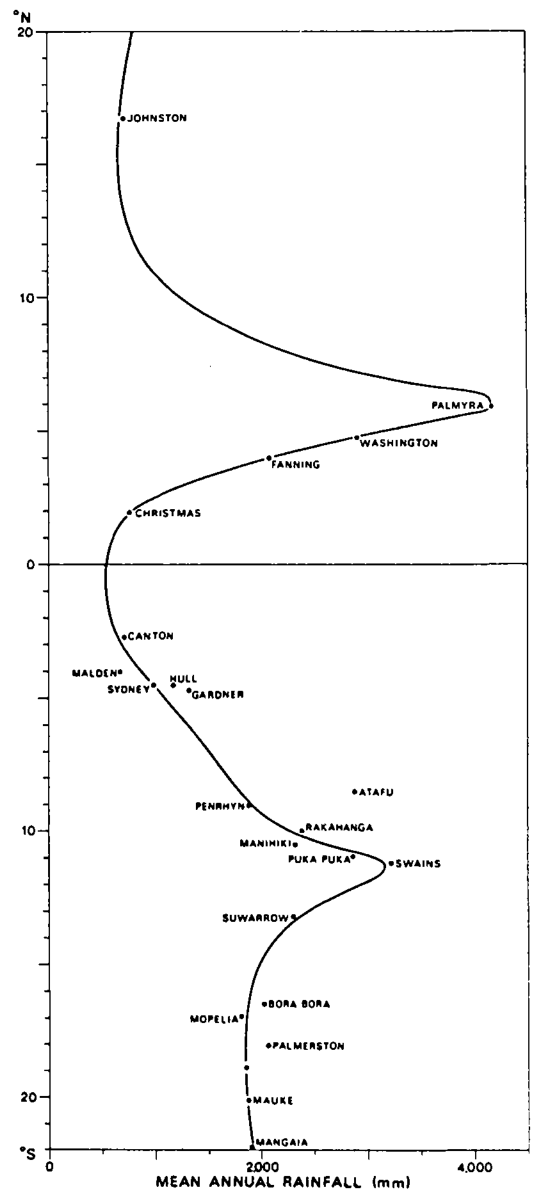

Figure 9. Latitudinal variation in mean annual rainfall from Johnston Island to the Line Islands, Phoenix Islands and Cook Islands (based on data in Taylor 1973 and later records).

Table 3. Changes in Pacific Ocean rainfalls

Station	Territory	Period of Record	Mean annual rainfall, mm				
			1883–1905	1906–1941	Change (%)	1942–1969	Change (%)
Suva	Fiji	1883–1969	2624	3348	+724 (27.59)	2945	-403 (12.04)
			1890–1919	1920–1939	Change (%)	1940–1971	Change (%)
Apia	Samoa	1890–1971	2718	3132	+414 (15.23)	2070	-262 (8.37)
			1899–1905	1906–1939	Change (%)	1940–1971	Change (%)
Rarotonga	Cooks	1899–1971	1872	2161	+289 (15.44)	2002	-159 (7.36)

Table 4. Changes in Indian Ocean rainfalls

Station	Territory	Period of record	Mean annual rainfall, mm						
			–1906	1907–1928	Change (%)	1929–1958	Change (%)	1959–1974	Change (%)
Amini Divi	Laccadives	1889, 1892-1974	1339	1513	+ 13.1%	1506	- 0.5	1573	+ 4.4
Minicoy	Maldives	1891-1974	1606	1676	+ 4.36	1566	- 6.56	1747	+ 11.56
Mahe	Seychelles	1891-1974	2575	2361	- 8.31	2224	- 5.80	2528	+ 13.67
Zanzibar	Zanzibar	1892-1950	1634	1424	- 12.85	–	–	–	–
Royal Alfred	Mauritius	1875-1974	1214	1306	+ 7.58	1313	+ 0.54	1349	+ 2.74
Tananarive	Madagascar	1890-1974	1396	1312	- 6.02	1278	- 2.59	1331	+ 4.15

Table 5. Temporal changes in West Indian rainfalls

Station	Territory	Period of record	Mean annual rainfall, mm						
			Before 1898	1899–1928	Change, mm(%)	1929–1958	Change, mm(%)	1959–1989	Change, mm(%)
Roseau	Dominica	1865-1989	2114	1874	- 240 (11.35)	2037	+ 165 (+ 8.80)	1820	- 217 (- 10.65)
Bridgetown/ Grantley Adams Airport	Barbados	1853-1988	1491	1207	- 284 (-19.05)	1376	+ 169 (+ 14.00)	1107	- 259 (- 18.82)
Port of Spain	Trinidad	1862-1968	1701	1507	- 194 (-11.41)	1666	+ 160 (+ 10.62)	1592	- 74 (- 4.44)
Georgetown	Guyana	1886-1972	2601	2237	364 (-13.99)	2414	178 (+ 7.96)	1969	- 446 (- 18.48)

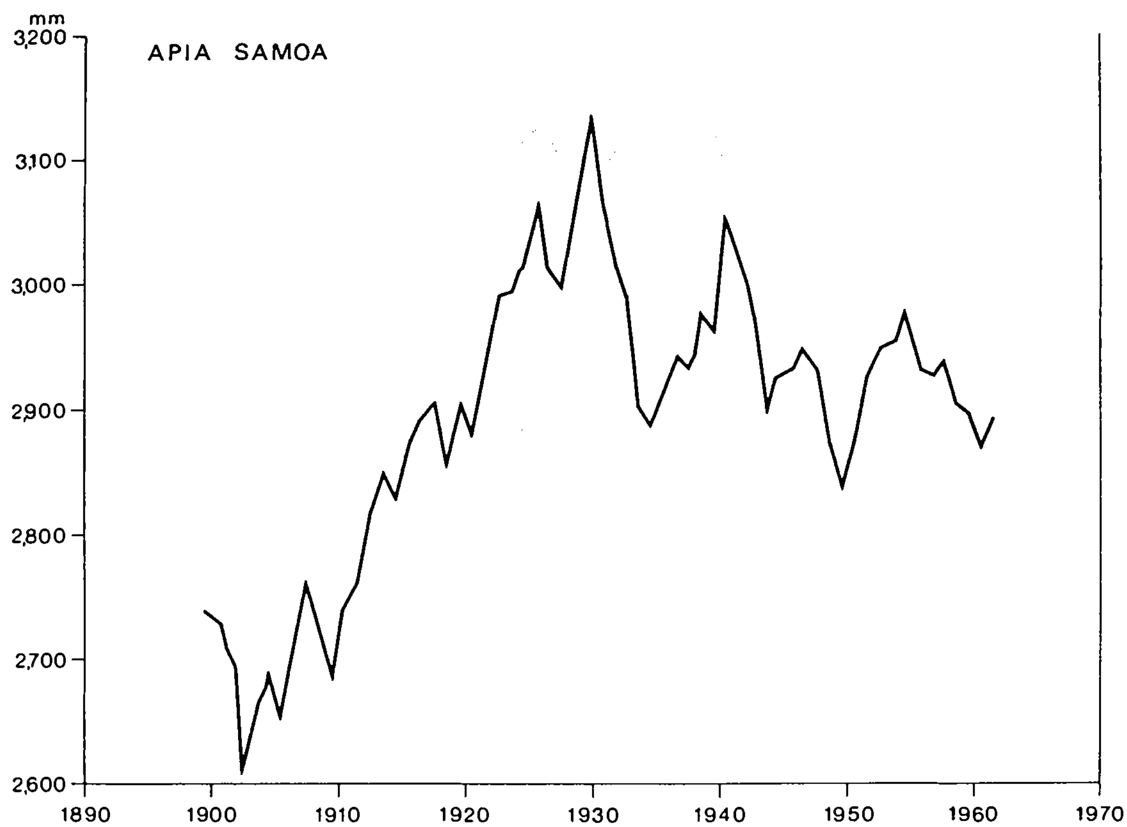

Figure 10. Twenty-year running means of annual rainfall at Apia, Western Samoa.

do seem to be characteristic of many tropical high and low islands over the last century. Where similar fluctuations have been identified in continental areas, however, it has often been shown that there are substantial spatial variations in the trends revealed (Stewart 1973; Parthasarathy and Dhar 1974).

Higher-frequency variations have also been identified using power spectrum analysis on records for 5 Caribbean and 10 Pacific Ocean islands with records ranging in length from 37 years (Banaba) to 116 years (Barbados). In the Caribbean the main cycles identified are long-term (more than 44 years), thus confirming the conclusions from the moving-average analysis; but cycles of 4.5, 5.5 and 7 years are also statistically significant. In the Pacific stations the 5.3 year cycle is particularly strong, though stations with longer records, such as Willis (in the Coral Sea), Fanning (Line Islands) and Funafuti, show significant cycles in the range 12-30 years.

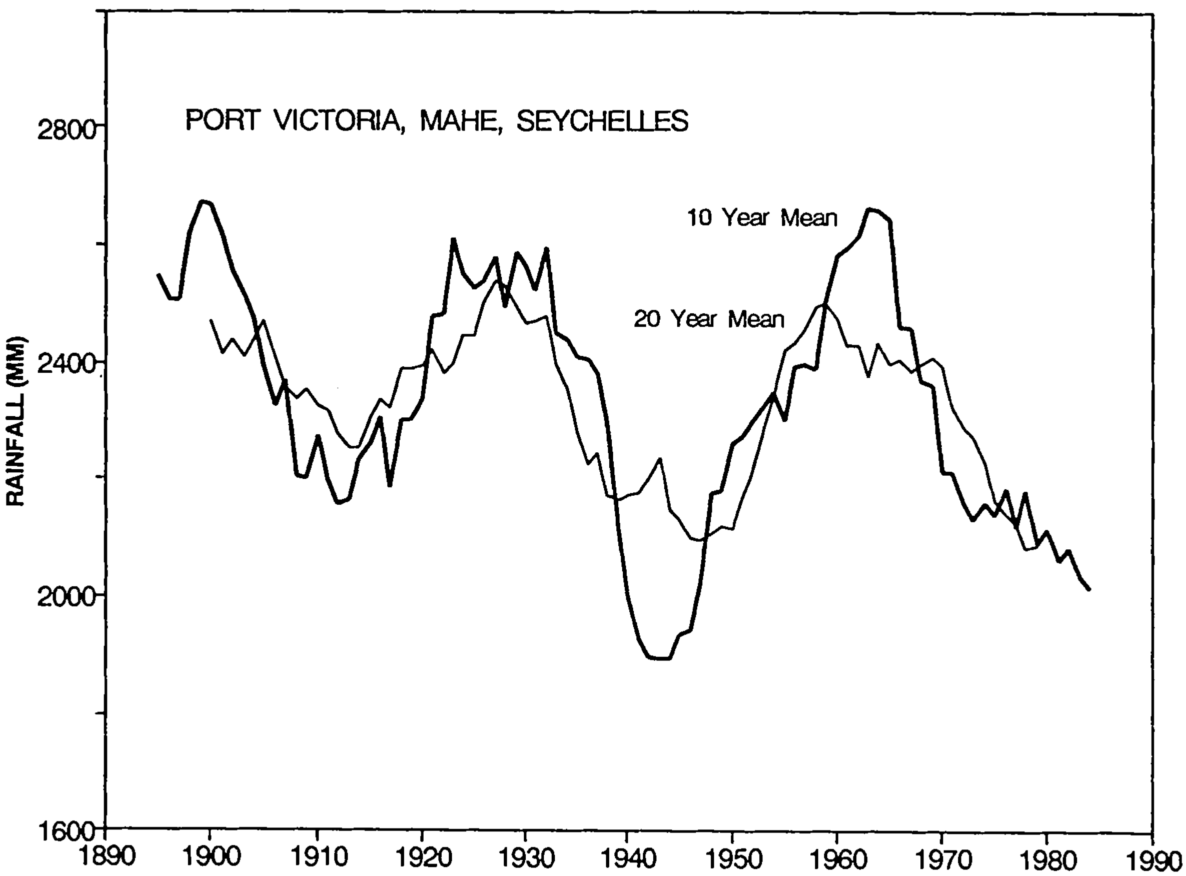

Figure 11. Ten- and twenty-year running means of annual rainfall at Port Victoria, Mahe, Seychelles.

Short-period high-magnitude fluctuations are also of great importance in some areas, notably in the Pacific equatorial islands, where they are associated with the El Niño phenomenon of coastal Peru. Figure 13 gives monthly rainfalls for the period 1942-1972 for Canton, Hull, Sydney and Gardner Atolls in the Phoenix Islands, which show highly coherent patterns. Periods of substantial rainfall (e.g. early 1953, early 1958, late 1965) are separated by long dry periods, and the variations are not only concurrent throughout the group but occur generally in the central and eastern Pacific equatorial area, including the southern Gilberts and the southern Line Islands. Figure 14 plots the spatial distribution of annual rainfall over this area for a very dry year (1968) and a very wet year (1958). In a dry year rainfall along the equator may be only a quarter of the long-term mean, whereas in a wet year it may be twice as much.

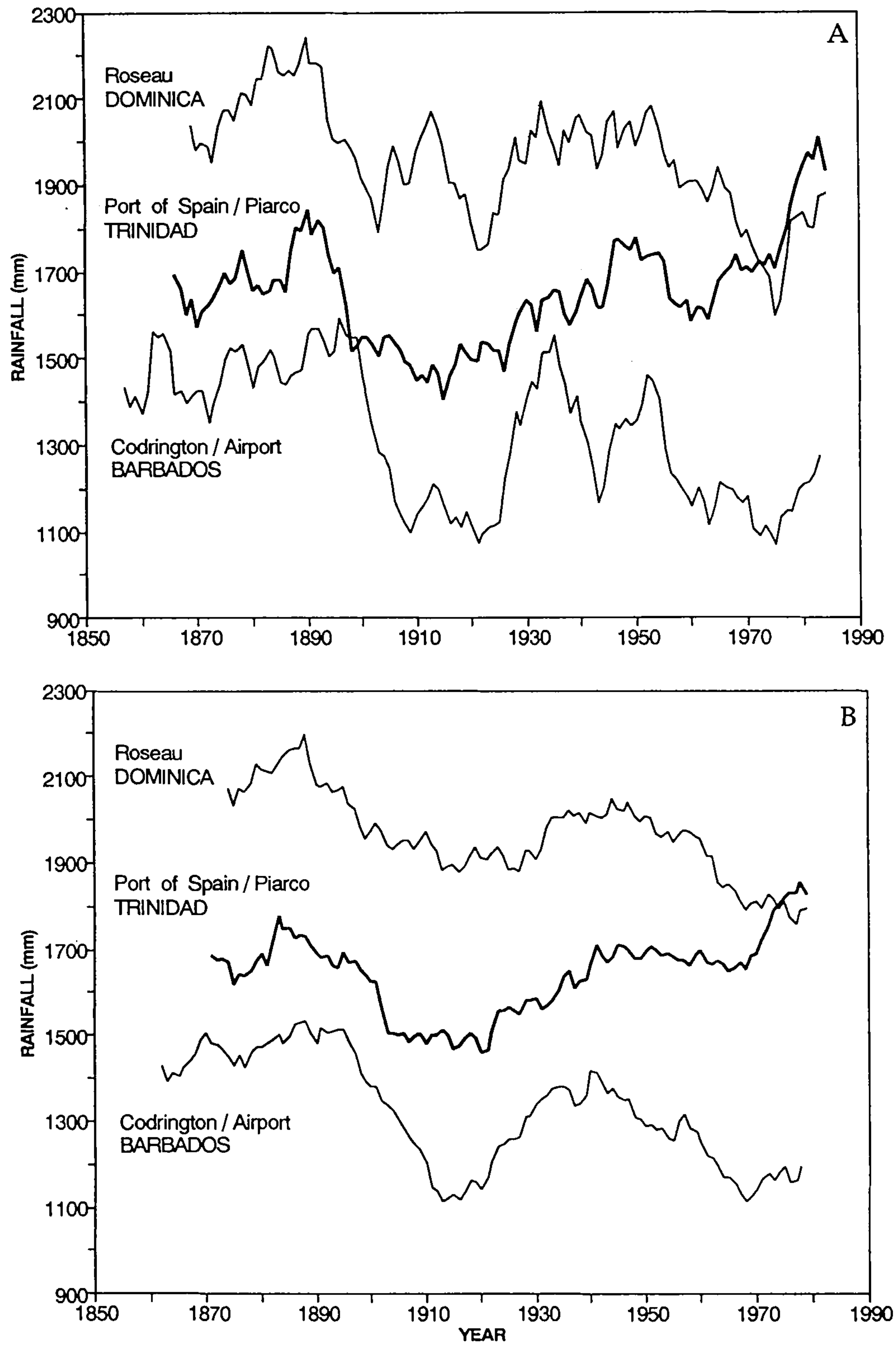

Figure 12. (A) Ten- and (B) twenty-year running means of annual rainfall at Roseau, Dominica; Port of Spain/Piarco, Trinidad; and Codrington/airport, Barbados.

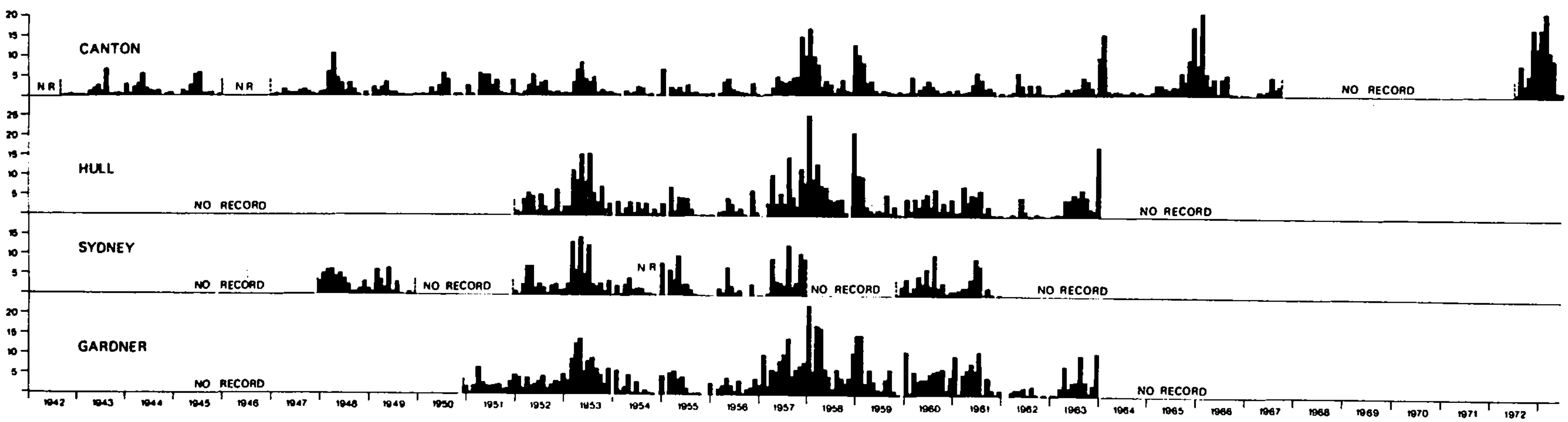

Figure 13. Variations in monthly rainfall at Canton, Hull, Sydney and Gardner Atolls, Phoenix Islands, 1942-1972.

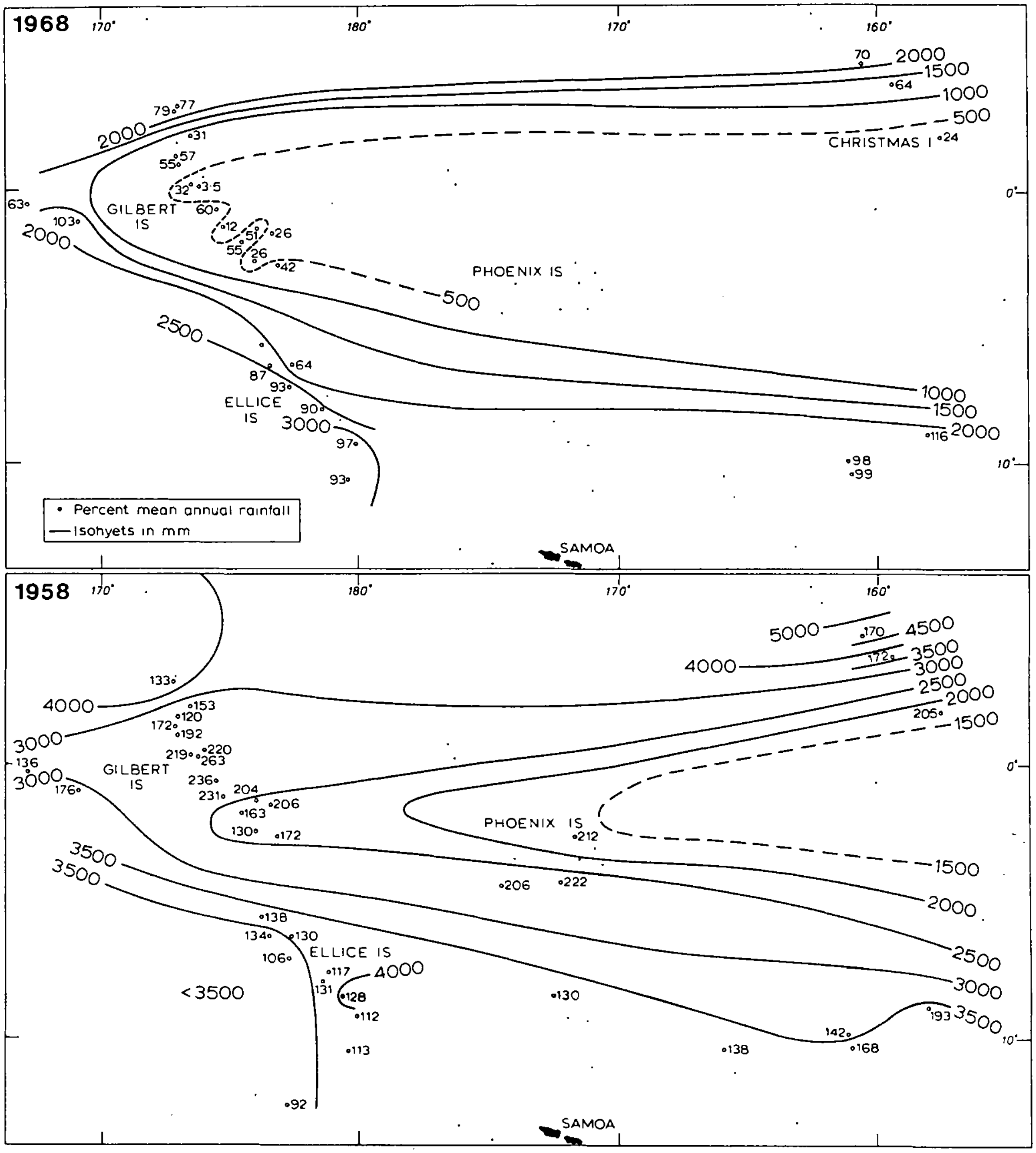

Figure 14. Rainfall distribution in the central equatorial Pacific in a dry (1968) and a wet (1958) year.

Table 6. Range of rainfall in the Southern Gilberts

Data for 1933, 1934, 1937-38, 1944-51 and 1953-69

Atoll	Wet Year 1940	Dry Year 1950	Ratio wet/dry
Little Makin	2811	1277	2.20
Butaritari	3946	1444	2.73
Marakei	3851	503	7.65
Abaiang	3555	316	11.26
Tarawa	3260	390	8.36
Maiana	3252	254	12.81
Abemama	3259	195	16.69
Kuria	2744	192	14.27
Aranuka	2836	149	19.02
Nonouti	2924	164	17.82
Tabiteuea	3275	190	17.26
Beru	3567	248	14.37
Nikunau	3763	162	23.19
Onotoa	3948	168	23.55
Tamana	3286	301	10.91
Arorae	3255	290	11.21

Data: Sachet (1957), and subsequent meteorological records.

Table 6 illustrates the magnitude of rainfall variations between events and how they increase from north to south in the Gilberts: note that on Nikunau a wet year (1940) brought 3759 mm (148 inches) compared with only 163 mm (6.4 inches) in the dry year of 1950. Extremes of these magnitudes are of obvious importance for atoll populations, more so than, for example, the predictable seasonal absolute droughts associated with monsoonal conditions on such atolls as Amini Divi, Lakshadweep (Laccadive Islands). Unpredictable droughts undoubtedly caused the abandonment of agricultural colonisation schemes on Sydney, Hull and Gardner Atolls in the Phoenix Islands, begun in 1938-1940 but abandoned during 1955-1963 (Knudson 1964). Two islands in this equatorial area (Banaba and Fanning) have rainfall records sufficiently long for power spectrum analysis: the most frequent cycles in both cases are short-term (3.6-3.7, 5.4-6.0 and 8.8-9.0 years).

The phenomena associated with these variations are reasonably well known (Ichiye and Petersen 1963; Bjerknes 1969; Rasmusson 1985; Cane 1986; Philander 1989, 1990), even though their causes are still not wholly understood. In normal years upwelling along the equator, extending westwards from the coast of South America, brings cool, nutrient-rich waters to the surface, and these support a zone of high productivity in the sea, forming the basis of the old 'On the Line' whale fishery, and also supporting large seabird colonies and guano deposition on equatorial islands. Under these conditions rainfall is low. But from time to time the upwelling is suppressed (Figure 15), sea temperatures rise 2-3°C in the central equatorial Pacific, and when the surface waters are warmer than the overlying air, instability occurs with consequent heavy rainfalls. Ocean productivity is severely reduced (Barber and Chavez 1983, 1986). During these periods seabirds populations suffer catastrophic reductions during the major 1982-1983 event all 18 breeding species of seabirds at Kiritimati, Line Islands suffered reproductive failure. The population of 10-12 million Sooty Terns *Sterna fuscata* disappeared and that of 8000 Great Frigate birds *Fregata minor* was reduced to less than 100 (Schreiber and Schreiber 1984, 1989). This resulted partly from lack of food but also because growth of land vegetation reduces nesting sites for ground-nesting seabirds and makes it difficult for larger birds such as boobies to become airborne. Similar effects were documented in the Galapagos for seabirds, marine mammals, marine iguanas, penguins, flightless cormorants and landbirds (Valle et al. 1987).

The spatial extent and temporal duration of the equatorial upwelling is thus a fundamental control of island environment over an area extending from the coast of South America (where the El Niño phenomenon was first recognised) westwards along the equator to the Caroline Islands. When upwelling is marked, islands experience droughts, land vegetation is sparse, and seabird colonies abundant, at least on undisturbed islands. When upwelling is suppressed, heavy rains occur, vegetation growth accelerates, and seabird colonies are much reduced. It follows that the input of phosphate to island soils through guano deposition is likewise episodic and correlated with El Niño events (Stoddart and Scoffin 1983).

The occurrence of historical El Niño events has been catalogued by Quinn et al. (1987) since the beginning of the sixteenth century (revising the earlier listing by Quinn et al. 1978). 47 'strong' or 'very strong' events are recorded between 1525 and 1983, with a mean periodicity of 9.9 years, and 32 'moderate' events between 1806 and 1987. Between 1803 and 1987 the mean time between moderate or stronger El Niños was ca 3.8 years. Major events occurred in 1578, 1728, 1791, 1828, 1877-78, 1891, 1925-26 and 1982-83, the last being one of the strongest recorded and certainly the best documented (Rasmusson 1983; Caviedes 1984; Gill and Rasmusson 1983; Glynn, ed. 1990; Hansen 1990; Barker and Chavez 1983, 1986).

Some of the consequences of El Niño events for the terrestrial ecology of equatorial islands have already been noted. In addition there is evidence that increased sea-surface temperatures during these events lead to thermal

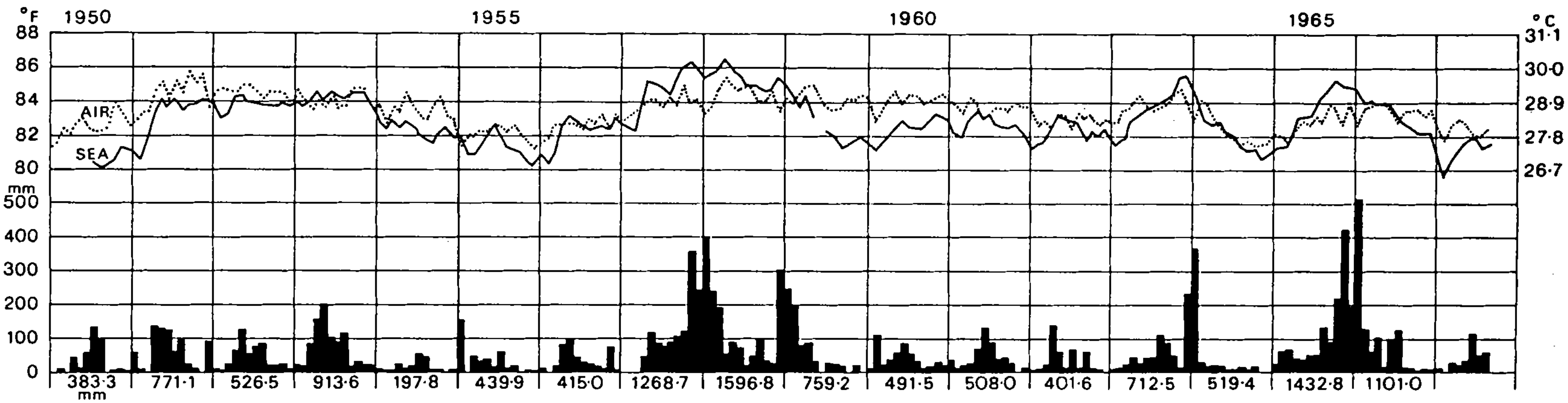

Figure 15. Relationships between monthly sea and air temperatures and monthly precipitation at Canton Island, Phoenix Islands (From Bjerknes 1969).

stress in corals and to widespread coral bleaching (Williams et al. 1987; Brown, ed. 1990; Cook et al. 1990; Glynn 1990; Rougerie 1991). Nor are El Niño effects limited to the area of Pacific equatorial upwelling: there is abundant evidence of global climatic and ecological response to these events (e.g. Duffy 1990).

Rainfall regime and seasonality

Changes in rainfall annual totals must clearly represent the sum of changes in monthly totals. Unless all monthly figures change proportionately there must also be changes in the seasonal distribution of rainfall. The nature of such changes has been analysed for Dominica, Minicoy (Lakshadweep), Viti Levu (Fiji) and Western Samoa (Figure 16). In the case of Suva, Viti Levu, changes in annual totals result mainly from changes in the rainfalls of only five months of the year, and especially in those for December, May and August; rainfall of the other seven months shows no significant changes over the period of record. In the case of Dominica, the high annual rainfalls of the wet periods results from the occurrence of a secondary maximum in November; this does not occur during dry years, when there is a single July maximum. In Western Samoa, the change from a low to a high rainfall period has resulted from increased rainfall in the wet months (October-January), i.e. the rainfall distribution became more seasonal. But the later change from wet to dry conditions involved not only a reduction in wet-season rainfall, but also an increase in dry-season amounts. At Minicoy in the Indian Ocean there was until 1958 a double peak in monthly rainfall (the main maximum in June and a secondary maximum in October), but during 1959-1974, when rainfall was about 8 per cent higher than in earlier decades, this double peak was replaced by a single maximum in July. This single peak during the wetter years also resulted in the occurrence of longer droughts in the dry season, in spite of the higher annual totals. Rainfall seasonality, which is of obvious importance in terms of geomorphic processes and vegetation growth as well as human activities, is thus related to annual totals in very complex ways. Some of the implications of changing seasonality are discussed later in this paper.

Magnitude and frequency of daily rainfalls

Of great importance, both in land erosion studies and in the ecology of nearshore areas subject to river discharge, is the frequency of high-magnitude and intensity rainfalls, which can be approached through the study of daily (or more frequent) rainfall records. Tropical islands, especially mountainous islands in trade-wind or monsoonal areas subject to hurricanes, may experience catastrophically high short-period rainfalls. Thus in 1911 Baguio on Luzon, Philippines, recorded a 24-hour rainfall of 1168 mm (3.8 feet), and in 1952 a station on Réunion, Mascarenes, recorded the world record 24-hour rainfall of 1870 mm (6.14 feet). Less extreme 24 hour maxima can also be of

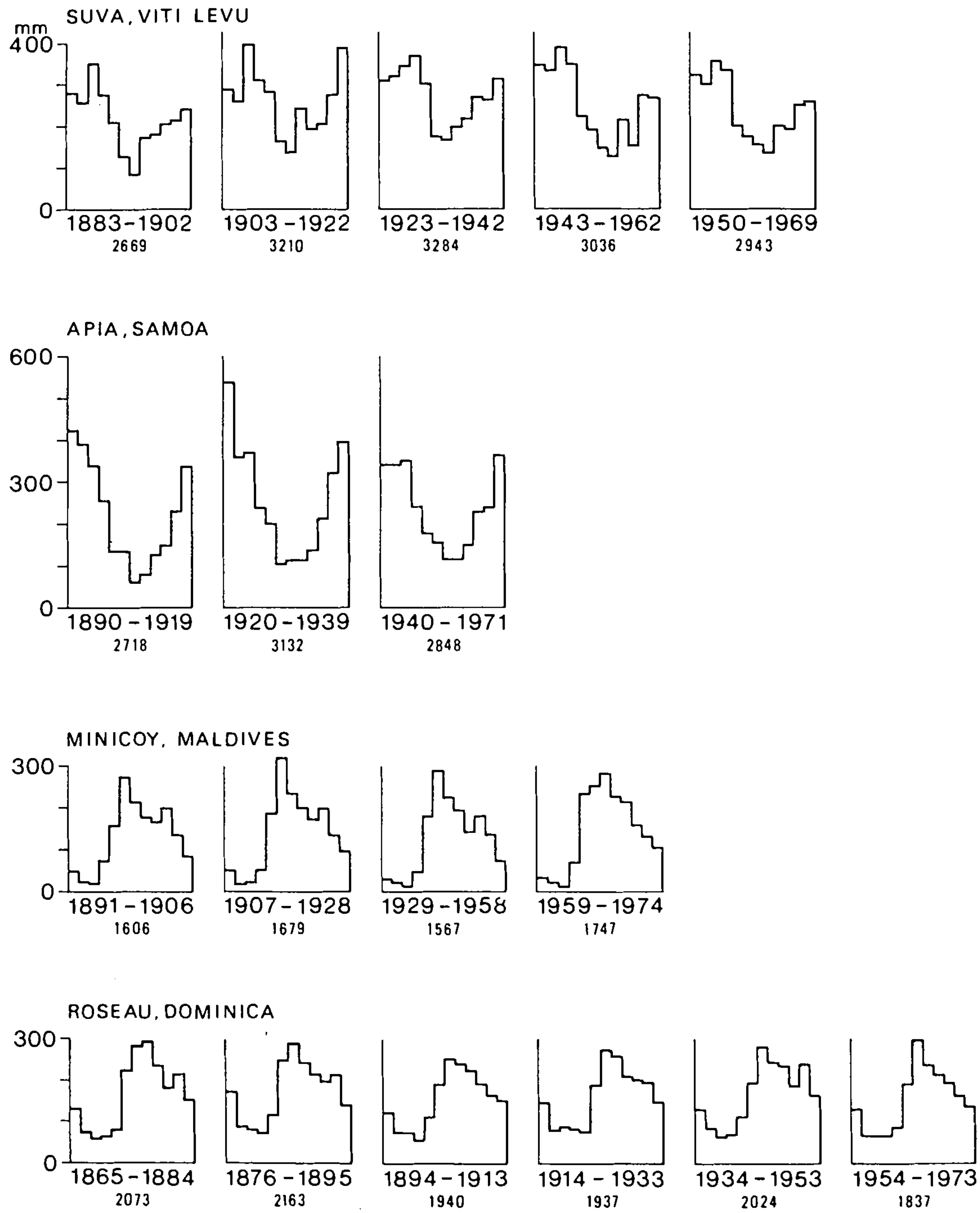

Figure 16. Changing seasonal distribution of monthly rainfall in different rainfall epochs at Suva, Viti Levu; Apia, Samoa; Minicoy, Maldives; and Roseau, Dominica. Figures beneath the histograms give annual means for each epoch.

Table 7. Frequency of high-intensity daily rainfalls in different rainfall epochs

Station	Period	Mean annual rainfall inches (mm)	Mean number of days > 1"/day	> 3"/day
St. Thomas				
BARBADOS	1889-1906	85.28 (2166)	22.0	2.3
	1907-1925	61.98 (1574)	11.7	0.6
	1926-1958	69.93 (1776)	14.5	1.1
	1962-1972	61.82 (1570)	11.6	0.7
Roseau				
DOMINICA	1921-1928	68.87 (1749)	11.7	0.8
	1929-1958	80.26 (2039)	18.9	1.7
	1959-1973	70.14 (1782)	17.2	1.1

great importance, however. The ecological and geomorphological effects of a high annual total made up of many small-magnitude rainfalls will be very different from those of a total composed of a small number of high-intensity falls. Cases where the transition from low to high annual rainfall epochs result from increases in the frequency of high-intensity rainfalls have been identified in the West Indies, and are documented in Table 7. Changed erosion rates, mass movements on slopes, increased sediment yields in rivers, sedimentation in nearshore areas, soil erosion exacerbated by deforestation, and lowered salinities in nearshore areas will all be accentuated under these conditions.

There are few analyses of rainfall intensity on atolls, though abundant data are available for study. Blumenstock and Rex (1960) showed from data collected at Enewetak Atoll, Marshall Islands, that the frequency distribution of daily rainfalls is highly skewed. Over the 13-month period August 1957-August 1958 the total rainfall at Enewetak Island in the southeast sector of the atoll was 1686 mm. Of this 871 mm, or 51.7%, fell on only 17 days, or 4.3% of the total number of days of record. The three highest daily falls were 77, 112 and 134 mm (3.04, 4.43 and 5.28 inches). Daily totals have also been studied at Aldabra Atoll, Indian Ocean (Hnatiuk 1979; Stoddart and Mole 1977; unpublished data), for the period 1968-1983, a total of 5845 days. During this period annual rainfall averaged 1103.3 mm, with extremes of 547.1 and 1467.4 mm. 22.3% of the total rainfall in this period fell on 0.79% of the total days, on days with rainfalls exceeding 50 mm (1.97 inches). Twelve days exceeded 100 mm per day (3.94 inches), and three exceeded 150 mm (5.91 inches). The three highest 24-hour rainfalls in this period were 159.3, 164.5 and 238.8 mm (6.27, 6.48 and 9.40 inches), the last on 19 April 1974. Further details are given

Table 8. Frequency of daily rainfalls at Aldabra Atoll, 1968-1983

Daily rainfall, mm	Number of days	Percent total number of raindays
0.1 - 5	1595	68.9
6 - 10	282	12.2
11 - 15	130	5.6
16 - 20	69	3.0
21 - 25	58	2.5
26 - 30	39	1.7
31- 35	44	1.9
36 - 40	18	0.8
41 - 45	19	0.8
46 - 50	9	0.4
51 - 55	11	0.5
56 - 60	6	0.3
61 - 65	3	0.1
66 - 70	6	0.3
71 - 75	3	0.1
76 - 80	2	0.1
81 - 85	3	0.1
86 - 90	2	0.1
91 - 95	1	0.04
96 - 100	1	0.04
101 - 200	13	0.6
200	1	0.04

in Table 8. There is a rather loose correlation between high annual rainfalls and the frequency of high-rainfall days. An earlier study (Stoddart and Mole 1977), over the shorter period 1968-1972, showed a better correlation between high annual rainfalls and the number of days with falls of 10-25 mm per day. This study also showed that 70% of all rain-days had less than 5 mm/day and 90% less than 15 mm/day (Stoddart and Mole 1977, 3, Table 12).

Magnitude and frequency of droughts

Mention has already been made of drought conditions in the southern Gilberts associated with the equatorial Pacific upwelling pattern (Table 6). Duration is a parameter as significant as absolute lack of rainfall in island droughts: Table 9 describes some atoll droughts characterised by both very low or even zero rainfalls and long duration. Three of the islands in Table 9 are located in the Pacific equatorial zone already described. In ecological terms drought can be defined by monthly rainfalls falling below a given threshold. Sequences of dry months, defined as months with less than 100 mm (3.94 inches) of rainfall, have been investigated: Table 10 and Figure 17 give data on

Table 9. Notable atoll droughts

Aranuka, Gilbert Islands:

29 months, August 1966–December 1968: total 146 mm (including zero rainfall February–June 1968 and August–October 1968).

Christmas Island (Pacific):

19 months, June 1949–January 1951: total 193 mm
Inlcuding 8 months, June 1949–February 1950, with a total of 22 mm

Malden Island:

14 months, January 1891–February 1892, total 220 mm
14 months, January 1895–February 1896, total 140 mm
9 months, June 1901–February 1902, total 45 mm

Alphonse Island, Amirantes:

7 months, June 1959–December 1959, total 7.6 mm

Aldabra Atoll:

6 months, June–November 1949, total 41 mm (three consecutive months with zero rainfall)

drought length so defined for several locations (Suva, Minicoy, Dominica, Apia, and Mahe) for each of the rainfall epochs previously identified. Longer drought sequences are more frequent in dry than in wet epochs, and very long droughts may occur in very dry periods. Gross fluctuations in drought frequency can also be identified over time using twenty-year moving averages, and Figure 17 also gives the twenty-year moving averages of numbers of dry months (with less than 100 mm rainfall per month) for the islands named. It is interesting to note that the frequency of extended periods of drought varies quite markedly at Minikoi even when the mean number of months of drought remains relatively invariant.

Drought duration has also been studied on a daily basis, using definitions both of no rainfall at all and of a threshold rainfall of 5 mm per day, for the five years 1968-1972 at Aldabra Atoll. The data in Figure 18 show an important asymmetry, in that wet periods are generally much shorter than dry periods, and some dry periods can be very long indeed.

An alternative approach to drought, utilising evapotranspiration measurements, has been explored in the context of agricultural potential for central Pacific islands by Nullet (1987), Nullet and Gianbucella (1988) and Gianbucella et al. (1988).

Table 10. 20 yr frequencies of length of droughts (months with less than 100 mm rainfall) in different rainfall epochs

Station	Period	Duration, months							
		1	2	3	4	5	6	7	8
Suva									
VITI LEVU	1883-1905	16.0	10.0	2.0	2.0				
	1906-1941	17.2	2.8	1.7					
	1942-1969	21.4	3.6	2.1	2.9				
Roseau									
DOMINICA	1865-1898	17.1	6.5	5.9	4.1	0.6	0.6		
	1899-1928	18.0	7.3	4.0	5.3	2.7	0.7		
	1929-1958	11.3	6.7	6.0	2.7	2.0	2.0		
	1959-1989	17.8	2.2	5.2	4.4	2.2	2.2	1.5	0.7
Minicoy									
MALDIVE IS.	1891-1906	13.8	2.5	5.0	3.8	10.0	2.5		
	1907-1928	15.5	1.8	4.6	3.6	6.4	2.7	1.8	0.9
	1929-1958	16.7	4.0	3.4	4.7	8.0	2.7	1.3	
	1959-1974	8.8	2.5	2.5	5.0	3.8	5.0	2.5	1.3
Apia									
SAMOA	1890-1919	10	7.3	3.3	2	2.6	0.6		
	1920-1939	17	6	1	3	0	1		
	1940-1971	18.1	5.6	3.8	1.3				
Mahe									
SEYCHELLES	1891-1904	17.1	4.3	7.1	4.3	1.4	1.4		
	1905-1922	15.6	8.9	10.0	1.1	1.1	1.1		
	1923-1937	22.7	9.3	4.0	1.3	1.3			
	1938-1954	18.8	3.5	3.5	5.9	3.5	2.4		
	1955-1968	17.1	5.7	7.1	2.9	1.4			
	1969-1989	12.4	7.6	1.9	4.8	3.8	1.9	1.0	

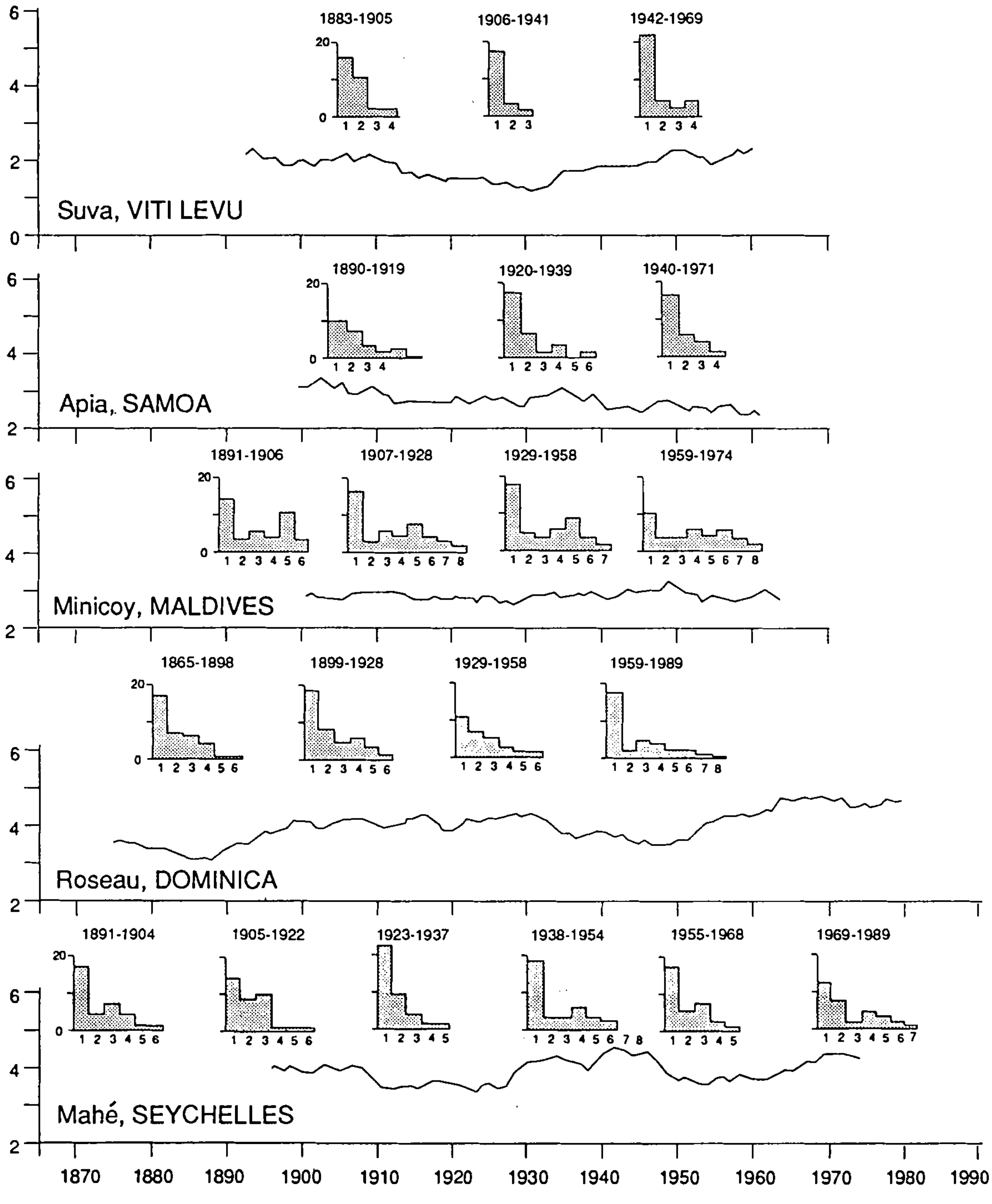

Figure 17. Twenty-year running means of numbers of dry months and frequency distribution (per 20 years) of lengths of dry periods for different time periods at Suva, Viti Levu; Apia, Samoa; Minicoy, Maldives; Roseau, Dominica; and Mahe, Seychelles.

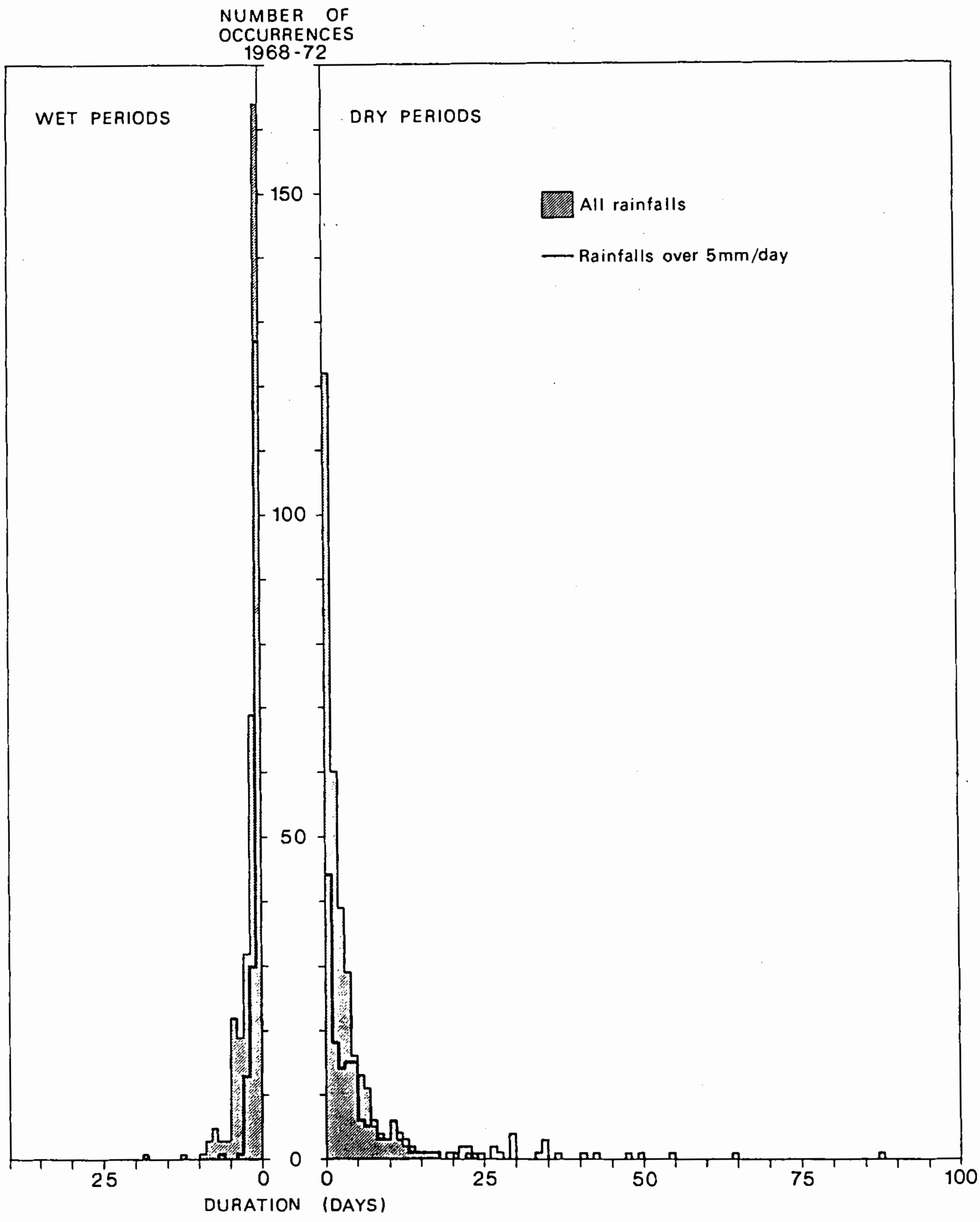

Figure 18. Frequency and duration of wet and dry spells at Aldabra Atoll, western Indian Ocean, 1968-1972 (From Stoddart and Mole 1977).

HURRICANES

Tropical hurricanes, defined as low pressure systems with wind speeds in excess of 120 km/h, are probably the single most important catastrophic event regularly experienced in the reef seas. Individual storm systems, moving at 15-25 km/h, have a mean area of 250,000 km^2 and diameter of 500-600 km; winds circulate round them in an anticlockwise direction in the northern hemisphere and clockwise in the southern. Figure 19 shows the general distribution of the hurricane belts, though it is clear from geomorphological evidence that hurricanes have occurred in the recent past in areas where there is no historical record of them.

Hurricanes act in a variety of ways (Stoddart 1971b, Dupon 1987). Wind speeds which may exceed 275 km/h (150 kts) can cause massive damage to forests and to economic crops such as coconuts and bananas, and also damage to houses, other installations such as fish traps and pearl farms, and equipment such as boats. Wind-generated storm surges can raise the local level of the sea 5 m or more above its tidally-predicted position and cause overtopping of islands and widespread inundation, especially in areas adjacent to coastal shelves. Thus Sally caused a 5 meter storm surge on Rarotonga in January 1987. These effects are particularly serious in microtidal reef areas where island altitudes are low. Wind-driven waves can lead to severe reef destruction, mobilisation of coarse sediment and its deposition on reef flats, and erosion of existing shorelines and land surfaces (waves 23.6 m high were measured during Hurricane Camille in the Gulf of Mexico in 1969: Earle 1975): reef blocks 4-6 meters in greatest dimension were deposited on reef fleets, on Raroia Atoll in 1903 and Rangiroa in 1906. Rainfall, especially in near-stationary storms, can reach extraordinary levels. There is a large literature on the economic (Weaver 1968) and social (Lessa 1964, Yamashita 1965, Barker and Miller 1990) consequences of such storms on small islands, where they may often serve as catalysts in accelerating change. There is, too, a large literature on their geomorphological and ecological effects, both terrestrial and marine (Stoddart 1971b).

Earlier studies, such as those of the effects of Ophelia at Jaluit Atoll, Marshall Islands, in 1958 (Blumenstock et al. 1961) and Hattie on the Belize reefs and cays in 1961 (Stoddart 1963) stressed the destructive effects on island morphology and vegetation of such intense storms, though it was clear from these studies that such effects varied both spatially in any particular storm and also with storm intensity. Bebe at Funafuti Atoll, Tuvalu, in 1972 demonstrated how storms impacting linear atoll islands (motus) may lead to land accretion rather than shoreline erosion. This storm had maximum winds in excess of 180 km/h and a surge up to 4 m above mean high water level. It formed a rubble ridge on the southeastern seaward reef flat 18-19 km long, 30-40 m wide and 3.5 m high; the volume of this ridge was estimated at 1.4×10^6 m^3 and its mass as 2.8×10^6 metric tonnes; the largest boulder included in it had a diameter of 7 m (Maragos et al. 1973; Baines et al. 1974).

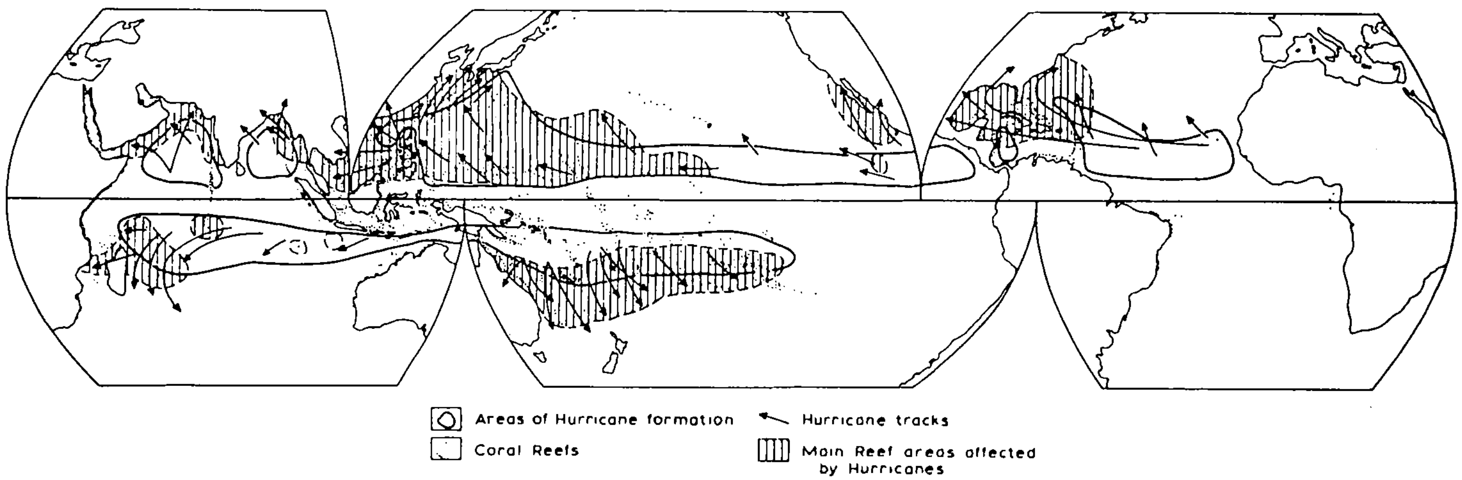

Figure 19. Distribution of main hurricane areas (From Stoddart 1971b).

Over succeeding years the ridge migrated shoreward until it became attached to the seaward beach of the motu (Baines and McLean 1976). A comparable storm on Ontong Java Atoll, Solomon Islands, in 1967 created a similar reef-flat rubble ridge 35 km long, 20 m wide, with a crest 1-3 m above mean sea-level, and this too migrated shoreward to become attached to motu shorelines over the next 19 years (Bayliss-Smith 1988). Bayliss-Smith concluded from the latter case that storms of such magnitude may be destructive on small islands (cays), but act as a source of sediment supply and episodic accretion on larger islands (motus), and also suggested that these effects varied systematically with comparable storms over the length of the Holocene (Bayliss-Smith 1988, 388-390). Woodley and co-workers (1981), however, also demonstrated the spatial variability of the effects of individual storms in terms of depth and aspect in the case of Hurricane Allen on the Jamaican reefs in 1980. This was one of the most intense storms of the century, with maximum wind speeds of 285 km/h and observed waves 12 m high in water only 15 m deep.

Extreme storms appear to have been more frequent in recent years, witness David and Gilbert in the Caribbean and Gulf of Mexico in 1979 and 1988 (Gilbert had the lowest pressure ever recorded in the western hemisphere [888 mb] and wind speeds up to 220 km/h) and Hugo in the Lesser Antilles in 1989. The six storms that occurred in the Tuamotu-Society Islands area between December 1982 and April 1983 (especially Orama), are particularly noteworthy in this respect (Figure 20). No observations on subaerial effects of these Pacific storms have been published, but reef damage was severe (Laboute 1985; Harmelin-Vivien and Laboute 1986). The frequency of central Polynesian storms appears directly linked to El Niño events, especially through increased sea-surface temperatures, and Emanuel (1987) has suggested that predicted global warming will lead to an increase in frequency and strength of major storms and an extension of the hurricane belts as sea-surface temperatures increase over future decades.

It is clear from the historical record that hurricane frequencies vary over time, even though analysis is made difficult by the paucity of accurate records in earlier years, especially in more remote locations. Milton (1974) has analysed changing frequencies in the Indian Ocean and Australian regions (Figure 21): his data show high frequencies between 1911 and 1921, low frequencies between 1921 and about 1945, and higher frequencies after that date. In the western Indian Ocean low storm frequencies in the period 1900-1929 were associated with lower sea-surface temperatures, and higher frequencies in the period 1930-1959 with higher temperatures. This temporal variability was also associated with spatial shifts: the locus of maximum activity, which was concentrated in the northern Madagascar area in the 1930s, shifted northwards in the decade 1941-1950 and southwards during 1951-1960 (Stoddart and Walsh 1979). On the Great Barrier Reef, over the period 1910-1969, storm frequencies were highest ca 1950 and lowest in the decade 1920-1930, with pronounced differences in latitudinal distribution (Coleman 1971; Stoddart 1978).

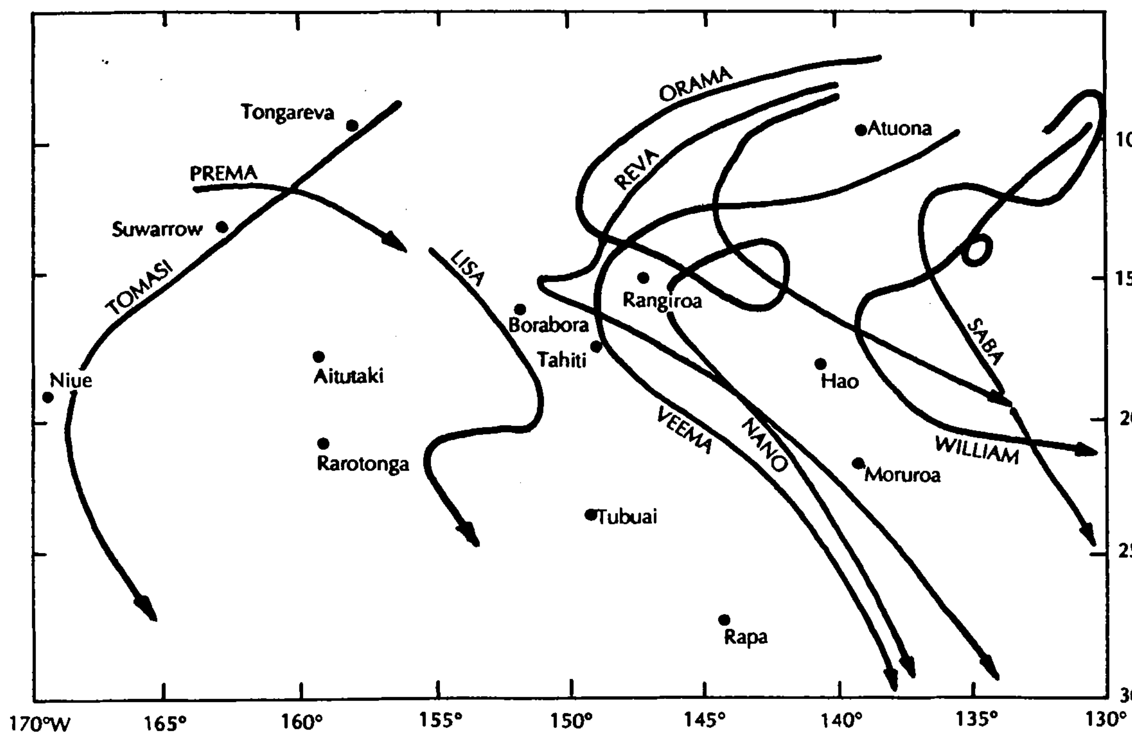

Figure 20. Hurricane tracks in the Tuamotu-Society Islands area during the period December 1982 through April 1983.

Likewise there have been marked changes in both regional frequency and intraregional spatial pattern of tropical cyclones in the North Atlantic/Caribbean, where reasonably comprehensive charted records extend back to 1871 (Neumann et al. 1978). Frequencies of tropical cyclones of at least storm intensity were high in the late nineteenth century, low in the 1910s and 1920s, very high from the 1930s to 1950. Though at the regional scale cyclone frequency has remained high in the 1960s, 1970s and 1980s, there has been a significant eastward shift in tracks into the Atlantic and frequencies over most of the Caribbean have fallen sharply (Eyre & Gray 1990, Walsh & Reading 1991). Furthermore, the frequency of cyclones of hurricane intensity has fallen in recent decades from 6.3 per annum in the 1950s to 3.5 and 4.0 per annum in the 1970s and 1980s respectively, with no indication yet of an increase with global warming.

Changes in tropical cyclone frequency over much longer timescales have been investigated for parts of the Caribbean using historical records (Walsh 1977, Walsh & Reading 1991) The particularly comprehensive records for the Lesser Antilles (Figure 22) indicate roughly equal peaks in cyclone activity in the periods 1765-92, 1804-37, 1876-1901 and 1928-58; very low frequencies from 1650-1764 at the height of the Little Ice Age, in 1793-1803 and

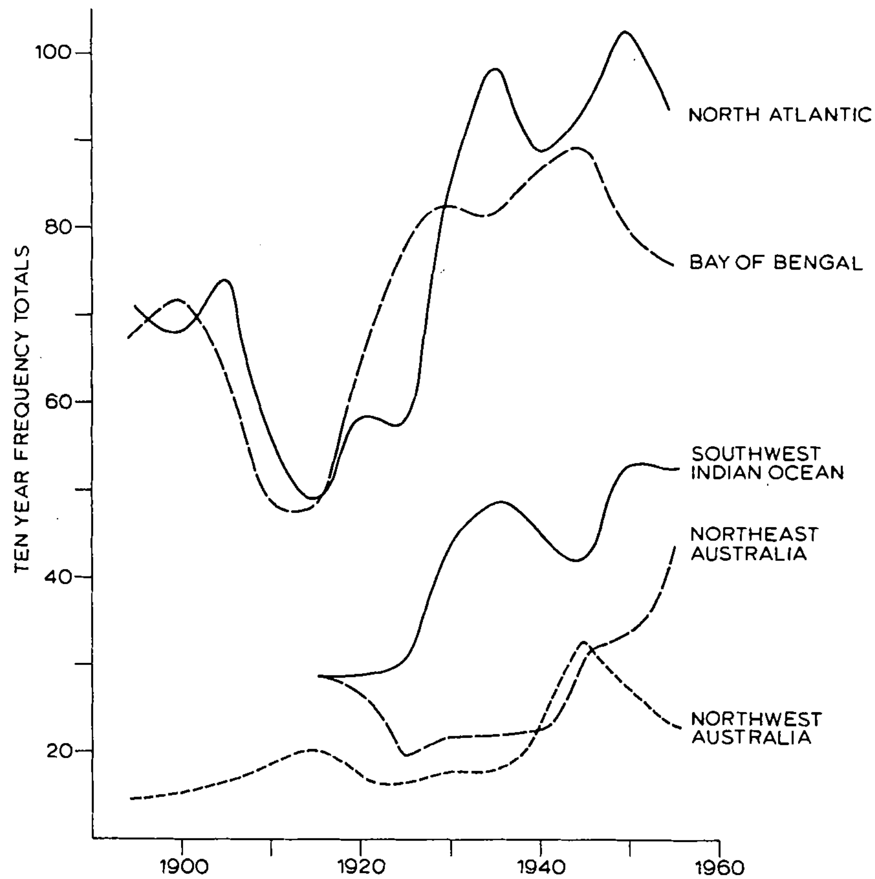

Figure 21. Ten-year frequency totals of hurricanes in different areas (From Milton 1974).

in 1835-75; and moderately low frequencies in 1902-27 and 1959-89. Even longer records for Hispaniola show a very similar temporal pattern and suggest very low frequencies during the whole of the sixteenth and seventeenth centuries. These sub-regional fluctuations can be used to reconstruct time series at a regional scale (Figure 23). Patterns of change at an individual island group level (Table 11) in the Lesser Antilles differ, reflecting latitudinal shifts in predominant tracks from epoch to epoch. Peak frequency in the Leewards/Virgins and in the French Islands and Dominica occurred in 1765-1793, but in 1876-1901 in the more southerly Windwards and in Trinidad and Tobago. For the charted period since 1871, the mean latitude at which cyclones passed westward through 61°W (which approximates to the north-south axis of the Islands) has been calculated. The mean track latitude was

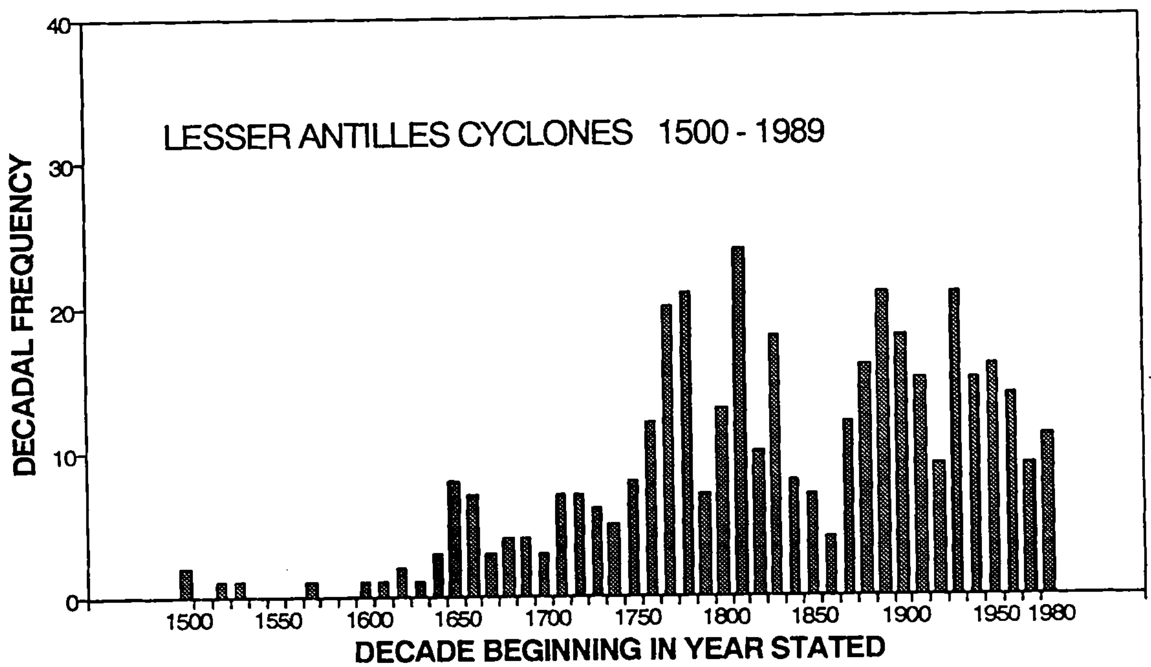

Figure 22. Ten-year frequency of cyclones in the Lesser Antilles, 1500-1989. Records are considered reasonably comprehensive since 1650. (From Walsh and Reading 1991)

21.0°N in 1871-75, more southerly at 18.4°N in 1876-1901, somewhat more northerly in both the 1902-27 (19.0°N) and 1928-58 (18.9°N) periods, but much further south at 17.7°N since 1959 .

Walsh & Reading (1991) found that the Atlantic/Caribbean appear to be more linked to shifts in key aspects of the atmospheric circulation rather than to changes in sea surface temperature or the frequency of El Niño events (as indicted by the chronology since 1500 of Quinn et al. 1987). Although strong El Niño events during the current century have been shown to be associated with a reduced frequency and changed spatial distribution of cyclones of hurricane intensity in the North Atlantic (Eyre & Gray 1990, Gray & Sheaffer 1991), it does not appear to have been the main influence on cyclones in the longer term.

IMPLICATIONS

The variations in climatic elements, notably rainfall, viewed as inputs to the terrestrial ecosystem, which we have identified, together with other perturbations, are on such a scale that they must have serious implications both for the structure of island ecosystems and for the processes at work within them.

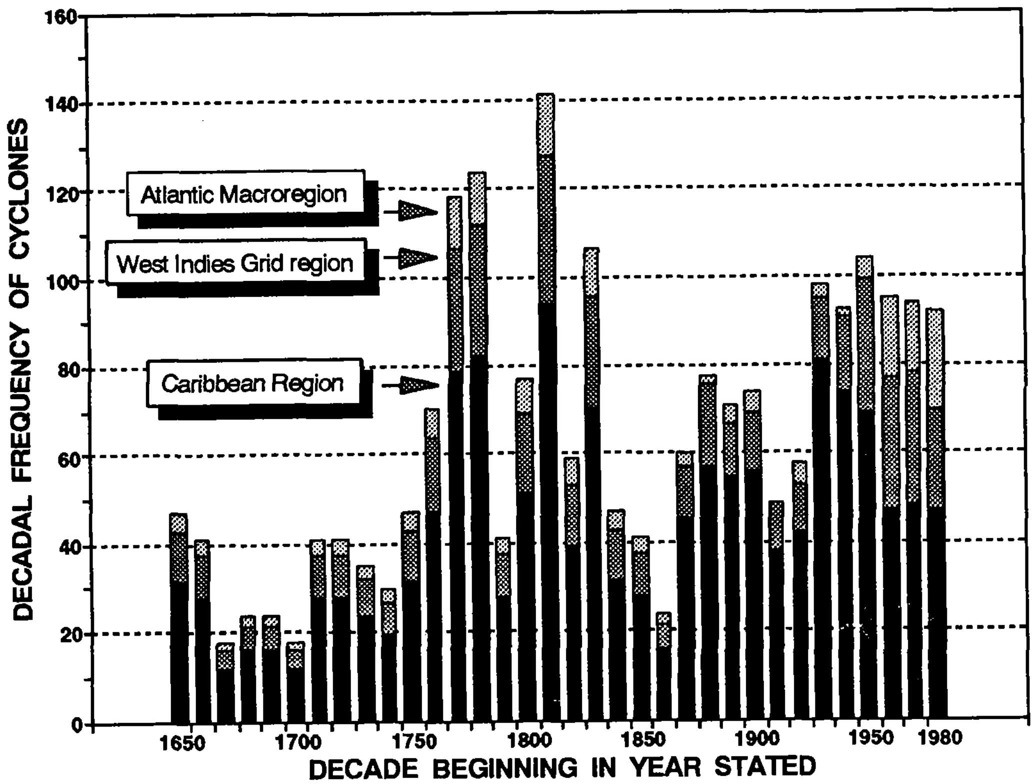

Figure 23. Reconstructed ten-year frequency of cyclones in the Atlantic, Caribbean and West Indian regions, 1650-1989. (From Walsh and Reading 1991).

Rainfall magnitudes, together with seasonality, form a basic input in the geomorphic system. Fournier (1960) has demonstrated a fairly simple relationship between suspended sediment yield in drainage basins and a parameter p^2/P, where p is the precipitation in mm of the month with highest rainfall, and P is the mean annual rainfall in mm. Fournier showed the relationship between 8 and 80 for values of p^2/P and from less than 100 to over 1500 tonnes/km^2/year for suspended sediment yield (cf. Stoddart 1969, 183). Table 12 shows how p^2/P varies between each of the identified rainfall epochs at Viti Levu and Samoa in the Pacific, Minicoy and Mahe in the Indian Ocean, and Dominica in the West Indies. The limits of variation at these stations are respectively 41.8-51.5 at Suva, 48.3-95.5 at Apia, 34.9-59.8 at Minicoy, 50.4-81.3 at Mahe, and 34.7-49.2 at Dominica. At Apia p^2/P during 1920-1939 was almost exactly twice that for 1940-1971. These figures vividly illustrate the fact that it is not possible to predict, for example, the erosional consequences of deforestation on tropical islands without understanding the rapid and large-scale fluctuations in intensity of erosional processes.

Table 11. Changes in mean annual frequency of tropical cyclones within the Lesser Antilles 1650-1989

Period	Leewards/ Virgin Is.	French Is. & Dominica	Windward Islands	Trinidad & Tobago	Lesser Antilles
1650-1764	0.28	0.22	0.17	0.01	0.56
1765-1793	1.07	0.96	0.29	0.14	1.96
1794-1805	0.27	0.00	0.00	0.09	0.36
1806-1837	0.94	0.94	0.62	0.09	1.79
1838-1875	0.42	0.21	0.24	0.03	0.66
1876-1901	0.88	0.69	0.81	0.15	1.96
1902-1927	0.54	0.62	0.62	0.00	1.27
1928-1958	0.71	0.55	0.61	0.13	1.74
1959-1989	0.42	0.45	0.42	0.10	1.13

Such variations in climate also have substantial ecological implications. We have already noted the correspondence between rainfall and vegetation in the Marshall Islands (Figure 8), as systematised in the 'Fosberg zones', originally described for the northern Marshalls and subsequently extended from Wake Island in the north southwards to Tuvalu (Fosberg 1956; Wiens 1962; Catala 1957; Amerson 1969). In Fosberg's zone 1 (Wake and Taongi) no coconuts grow. In zone 2 at Bikar there is *Pisonia* forest. In zone 3 there is *Cordia, Pemphis,* mixed forest and coconuts. In zone 4 there is *Neisosperma* forest and breadfruit; in zone 5 coconuts and breadfruit; and in zone 6 dense forest. Zones 7, 8 and 9 mirror zones 5, 4 and 3 on the south. If, as in Figure 7, these zones are defined simply in terms of mean annual rainfall, then fluctuations over time in the mean of ±25 per cent are sufficient to move an atoll from the centre of one zone to the centre of another; smaller fluctuations can take an atoll across a zonal boundary. We have seen that fluctuations of up to or exceeding 25 per cent have occurred at some tropical stations between successive rainfall epochs, and changes of 10 per cent are common.

This observation raises important questions about the environmental controls of such vegetation units, and about the rates at which vegetation (especially shrubs and trees) can respond to changes in environmental conditions. It may, for example, be hypothesized that such responses are asymmetric, with herbaceous vegetation responding more rapidly to increasing wetness than to drought, and shrubs and trees being more responsive to drought than increasing wetness. Further studies are needed of the relationships between such parameters as rainfall periodicity and

Table 12. Variation in p^2/P in different rainfall epochs for tropical island stations

Period	Mean Ann. Rainfall mm	Highest Mon. Mean mm	p^2/P
SUVA, FIJI			
1883-1902	2682	361 (Mar)	48.7
1903-1922	3210	406 (Mar)	51.5
1923-1942	3284	371 (Apr)	41.8
1943-1962	3036	390 (Mar)	50.2
1950-1969	2943	360 (Mar)	44.1
APIA, SAMOA			
1891-1919	2718	427 (Jan)	67.1
1920-1939	3132	547 (Jan)	95.5
1940-1971	2870	371 (Dec)	48.3
ROSEAU, DOMINICA			
1865-1884	2073	298 (Aug)	43.0
1876-1895	2163	296 (July)	40.6
1894-1913	1940	259 (July)	34.7
1914-1933	1937	274 (July)	38.8
1934-1953	2024	284 (July)	39.9
1954-1973	1837	300 (July)	49.2
MINICOY, MALDIVE ISLANDS			
1891-1906	1606	274 (June)	46.7
1907-1928	1679	317 (June)	59.8
1929-1958	1567	292 (June)	54.4
1959-1974	1747	283 (June)	34.9
MAHE, SEYCHELLES			
1891-1904	2600	460 (Jan)	81.3
1905-1922	2192	388 (Jan)	68.7
1923-1937	2652	390 (Jan)	57.3
1938-1954	2038	320 (Jan)	50.4
1955-1968	2585	451 (Jan)	78.6
1969-1989	2082	358 (Jan)	61.7

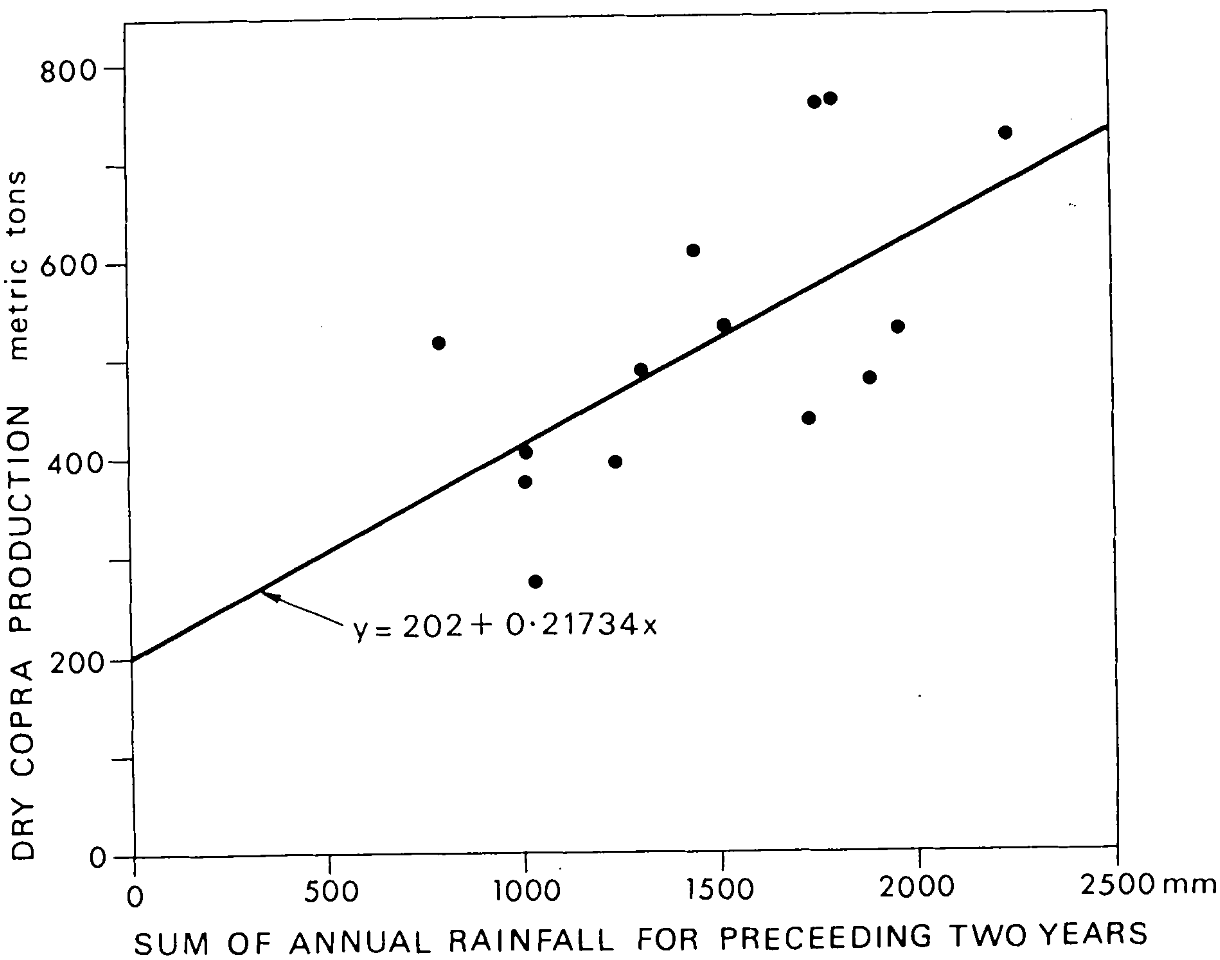

Figure 24. Relationship between rainfall and copra production at Kiritimati [Christmas Island], Line Islands (From Jenkin and Foale 1968).

hurricane magnitude and frequency and the life-cycles of island plants. And as demonstrated by the response of the Christmas Island seabird populations to the 1982-1983 El Niño event, perturbations in climate and vegetation are transmitted through other components of the island ecosystem.

The data also suggest substantial constraints on agricultural activities. Jenkin and Foale (1968) have shown a relationship between copra production and the aggregate rainfall of the two previous years at Kiritimati (Line Islands) (Figure 24), and in Trinidad Smith (1966) has shown that there is an even closer relationship between copra yield and the rainfall of the dry season of the two previous years (Figure 25). Such relationships have been observed at many tropical stations (e.g. Patel and Anandan 1936), not only for copra but also for sugar (Smith 1962; Chang et al. 1963; Oguntoyinbo 1966). There is

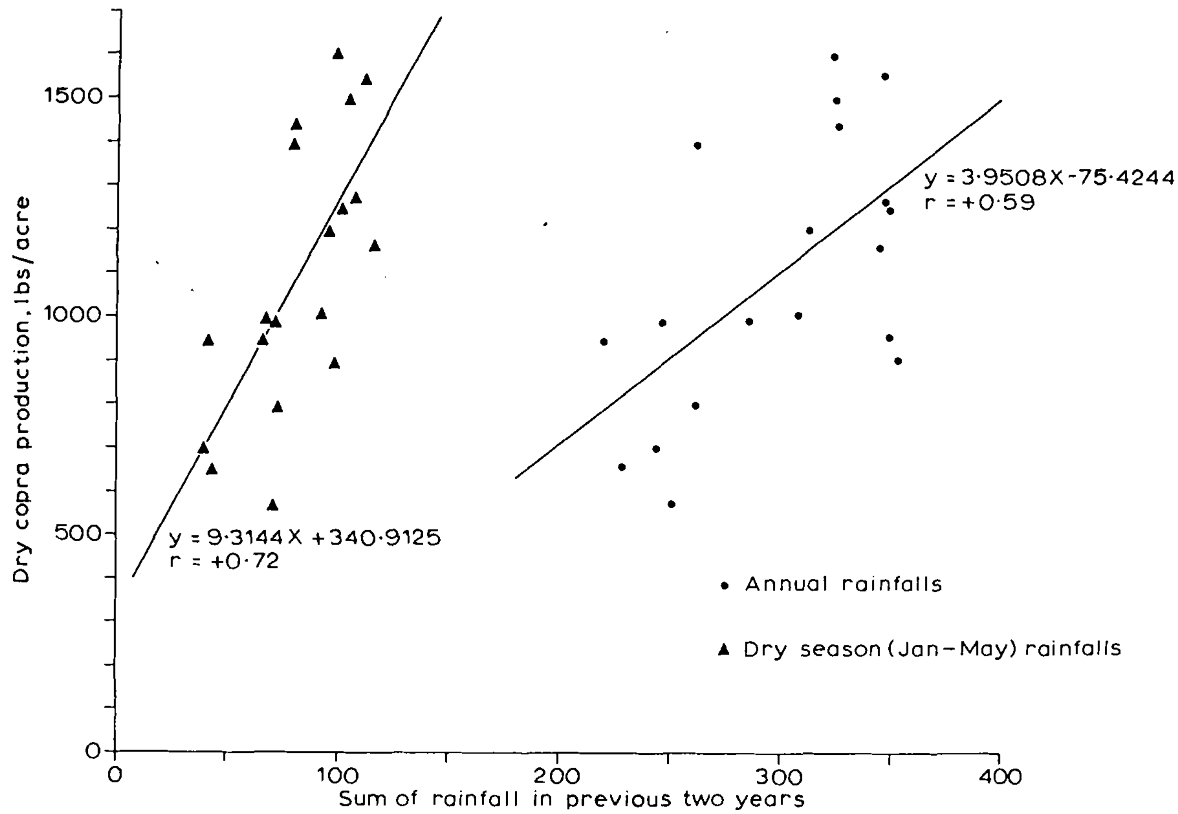

Figure 25. Relationship between annual and dry season rainfall and copra production, Perseverance Estate, Trinidad, 1941-1959 (Data from Smith 1966).

little doubt that, as Marshall (1956) suggested, variations in rainfall on the scale now identified must have had profound consequences for human activities and especially for agriculture.

Finally, it is worth noting that the effects of these climatic variations must extend throughout the terrestrial ecosystems of islands. Just as the Fosberg zones demonstrate a close linkage between mean annual rainfall and vegetation, so Amerson (1969) has shown an equally close linkage between rainfall and bird species diversity (Figure 26). At Aldabra Atoll D . W. Frith (1979) has demonstrated how rainfall seasonality controls insect abundance, and C. B. Frith (1976) has shown how insect numbers affect breeding performance in predominantly insectivorous landbirds. Likewise at the same location, Coe et al . (1979) have used an equation developed by Rosenzweig (1968) to calculate net primary production under different rainfall conditions. They have shown how food consumption by the Giant Tortoise *Geochelone gigantea* varies seasonally from 380±64.8 g dry mass/day to only 110±77.9 g in the late dry season. On a longer time scale they suggest that secondary

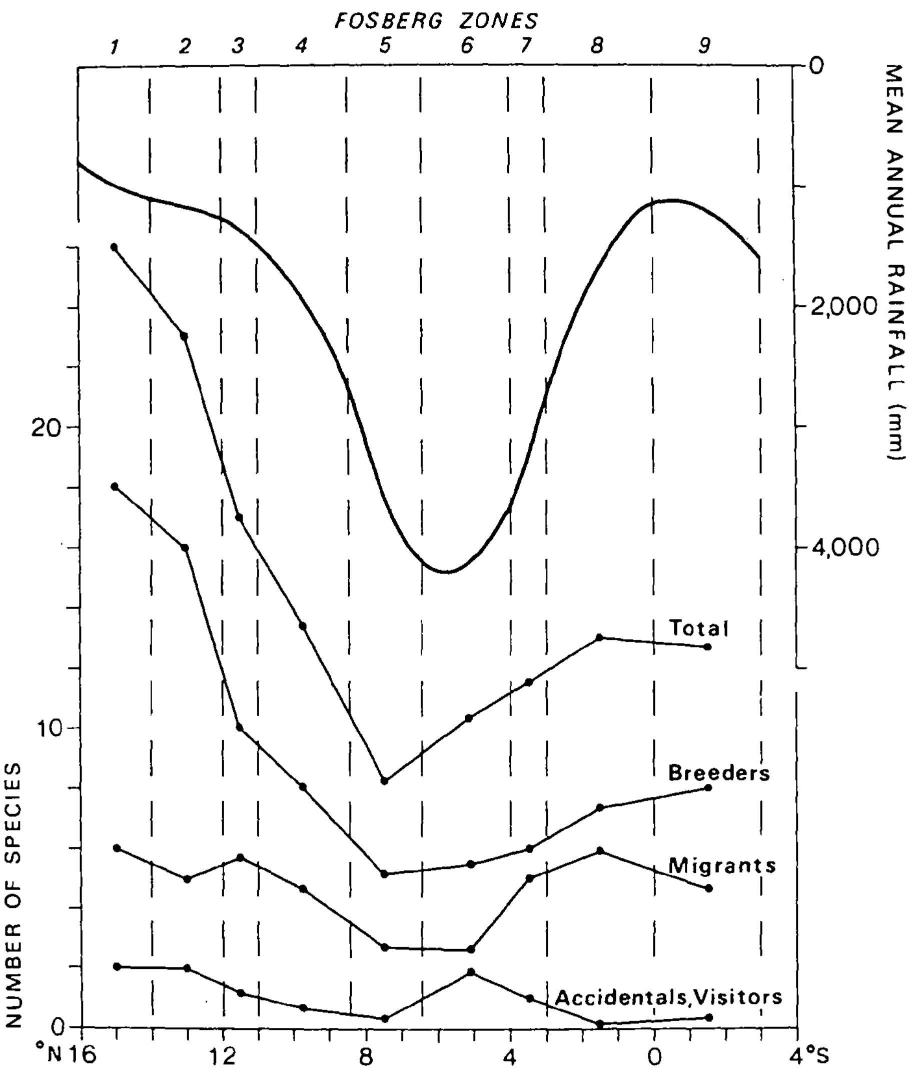

Figure 26. Relationship between mean annual rainfall, latitude, and numbers of birds species for the Marshall and Gilbert Islands (Data from Amerson 1969).

production is itself controlled by primary production, ranging for tortoises from 624 kg/km^2/year in a dry year (547 mm annual rainfall) to 3245 kg/km^2/year in a wet year (1487 mm); production is itself a key to biomass. Climate and its variability is thus seen as a fundamental control of the functioning of the terrestrial ecosystem of Aldabra Atoll. We may go further and speculate that substantial variations in rainfall, as well as catastrophic damage by hurricanes, may over quite short periods of time, measured in decades, have implications for changes in species diversity and even for the extinction or survival of individual taxa on islands.

CONCLUSION

We have therefore shown that island environments are highly variable in a number of critical ways, on time scales ranging from the last few thousand years to very short period fluctuations. We suggest that analyses of human adaptability on tropical islands should begin with a presumption of temporal variability in critical environmental inputs, rather than with a reliance on long-term mean conditions. We have suggested that such changes may have important implications for the economic viability of island populations, through effects on crop yields, and we have shown that in some cases, at least, such as the Phoenix Islands, environmental variations in recent years have exceeded even the thresholds of survival for indigenous human communities. In these responses human populations mirror those of other components, both plant and animal, of the terrestrial island ecosystem.

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