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Chapter 2

Observed and projected changes in surface climate of the tropical Pacific

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> *'Warming of the climate system is unequivocal, as is now evident from observations of increases in global average air and ocean temperatures, widespread melting of snow and ice, and rising global average sea level.*
>
> *Most of the observed increase in global average temperatures since the mid-20th century is very likely due to the observed increase in anthropogenic greenhouse gas concentrations.'* (IPCC 2007)[i]

i IPCC (2007) Summary for Policymakers. In: S Solomon, D Qin, M Manning, Z Chen, M Marquis, KB Averyt, M Tignor and HL Miller (eds) *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge University Press, Cambridge, United Kingdom, and New York, United States of America.

Contents

Page

2.1 Introduction

Weather is defined as the relatively instantaneous state of the atmosphere as described from day-to-day by measurable variables, such as air temperature, rainfall, wind speed and direction, cloud cover and humidity. Climate is the long-term average weather – what is expected at a particular time and place – and is based on observations over many years. Descriptions of climate can include both the average values and measures of variability from year-to-year. A climate change is then defined as a significant change in what we expect the weather to be like at a particular location and season[1]. Such a change could be in average values and/or in the variability around the average, i.e. the range of extremes. Determining the nature and significance of changes in climate globally or regionally is dependent on long, homogeneous weather observations from as many locations as possible[2].

Humans, and the natural and managed ecosystems that we rely on for the goods and services they provide, are adapted to prevailing climatic conditions. Global and regional climate has varied in the past on a range of time scales. For example, orbital forcing of the climate system over the past 450,000 years resulted in major swings in global climate between four cooler periods of widespread glaciation (ice ages), interspersed with warmer interglacial periods lasting from 10,000 to 30,000 years. Warming of the global climate by 4–7°C since the last glacial maximum about 21,000 years ago is estimated to have occurred at a rate 10 times slower than the observed warming during the 20th century[3]. Climate also varies on interannual to decadal time scales due to a range of forcings that are internal (such as El Niño-Southern Oscillation and North Atlantic Oscillation) or external (such as volcanic aerosols, amount of incoming solar radiation) to the climate system[4]. We have, however, entered a new era of rapidly changing global climate as a consequence of human activities.

'All these changes characterise a carbon cycle that is generating stronger climate forcing and sooner than expected.' (Global Carbon Project 2008)[5]

The evidence for increasing greenhouse gases due to human activities since the late 18th century is unequivocal[6–8]. The atmospheric concentration of the main greenhouse gas, carbon dioxide (CO_2), rose from 280 parts per million (ppm) in 1750 to 385 ppm in 2008[9], a 38% increase and the highest concentration of the last 800,000 years[10] and possibly the last 20 million years. The concentrations of other greenhouse gases such as methane and nitrous oxide have also risen over this period. Not only are atmospheric concentrations of greenhouse gases rising but the rate of increase is accelerating (**Figure 2.1a**). These concentrations are projected to continue to rise over the 21st century (**Figure 2.1b**). The annual mean growth rate of the main greenhouse gas (CO_2) was 2.0 ppm per year for 2000–2007 compared to an average annual growth rate of 1.5 ppm per year from 1990 to 1999[5,11,12].

This increase in atmospheric greenhouse gases is causing significant positive radiative forcing of the global climate system[8], i.e. global warming (**Figure 2.2**). The most recent assessment of the Intergovernmental Panel on Climate Change, the Fourth Assessment Report (IPCC-AR4), indicates that average global land and ocean surface temperatures (based on instrumental observations) have warmed by 0.74°C over the last 100 years (1906–2005) and that the rate of warming over the past 50 years (0.13°C per decade) is almost twice that of the past 100 years (0.07°C per decade)[13]. Assessments since IPCC-AR4 highlight that the pace of climate change is accelerating and is tracking the upper end of the more pessimistic scenarios for the 21st century[14,15].

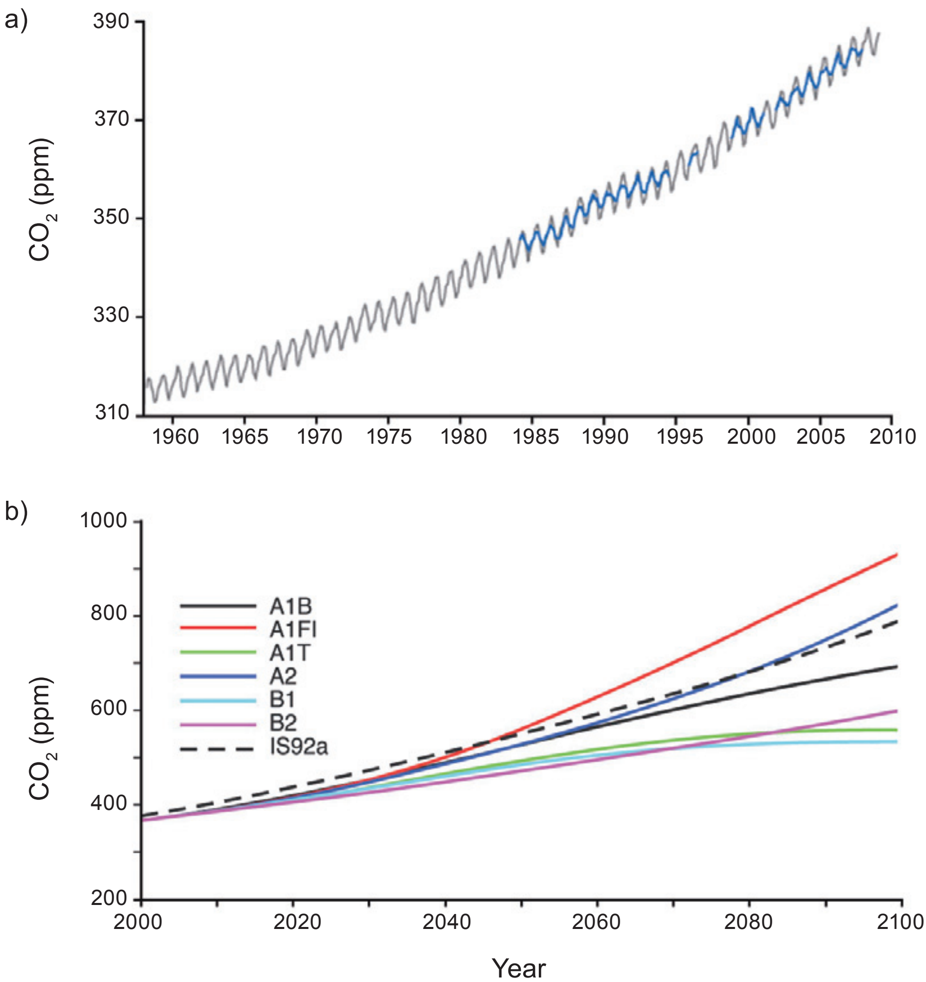


Figure 2.1 (a) Monthly atmospheric carbon dioxide (CO_2) concentration (ppm) for Mauna Loa, Hawaii (March 1958 – February 2009) (grey) and Christmas Island, Kiribati (March 1984 – December 2007) (blue), illustrating the increase, and accelerating rate of increase, of this greenhouse gas in the northern and equatorial central Pacific (source: World Data Centre for Greenhouse Gases)[93]; (b) Projected concentrations of atmospheric carbon dioxide (CO_2) concentrations to 2100 for six IPCC SRES scenarios (source: Meehl et al. 2007)[71].

The relatively modest global warming observed to date has already been associated with changes in the global climate system, such as more intense rainfall, more frequent droughts, sea-level rise, loss of Arctic sea ice, melting of land-based ice[13] and a widening of the tropical climate belt[16]. The rate of warming is also about twice as fast for land masses compared to oceans and, in the Northern Hemisphere, for high versus low latitude regions. Warming of the ocean varies between basins, with that of the Pacific being 'punctuated' by El Niño events and Pacific decadal variability compared to the steadier observed warming of, for example, the Indian Ocean[13]. Many of the instrumental observations of recent climate changes are, however, centred on temperate land areas with relatively little detail for tropical islands and ocean regions, which are the focus of this publication.

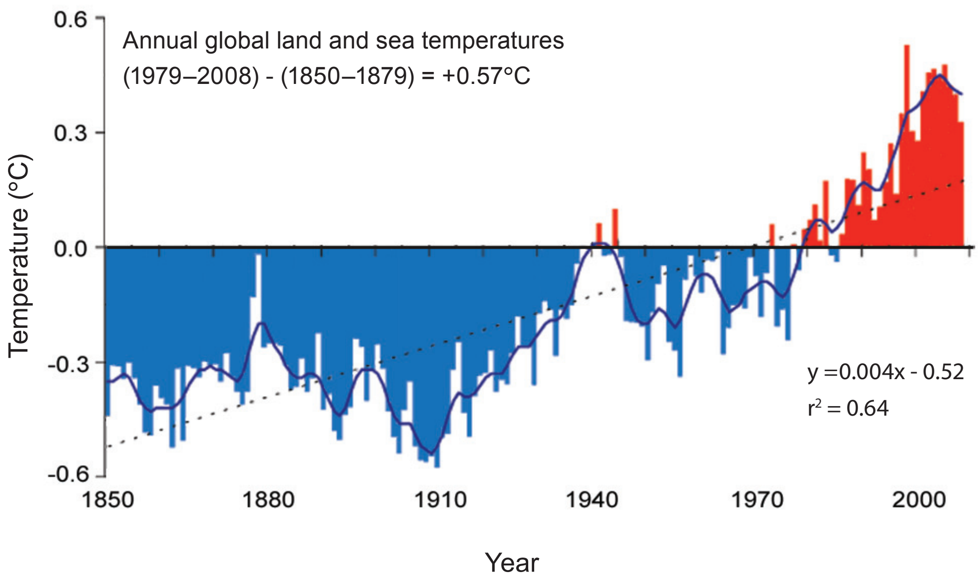


Figure 2.2 Annual global land and sea temperature anomalies (from 1961–1990 mean) for the period 1850–2008, illustrating the observed warming of global temperatures and the occurrence of the warmest years in the observational record in recent years. Thick line is 10-year Gaussian filter emphasising decadal variability, and dashed line is linear trend (source: HadCRUTV3, Jones et al. 1999, Brohan et al. 2006, Rayner et al. 2003, 2006)[94–98].

Assessing the nature, magnitude and significance of recent climate variations and trends relies on high-quality observational records from a range of locations. In particular, there is a paucity of reliable precipitation measurements for tropical oceanic regions, and discrepancies among observational and remotely sensed products make assessing changes in precipitation in these regions difficult. Despite some data limitations, the observed changes in climate are, however, driving changes in the world's biological and physical systems that are consistent with a rapidly warming climate[17,18].

The Pacific Island countries and territories (PICTs) are scattered across the large area of the tropical Pacific Ocean. These relatively small land areas encompass diverse environments, from high-elevation volcanic islands to low-lying coral atolls (Chapter 1). Although the magnitude of warming is, and is expected to be, greater over land areas than the oceans, and greater at higher compared to lower latitudes, significant changes are also occurring in the tropical Pacific Ocean[ii] (**Figure 2.3**). In this chapter, which builds on earlier assessments[19,20], we first review some of the uncertainties in accurately projecting future climate changes in the tropical Pacific, then provide a brief description of the current surface climate of the region and the major controls on seasonal and interannual climate variability. We then present examples of how some components of the recent surface climate are already changing with the global warming observed to date. Finally, we describe how surface climate is likely to change in coming decades for two scenarios – low emissions and high emissions. The specific focus of this chapter is to help provide the basis for assessing the vulnerability of fisheries and aquaculture in the Pacific to the regional consequences of global climate change.

2.2 Understanding how climate will change – uncertainties

'*Many aspects of tropical climatic responses remain uncertain*.' (Christensen et al. 2007)[21]

A number of factors must be taken into consideration to understand and document the potential consequences and impacts of a rapidly changing climate. First, we need high-quality observations of modern climate conditions that are sufficiently detailed to determine, for example, the climatic envelope of particular organisms. Such records need to be long, continuous and ongoing to also detect changes in current conditions. Unfortunately, such high-quality data for the vast area of the tropical Pacific and the scattered and isolated islands[2] are relatively scarce compared, for example, to records for temperate Northern Hemisphere land areas[iii].

Secondly, we need sufficient understanding of the complex physics of the global climate system, and the various interactions between the atmosphere, ocean, land, cryosphere and biota, to realistically model current climate conditions. Such global climate models then provide the basis for projecting future changes as a consequence of radiative forcing by greenhouse gases. The present generation of global climate models still has difficulties in correctly simulating certain components of the present day, tropical western Pacific climate[22,23]. Present climate models are also fairly coarse (~ 150 km grid size) in their spatial resolution, which makes projecting to the regional scales most relevant to human populations a challenge[21,24]. Current models, for example, do not differentiate the small PICTs from the ocean, and may not adequately incorporate some of the ocean-atmosphere interactions of importance to island climates. Although there are various dynamic or statistical methods to

ii Defined here as the area 25°N to 25°S and 130°E to 130°W.

iii The Pacific Climate Change Science Program has recently undertaken recovery and homogenisation of historical climate data for many Pacific Island countries and made them available at www.bom.gov.au/climate/pccsp

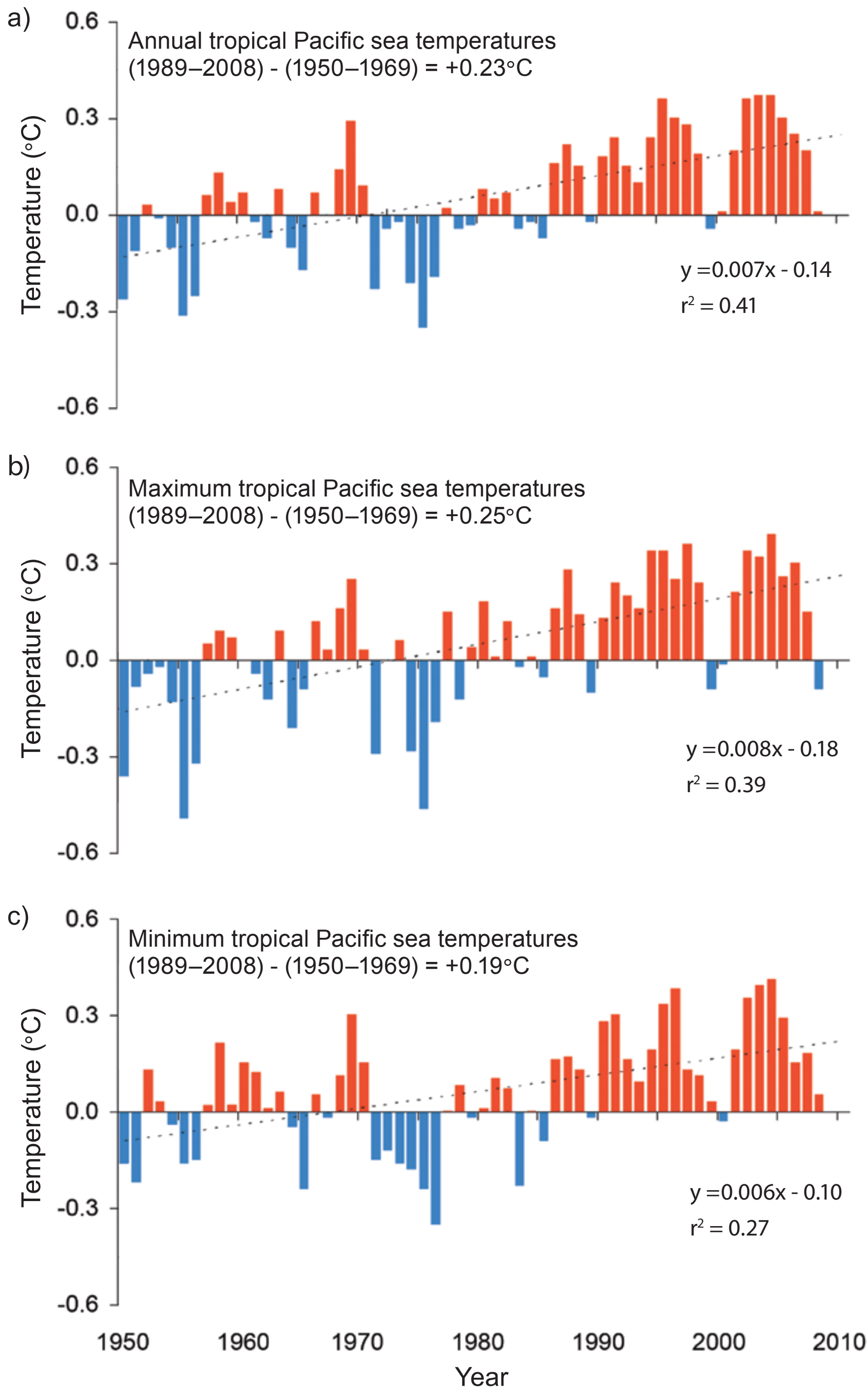


Figure 2.3 Anomalies for tropical Pacific sea surface temperatures, from the 1961–1990 mean, for the period 1950–2008 for (a) annual average temperatures, (b) annual maximum temperatures and (c) annual minimum temperatures, illustrating the recent acceleration in the rate of global warming, and that significant warming (~ 50% of global average) is already occurring in the tropical Pacific. Dashed line is linear trend (source: HadCRUTV3, Jones et al. 1999, Brohan et al. 2006, Rayner et al. 2003, 2006, Climate Research Unit)[94–99].

'downscale' coarse-resolution model outputs to finer spatial scales, these rely on a high degree of realism in the 'coarse' model and corrections for bias in SST. There is a need to produce more reliable models with higher spatial resolution, especially for the tropics[25].

Thirdly, although based on the same physical laws, different global climate models vary in the way in which they handle or 'parameterise' key small-scale processes that are unresolved by the models. This can lead to slightly different results for both present and future climate simulations. Such parameterisations (e.g. cloud formation or ocean eddy mixing) rely on a thorough understanding of the physical processes involved and are a vital component in climate modelling as they keep computational costs down to manageable levels. There is, therefore, no single 'perfect' global climate model and the most recent IPCC-AR4 often presents a multi-model average across a large number of relatively independent climate projections to account for the inter-model variability. This averaging tends to remove opposing biases in the models and is considered a suitable method for obtaining useful output, though it can also remove climate extremes that may be real[26] (Chapter 1).

Finally, projecting future climates also depends on projecting future greenhouse gas concentrations (i.e. the link to radiative forcing). The latter depends on a variety of socio-economic factors – basically the global response and level of commitment to reducing and stabilising greenhouse gas emissions in the atmosphere (mitigation), which in turn will affect the magnitude and timing of future climate changes. To take account of the uncertainty in future human emissions, the IPCC constructed a number of plausible scenarios, ranging from very carbon-intensive futures with high emission rates, to scenarios where emissions are 'reined in' very quickly. These scenarios specify the concentrations of greenhouse gases used as inputs for the climate models. Here, we use two scenarios from the Special Report on Emissions Scenarios (SRES)[27], which were used widely for the IPCC-AR4:

1. the relatively low emissions SRES B1 scenario (hereafter B1) which assumes that the concentration of the main greenhouse gas, CO_2, will be 500–600 ppm by the end of the 21st century, and
2. the most commonly modeled higher SRES A2 scenario (hereafter A2) with CO_2 concentrations reaching ~ 750–800 ppm by the end of the 21st century (**Table 2.1**).

Table 2.1 Projected global air and sea temperature changes, sea-level rise (relative to 1980–1999) and carbon dioxide (CO_2) concentrations for the low emissions B1 and high emissions A2 scenarios for 2100 (source: Bindoff et al. 2007, Meehl et al. 2007)[71,112]. In the shorter term, values for both scenarios are very similar, with projected warming by 2035 of ~ 0.9°C, sea-level rise of ~ 0.10 m and atmospheric CO_2 concentrations of ~ 400 ppm.

Scenario	Temperature (°C)	Sea-level rise (m)*	CO_2 (ppm)
B1	+1.8 (1.1–2.9)	0.18–0.38	500–600
A2	+3.4 (2.0–5.4)	0.23–0.51	750–800

* See Chapter 3 for updated estimates of sea-level rise.

Projections are presented for the near-term, 2026–2035 (hereafter 2035), and long-term, 2090–2099 (hereafter 2100)[iv]. Many scientists consider emissions reductions well below those of the B1 scenario are necessary to avoid dangerous climate change and significant ecosystem impacts[28,29]. Unfortunately, we are currently tracking above the high emissions A2 scenario[11,30].

There are, therefore, a range of uncertainties in projecting the nature, magnitude and consequences of how the surface climate of PICTs will change in the future. This 'explosion of uncertainty'[31] needs to be considered when attempting to delineate the range of potential impacts of a rapidly changing climate on marine ecosystems of the tropical Pacific. The planet, however, is already committed to ongoing, rapid, possibly intensifying climate changes for the foreseeable future. Climate will continue to change due to gases already in the atmosphere, even with drastic mitigation strategies[32,33], and there is the spectre of irreversible changes on the scale of thousands of years[34]. Our challenge in confronting a rapidly changing climate is to avoid the unmanageable, and manage and be prepared for the unavoidable[35].

2.3 Present-day surface climate

2.3.1 Atmospheric circulation

Surface climates of islands in the tropical Pacific are dominated by the vast surrounding ocean and the large-scale atmospheric and oceanic circulations[36–38]. The major atmospheric circulation features (**Figure 2.4**) include the northeast and southeast trade wind regimes, which originate in the subtropical high-pressure belts of each hemisphere where air sinks and dries. These tropical easterly flows are characterised by their constancy in speed and direction, although they tend to be strongest in the winter season of each hemisphere and to extend further polewards in the summer seasons.

The trade winds from the two hemispheres converge in the Intertropical and South Pacific convergence zones (ITCZ and SPCZ), where rising air forms the ascending branch of the Hadley circulation. The Hadley circulation represents the main north to south component of the Pacific atmospheric circulation. In addition, the Walker circulation operates in the east to west plane of the tropical Pacific with, normally, rising air over Indonesia and sinking air in the southeast tropical Pacific. This circulation is intimately linked to the major source of interannual tropical climate variability, El Niño-Southern Oscillation.

iv Examination of 10-year periods is based on multi-model averages (see Chapter 3, which uses 20-year averages but a smaller subset of models). The two approaches, which cancel out differences among models and use 10–20 year averages, are likely to produce relatively stable scenarios[26].

a)

b)

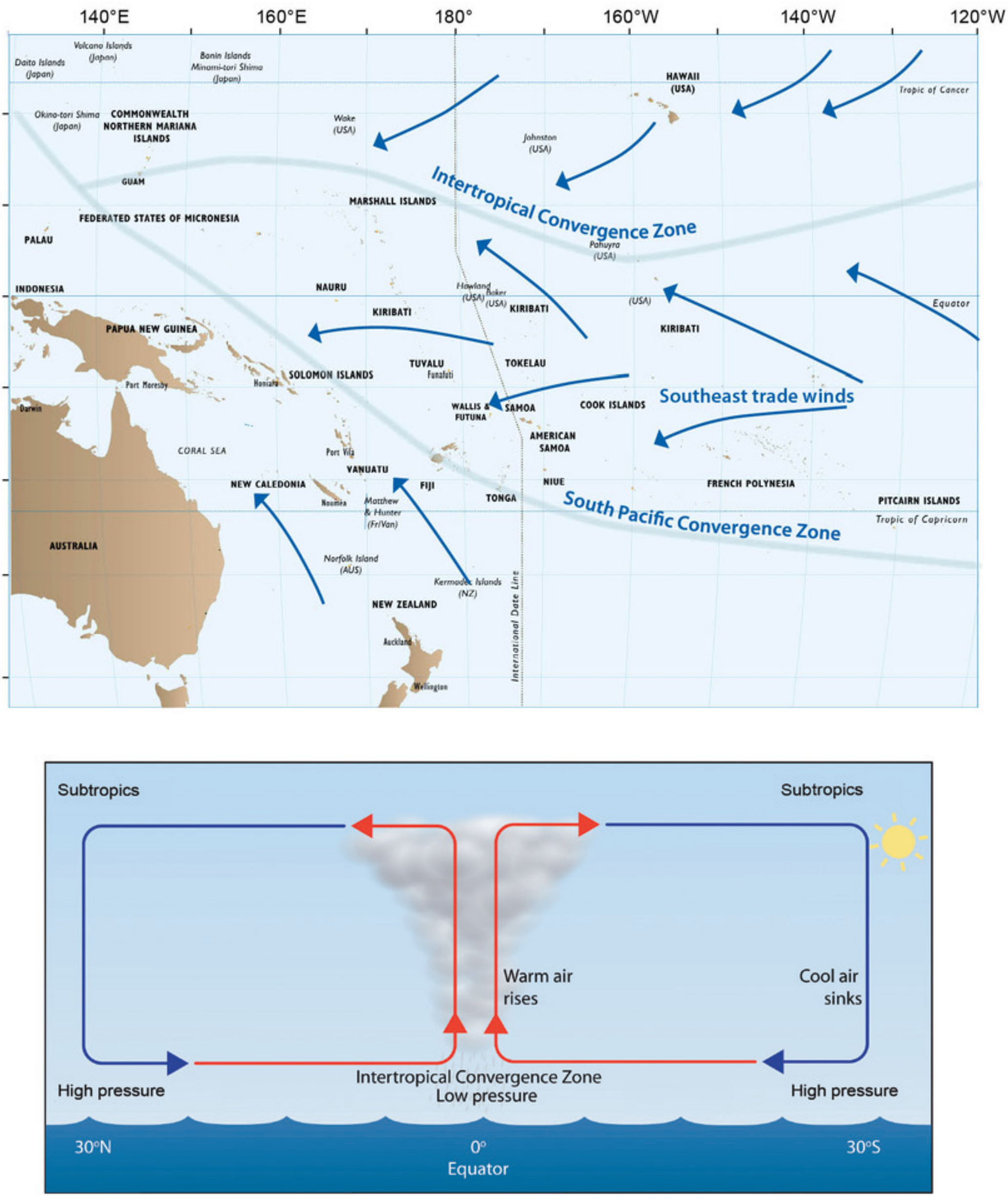


Figure 2.4 (a) Major atmospheric circulation features in the tropical Pacific; (b) cross-section illustrating the Hadley circulation of the region.

The SPCZ (**Figure 2.5**) is one of the most significant features of subtropical Southern Hemisphere climate[39,40]. It is characterised by low-level convergence of air flow leading to uplift and a band of cloudiness and rainfall stretching from the 'Warm Pool' in the western Pacific southeastwards towards French Polynesia[39–41]. The SPCZ shares some characteristics with the ITCZ, which lies just north of the equator, but is more subtropical in nature, especially east of the dateline[42]. To the west, it is linked to the ITCZ over the Warm Pool. To the east, it is maintained by the interaction of the trade winds and transient disturbances in the mid-latitude westerly winds emanating from the Australasian region. The SPCZ tends to lie over a region of large sea surface temperature (SST) gradient, rather than the maximum of SST, and is most active in the austral summer period, November–April. The location of the SPCZ convergence maximum varies considerably between seasons, by 10–15° of latitude (**Figure 2.6**). This causes large variability in rainfall throughout the southwest Pacific.

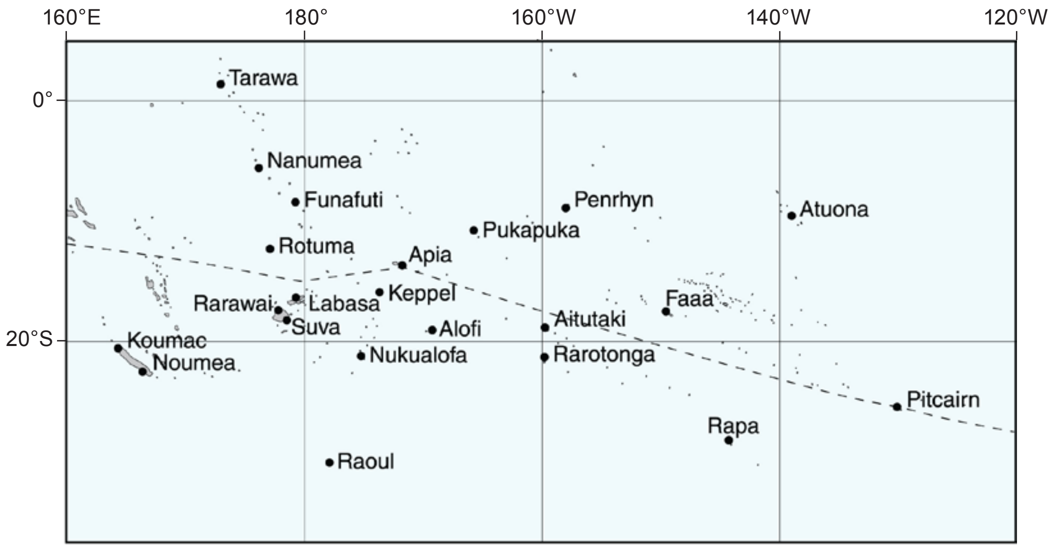


Figure 2.5 Mean position of the South Pacific Convergence Zone (SPCZ) (dashed line), defined as a maximum of low-level convergence (source: Folland et al. 2002)[53]. The zonal portion of the SPCZ lies west of 180º, and the diagonal section to the east. The locations of stations used in the analyses of instrumental temperature and rainfall records (Section 2.4) are also shown.

2.3.2 Temperature

The tropical Pacific consists primarily of ocean, with extremely small land areas (Chapter 1), and therefore the mean temperature climate is dominated by the average SSTs. The SSTs are similar to air temperatures for low-lying land areas and thus are a good proxy for surface temperatures. In general, average SSTs are warmer in the western Pacific compared to the eastern Pacific (**Figure 2.7a**). Annual maximum temperatures of around 30°C characterise the Warm Pool (**Figure 2.7b**) (Chapter 3) and annual minimum water temperatures (**Figure 2.7c**) exceed the threshold considered suitable for coral reef growth[43] throughout the region. The annual range of SSTs is < 2°C throughout much of the western tropical Pacific (**Figure 2.7d**). This is quite a narrow range to which marine organisms will be adapted, meaning small temperature changes may have relatively large impacts.

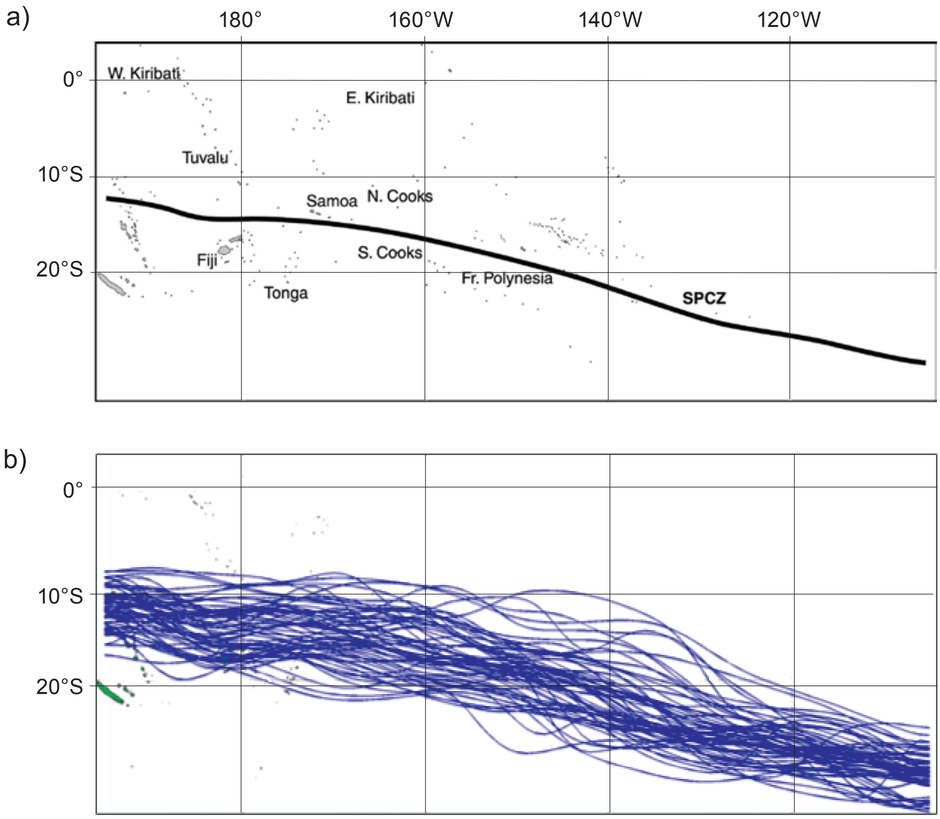


Figure 2.6 Average position of the South Pacific Convergence Zone (SPCZ), in (a) November–April, and (b) individual positions in all seasons, illustrating the considerable interannual variability in the SPCZ's mean position, which can give rise to substantial rainfall anomalies either side of its locations. The SPCZ is defined here as the maximum of convergence calculated from monthly mean reanalysis of wind speeds measured at 10 m elevation and direction with the mean SPCZ location at each longitude taken as the position of maximum convergence between the equator and 30°S.

2.3.3 Seasonal variation in winds, rainfall and temperatures

The average seasonal variation in the surface climate of the tropical Pacific is described here using monthly average wind fields, SSTs and rainfall (**Figure 2.8**). How this large-scale Pacific seasonality translates into average island climates is illustrated for five stations (**Figure 2.9**). These show the very small annual air and sea temperature variation of the near-equatorial sites of Tarawa and Funafuti, with an austral summer rainfall maximum; the greater annual air and sea temperature range of Nadi, with a more marked austral summer rainfall maximum; and the greater annual air and sea temperature and range for subtropical Rarotonga and Pitcairn, with little seasonality in monthly rainfall.

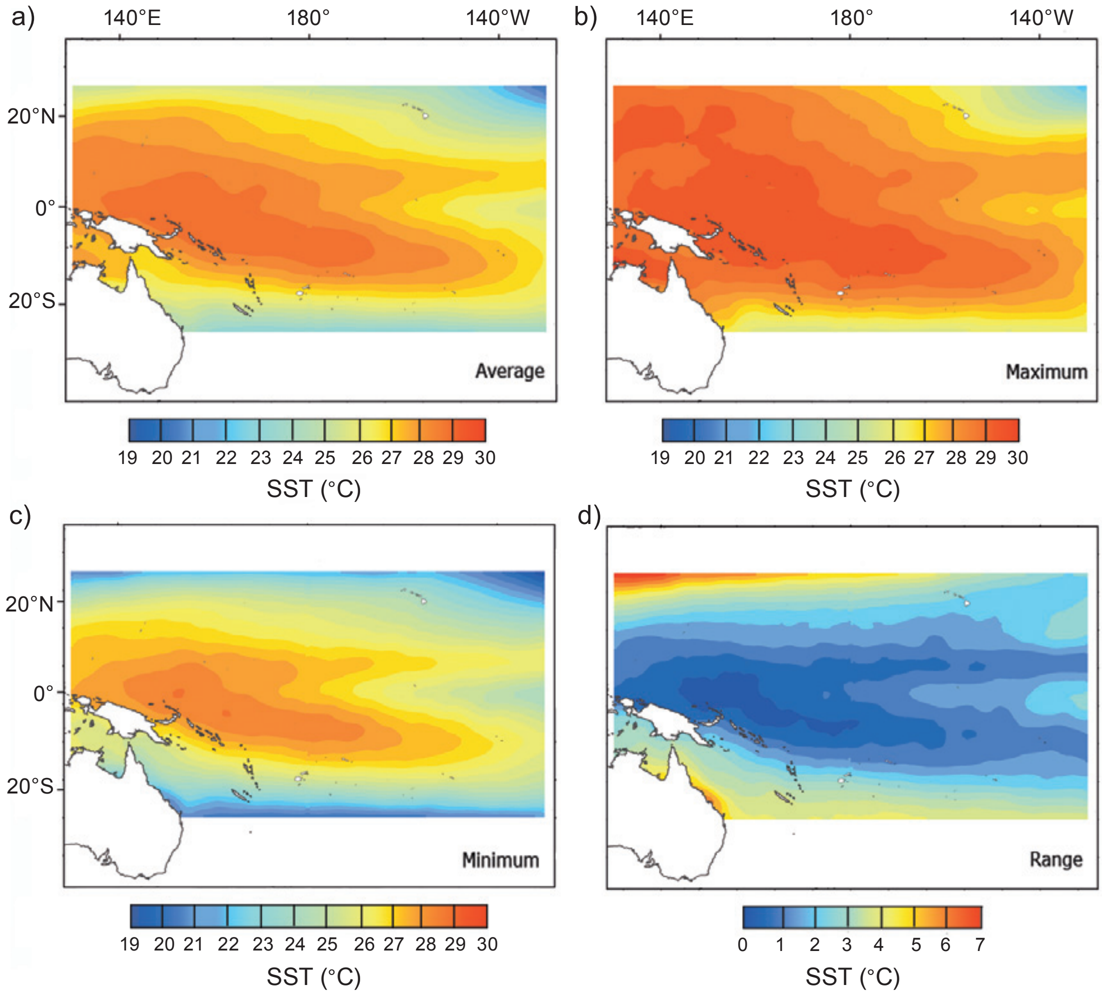


Figure 2.7 Average sea surface temperature (SST) in the tropical Pacific, 1950–2007, for (a) annual average, (b) annual maximum, (c) annual minimum and (d) annual temperature range (source: Rayner et al. 2003, 2006, HadISST)[97,98,100].

2.3.4 Tropical cyclones

Tropical cyclones are the most destructive weather disturbances that affect parts of the Pacific[44]. They are rarely observed within about 5–10° of the equator and, thus, their main impacts are on the islands of the southwest and northeast tropical Pacific (**Figure 2.10**) during the respective summer seasons. Tropical cyclones bring strong winds, high rainfall, storm waves and destructive storm surges to affected islands and are currently graded in the region from category 1 through to the most severe, category 5 (**Table 2.2**). In the southwest Pacific, tropical cyclones usually develop in the summer season, from November to April, but occasionally occur in May. Peak cyclone occurrence is usually from January to March.

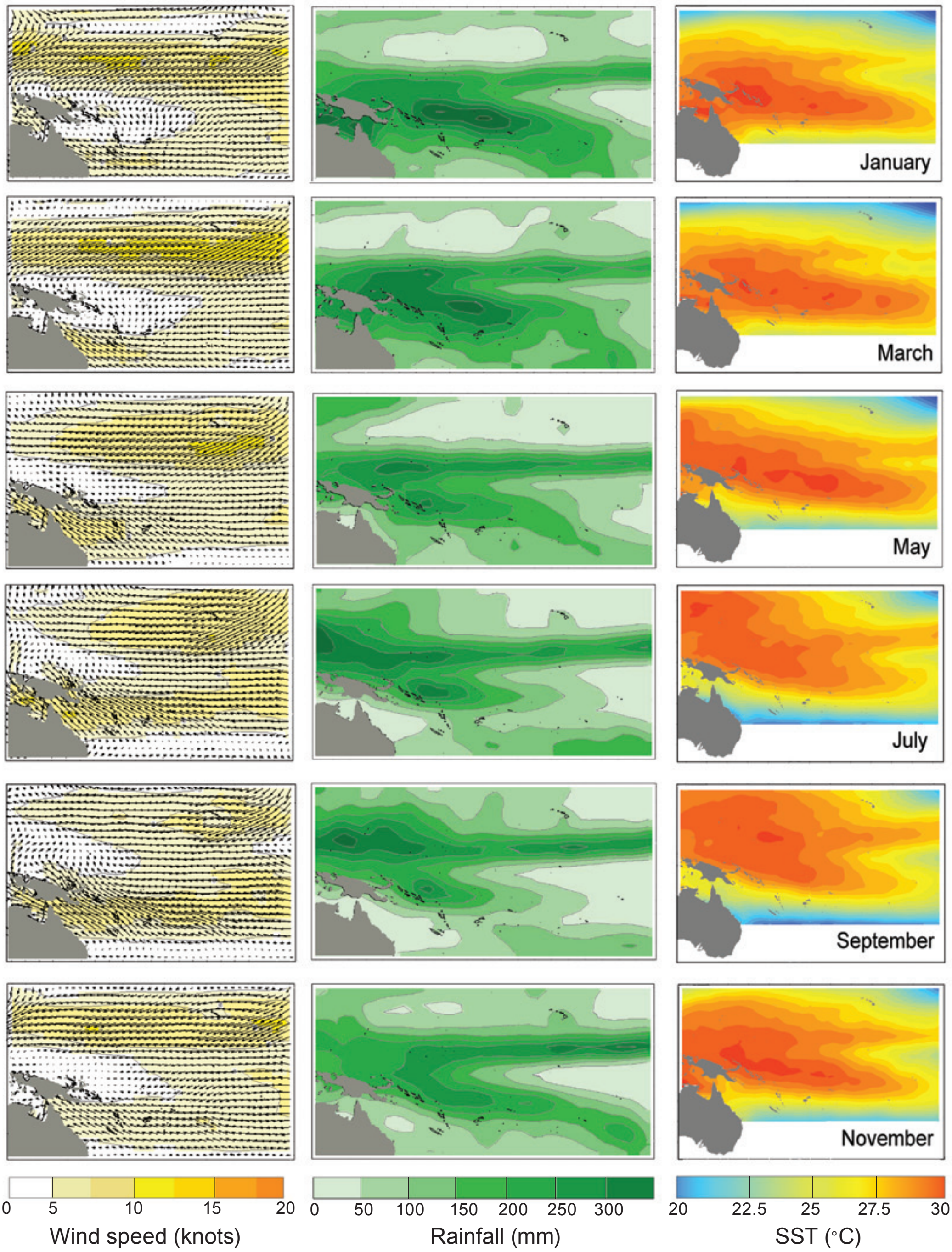


Figure 2.8 Surface climate averages, every two months throughout the year, for the tropical Pacific. Left: monthly average wind speed and direction (1971–2000) (NCEP Reanalysis). Centre: monthly rainfall, mm/month (1979–1995) (source: CAMS-OPI climatology, Janowiak and Xie 1999)[102,103]. Right: monthly average sea surface temperature (HadISST)[100].

Several tropical cyclones usually occur between Vanuatu and Niue each year with some affecting other areas. About half of the tropical storms that develop in the Pacific reach cyclone force with mean wind speeds of at least 118 km/h. Regionally, the highest frequencies of tropical cyclones occur between New Caledonia and Vanuatu in the Coral Sea, and towards Fiji (**Figure 2.11**).

Table 2.2 Tropical cyclone category and severity scale indicating typical magnitude of effects (source: Australian Bureau of Meteorology)[113].

Category	Strongest gust (km/h)	Average maximum wind (km/h)	Central pressure (hPa)	Typical effects
1	< 25	63–88	> 985	Negligible house damage. Damage to some crops, trees and caravans. Craft may drag moorings.
2	125–164	89–117	985–970	Minor house damage. Significant damage to signs, trees and caravans. Heavy damage to some crops. Risk of power failure. Small craft may break moorings.
3	165–224	118–159	970–955	Some roof and structural damage. Some temporary dwellings destroyed. Power failures likely.
4	225–279	160–199	955–930	Significant roofing loss and structural damage. Many caravans destroyed and blown away. Dangerous airborne debris. Widespread power failures.
5	> 279	> 200	< 930	Extremely dangerous with widespread destruction.

2.3.5 Intra-annual, interannual and decadal Pacific climate variability: ENSO, PDO and SAM

Superimposed on the average seasonal cycles of surface climate (Figure 2.8) and observed trends in surface climate (Section 2.4), are various sources of natural climate variability that modulate atmospheric and oceanic climate on time scales from weeks to decades.

The El Niño-Southern Oscillation (ENSO) phenomenon is the principle source of interannual global climate variability. This highly coupled ocean-atmosphere phenomenon is centred in the tropical Pacific. ENSO has significant impacts on climate and society, both within the region and, through teleconnections, in many distant parts of the world[13,45–48]. ENSO fluctuates between two phases, El Niño and La Niña, that disturb the normal atmospheric and oceanic circulations in the tropical Pacific (**Figure 2.12**). Both phases typically evolve over a period of 12–18 months and have some predictability once they have started to develop[49,50].

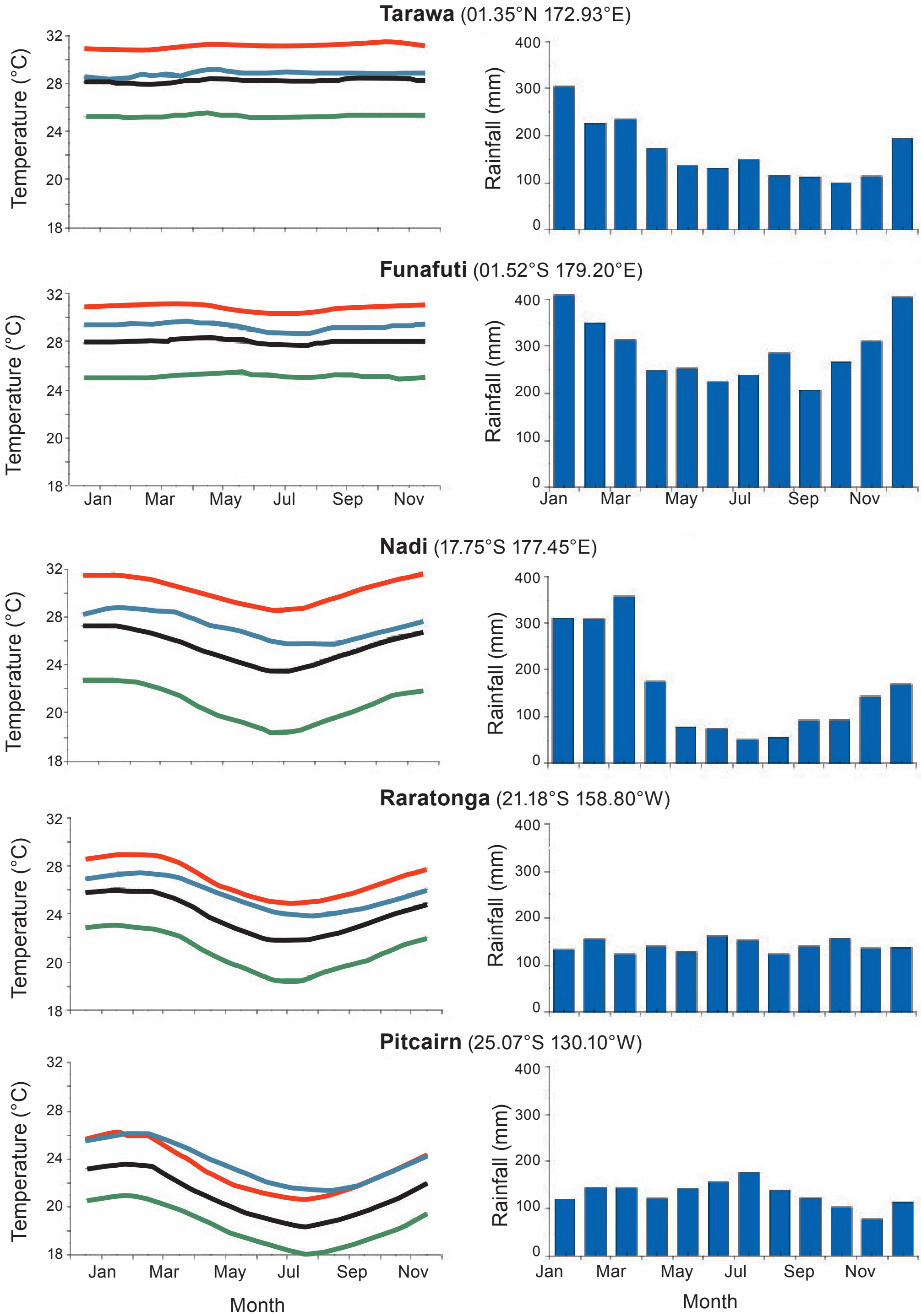


Figure 2.9 Average seasonal climate for five Pacific island stations, 1951–1980 (source: NIWA)[92]. Left: monthly average air temperature (black), monthly maximum air temperature (red), monthly minimum air temperature (green) and sea surface temperature (blue). Right: monthly total rainfall. The latitude and longitude of each station are also shown.

During El Niño events, the easterly trade winds weaken along the equator and a large part of the Pacific experiences unusually warm SSTs. This is associated with a weakening of the horizontal Walker circulation and strengthening of the meridional Hadley circulation. The centre of intense tropical convection shifts eastward towards the dateline and the ITCZ and SPCZ move closer to the equator. As a result, some regions experience drought conditions, while others receive much greater than average rainfall. There are also shifts in the dominant location of tropical cyclone activity. The slope of the thermocline (separating warmer surface and cooler deeper waters) flattens across the Pacific and the Warm Pool shifts eastwards (Chapter 3).

Climate anomalies during La Niña events are typically opposite to those of El Niño episodes, with stronger trade winds and large parts of the tropical Pacific experiencing cooler than normal SSTs. There are changes in the usual locations of tropical cyclones and a shift of the heaviest rainfall zone to the far western tropical Pacific. The depth of the thermocline increases from east to west across the Pacific during La Niña events.

The reliability of seasonal outlooks for ENSO conditions has improved significantly and is based on being able to successfully observe and model the development of SST anomalies in the tropical Pacific up to a year in advance of an event[47]. Two commonly used indices of ENSO activity are (1) the Southern Oscillation Index (SOI), which measures the atmospheric component and reflects anomalies in sea-level pressure between Tahiti in the southwest Pacific and Darwin in northern Australia (**Figure 2.13a**), and (2) the Niño 3.4 region (5°N–5°S, 170°W–120°W) average SST anomaly, which captures the oceanic component of ENSO (**Figure 2.13b**). These indices are very similar, indicating the highly coupled ocean-atmosphere nature of ENSO, but also show differences in the timing and magnitude of individual events that typically recur every 3 to 7 years.

Each ENSO event evolves slightly differently, and there have been suggestions that the major features of ENSO (termed 'ENSO-Modaki') have recently changed[51]. However, there are features common to these different ‘flavours’[52], and features typical of the two phases can be determined by averaging the surface climate anomalies across several events. Large parts of the tropical Pacific typically experience significantly warmer than normal SSTs during El Niño events and, conversely, significantly cooler than normal SSTs during La Niña events (**Figure 2.14**). Different patterns of rainfall are also associated with the two phases (**Figure 2.15**).

ENSO events also affect the spatial occurrence of tropical cyclone activity. During El Niño episodes, the overall number of tropical cyclones tends to be lower, with highest occurrences between Vanuatu and Fiji, and chances of occurrence higher further east in Samoa, southern Cook Islands and French Polynesia. During La Niña events, tropical cyclones are more frequent in the Coral Sea, with highest occurrence around New Caledonia, and higher occurrence between the coast of Queensland and Vanuatu in the Coral Sea. There is an absence of tropical cyclones from Cook Islands eastwards during La Niña episodes (**Figure 2.16**).

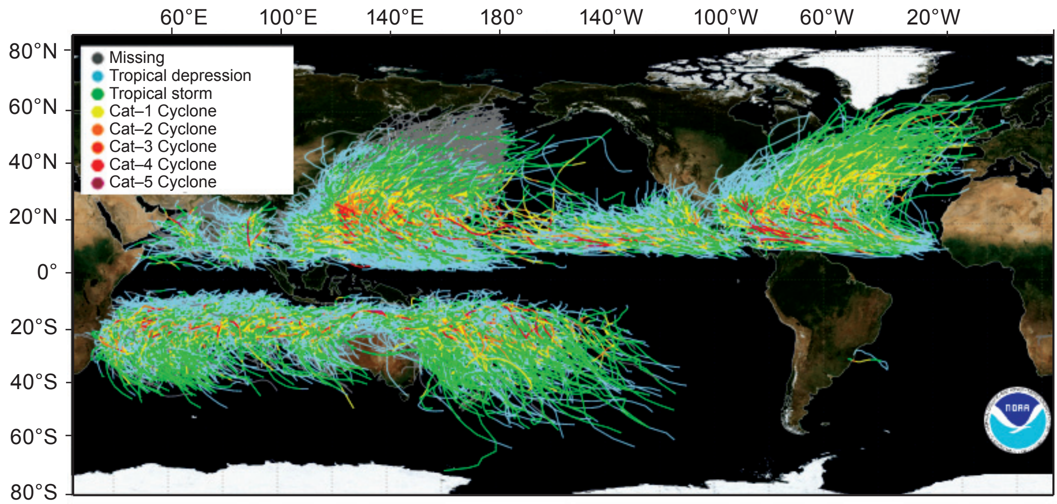


Figure 2.10 Tropical storm and cyclone tracks, 1947–2007, illustrating the areas of the tropical Pacific that are regularly affected by tropical storm activity and the near-equatorial region, which does not experience disturbances from these severe weather events (source: IBTrACbtracs)[104].

The location of the SPCZ also varies systematically with ENSO-related expansion and contraction of the Warm Pool[53]. Such movements can result in very large anomalies in precipitation on either side of the mean location of the SPCZ[54] as it moves northeast during El Niño events and southwest during La Niña episodes (**Figure 2.17**).

The interannual variability of ENSO and the strength of its climate teleconnections are modulated on decadal timescales by a long-lived pattern of Pacific climate variability known as the Pacific Decadal Oscillation (PDO)[55,56] or the Interdecadal Pacific Oscillation (IPO)[57]. The PDO is the North Pacific part of a Pacific basin-wide pattern encompassed by the IPO. It is described by an 'El Niño-like' pattern of Pacific SST anomalies and appears to persist in either a warm or cool phase for several decades (**Figure 2.13c**). Warm phases characterised the 1920s to 1940s and the period from the mid-1970s to at least the 1990s. In these periods, ENSO was a weaker source of interannual climate variability. They were preceded and separated by IPO and PDO cool phases from the 1900s to 1920s, and 1940s to 1970s, when ENSO was a major source of interannual climate variability[58]. Decadal variability in the SST field of the tropical Pacific is associated with decadal variability in atmospheric characteristics such as sea-level pressure, winds and precipitation[58,59].

The Southern Annular Mode (SAM) is the most important source of variability in the atmospheric circulation of the mid to high latitudes of the Southern Hemisphere, where it operates on time scales longer than ~ 50 days. It is characterised by a zonally symmetric pattern of atmospheric circulation with pressure anomalies of opposite sign in mid and high latitudes[60,61]. SAM fluctuates between two phases, which can be defined by the sea-level pressure difference between ~ 45°S and ~ 65°S. The strength of the westerly winds of the Southern Ocean is enhanced in the positive phase of SAM and weakened during the negative phase. The two phases influence surface climate patterns of rainfall and temperature over Antarctica and the mid-high latitudes of Southern Hemisphere land masses. Variations in the strength of the zonal westerly winds associated with SAM also influence sea level and ocean circulation patterns (Chapter 3). There is also observational evidence that SAM has become more positive in recent decades, enhancing westerly winds in the Southern Ocean[13].

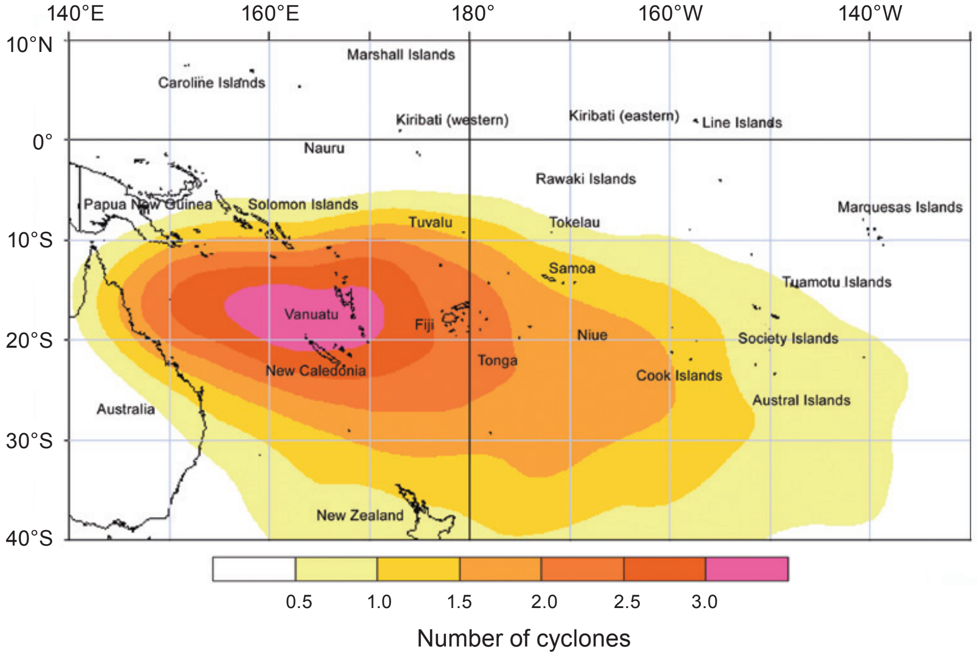


Figure 2.11 Average spatial occurrence of tropical cyclones in the southwest Pacific over the period 1970–1971 to 2001–2002, illustrating maximum numbers near Vanuatu (source: NIWA)[92].

2.4 Recently observed climate trends

Average global air and sea surface temperatures have already warmed significantly (**Figure 2.2**) and the rate of warming appears to be accelerating (0.12°C per decade for the period 1950–2008 compared to 0.04°C per decade from 1850 to 2008). Sea surface temperatures from 1950 to 2007 averaged over the tropical Pacific (25°N–25°S, 130°E–130°W) also show significant warming. For SST, however, the rate of change in the region is less than the global average, with values for the annual average of 0.07°C per decade, for the annual maximum of 0.08°C per decade and annual minimum of 0.06°C per decade (**Figure 2.3**). Linear trends can be distorted by the choice of the start and end year, but a comparison of temperature differences for two 20-year subperiods (1989–2008 and 1950–1969)[1] supports the linear trend analyses showing that there have been significant changes in temperature globally (+0.44°C), and in tropical Pacific SSTs (+0.23°C, +0.26°C and +0.19°C, for average annual, maximum and minimum temperatures, respectively). These large-scale averages, however, hide important spatial differences in the patterns of warming in the tropical Pacific (**Figure 2.18**), with some regions warming more than others, and some parts showing no significant trends as yet.

Here, we look in more detail at observed changes in various climate variables, mainly in the south Pacific where considerable effort has improved and homogenised the instrumental climate records[2,62]. The observational networks were still quite sparse in the first half of the 20th century but improved markedly after World War II.

2.4.1 Surface air temperature

Surface air temperatures, averaged over the southern tropical Pacific, have warmed significantly since the late 19th century (**Figure 2.19**). The rate of recent warming has accelerated (from 0.03°C per decade from 1850 to 0.11°C per decade since 1950) and 8 of the 10 warmest years between 1850 and 2007 have occurred in the last 10 years. As noted for SSTs, the changes are not uniform and the southern tropical Pacific can be divided into distinct areas on the basis of observed temperature trends: southeast trades (16°S to 27°S, 164°E to 144°W), central Pacific (2°N to 17°S, 172°W to 139°W) and convergence zones (1°N to 12°S, 169°E to 179°E). Temperature trends in the central Pacific and convergence zones are quite similar but differ from those in the area of the southeast trade winds[54,63,64].

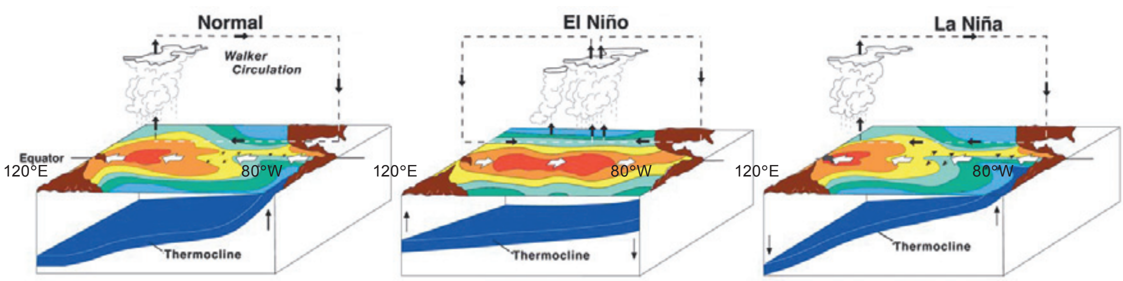


Figure 2.12 Pacific climate conditions during normal years (left), El Niño events (centre) and La Niña events (right) (source: McPhaden 2004)[47].

Average surface air temperatures in the area of the southeast trades have increased significantly but at a slightly lower rate than for the southern tropical Pacific as a whole (**Figure 2.20a**). The rate of warming since 1950 is 0.08°C per decade and 5 of the 10 warmest years on record have occurred in the last 10 years. There are, however, differences in the observed trends for daily maximum and minimum temperatures and between seasons over the period of high-quality observational data, 1941–1990 (**Table 2.3**). Significant warming occurred in annual and austral summer and autumn daily mean and daily minimum air temperatures by ~ 0.4–0.5°C. Warming of daily maximum air temperatures over this period was not significant and nor were the slight decreases observed in the daily temperature range.

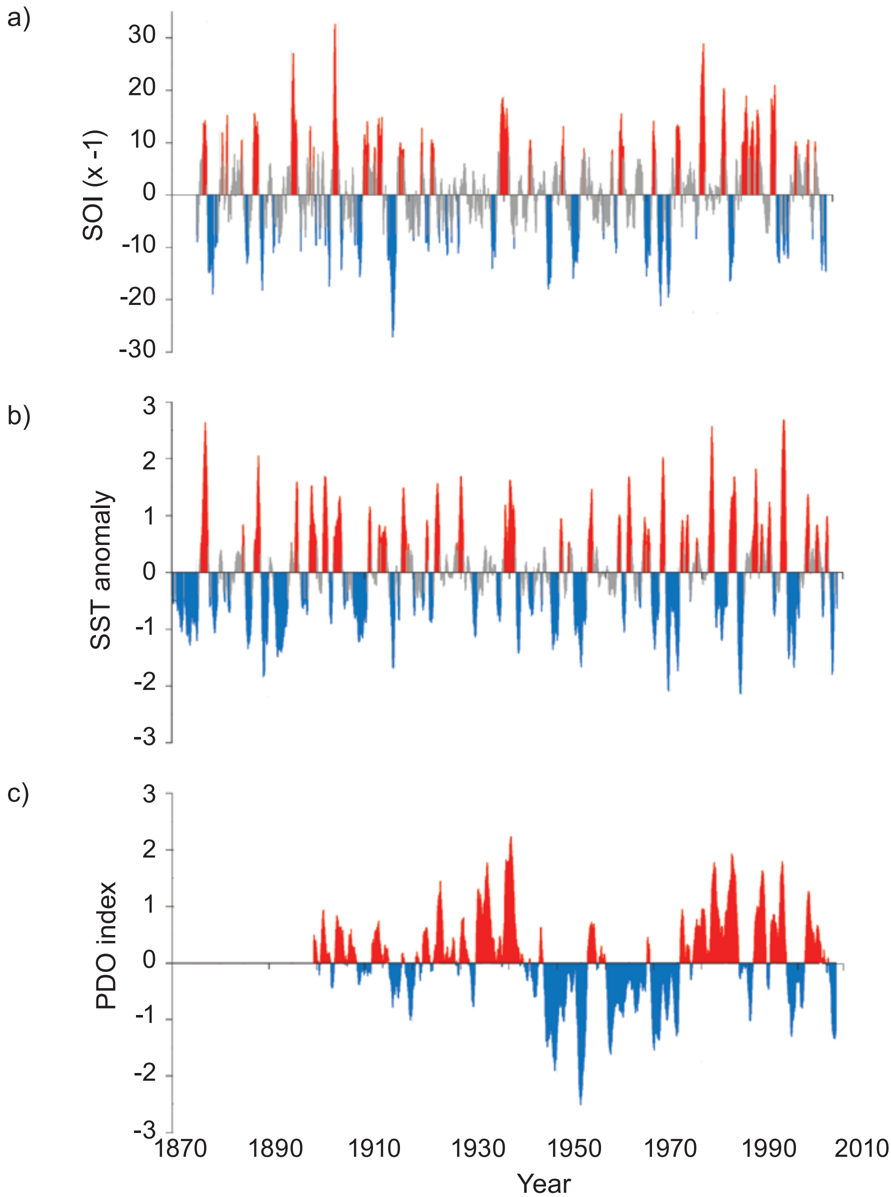


Figure 2.13 Indices of interannual and decadal climate variability in the tropical Pacific. (a) Southern Oscillation Index (SOI), standardised monthly anomaly of sea-level pressure difference between Tahiti and Darwin, values filtered with 5-month running mean and multiplied by -1; monthly values ±1 SD highlighted in red (positive El Niño) and blue (negative La Niña); for the period 1876–2008 (source: Troup 1965, Australian Bureau of Meteorology)[45,105]; (b) Niño 3.4 region (5°N–5°S, 170°W–120°W) sea surface temperature index, i.e. monthly sea surface temperature (SST) anomalies from 1971–2000 mean filtered with 5-month running mean and then standardised by 1971–2000 SD; monthly values ±1 SD highlighted in red (positive El Niño) and blue (negative La Niña); for the period 1871–2008 (source: Trenberth 1997, HadISST, Rayner et al. 2003, National Oceanic and Atmospheric Administration (NOAA))[97,100,106,107]; (c) Pacific Decadal Oscillation (PDO) Index based on Empirical Orthogonal Function (EOF) analyses of Pacific SSTs; monthly values filtered with 12-month running mean, for the period 1900–2008 (source: Mantua et al. 1997, Zhang et al. 1997)[55,56].

Warming of air temperatures in the central Pacific area is significant but is slightly less since 1850 than for the other regions (**Figure 2.20b**). The current rate of warming since 1950 is, however, similar at 0.10°C per decade and three of the 10 warmest years have occurred in the most recent 10-year period. Although daily mean, maximum and minimum temperatures have all warmed by ~ 0.3–0.4°C (**Table 2.3**) in all seasons, there have been no significant changes over the period 1941–1990.

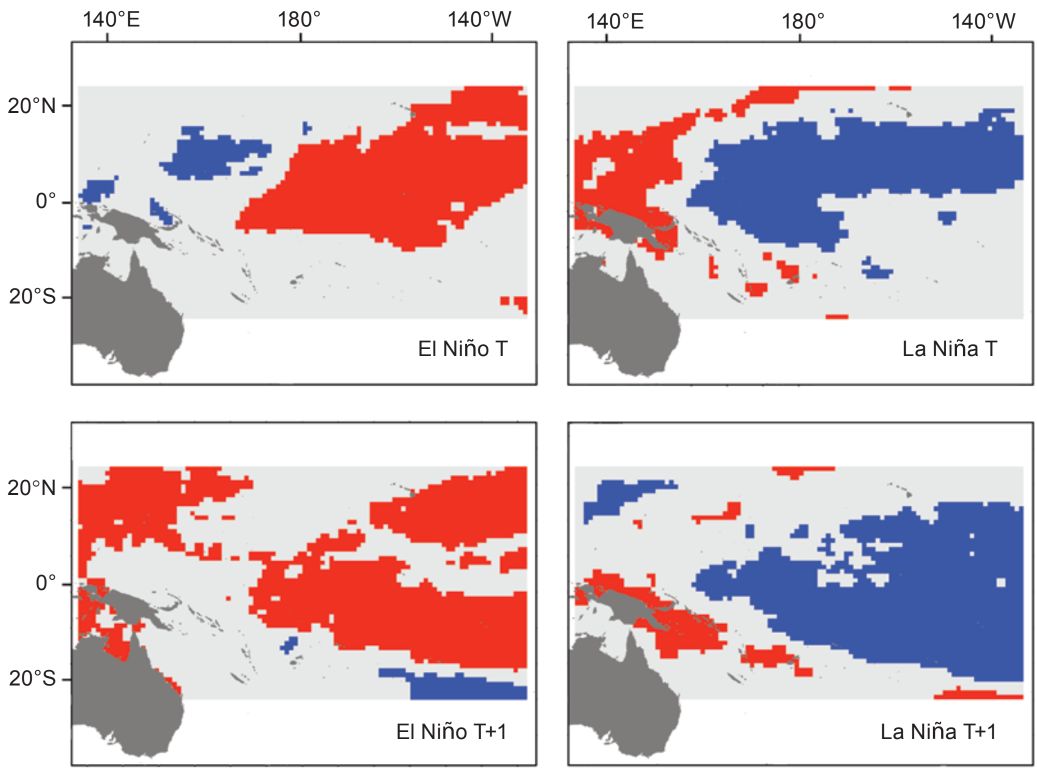


Figure 2.14 Areas with significantly warmer (red) or cooler (blue) maximum annual sea surface temperature (SST) during El Niño events (left) and La Niña events (right) for year T (top panels) and year T+1 (lower panels). SST anomalies were averaged for two years spanning 20 El Niño and 20 La Niña events and tested for significant difference from similar averages for 20 ENSO-neutral years within the period 1871–2002 (source: Rayner et al. 2003, 2006, HadISST)[97,98,100].

The area covered by the ITCZ and SPCZ also shows significant warming since 1850 (**Figure 2.20c**). The rate of warming since 1850 of 0.02°C per decade was slightly lower than for the other regions but has recently (since 1950) accelerated to 0.15°C per decade, and 5 of the 10 warmest years have occurred within the most recent 10-year period. Changes in daily temperature characteristics are not significant over the period 1941–1990, although in contrast to the other regions, a warming of maximum daily temperatures and slight cooling of daily minimum temperatures have resulted in an increase in the daily temperature range.

Table 2.3 Differences in seasonal and annual mean, maximum and minimum daily surface air temperatures (°C) determined from linear trend analyses from 1941 to 1990, for three areas in the Pacific (see Section 2.4 for the definition of these areas). Daily southern tropical temperature range is also shown.

Area	Surface air temperature (°C)			
	Mean	Maximum	Minimum	Daily range
Southeast trade winds				
DJF	0.46*	0.43	0.49*	-0.06
MAM	0.45*	0.36	0.52*	-0.16
JJA	0.36	0.33	0.35	0.02
SON	0.44	0.33	0.52	-0.19
Annual	0.45*	0.37	0.47*	-0.10
Central Pacific				
DJF	0.41	0.34	0.43	-0.09
MAM	0.43	0.26	0.52	-0.26
JJA	0.40	0.39	0.35	0.04
SON	0.49	0.37	0.53	-0.16
Annual	0.43	0.34	0.45	-0.11
ITCZ and SPCZ				
DJF	0.01	0.21	-0.08	0.29
MAM	-0.02	0.11	-0.09	0.20
JJA	0.05	0.24	-0.14	0.38
SON	0.01	0.06	-0.05	0.11
Annual	0.02	0.16	-0.09	0.25

* Indicates significance at the 5% level; ITCZ = Intertropical Convergence Zone, SPCZ = South Pacific Convergence Zone; DJF = December–February; MAM = March–May; JJA = June–August; SON = September–November (source: Salinger 1995)[64].

2.4.2 Rainfall

'At present, documenting interannual variations and trends in precipitation over the oceans remains a challenge.' (Trenberth et al. 2007)[13]

How much rain falls, when it falls, and the frequency and intensity of rainfall events all shape the 'climate envelope' of a given location. Detecting trends in rainfall, especially over tropical ocean regions, is hampered, however, by the relative lack of long-term observations, the inherently larger variability of rainfall in space and time compared to temperatures, and the high interannual variability of Pacific rainfall due to ENSO events[65,66].

Trends in rainfall and rainfall extremes have been examined for the period 1961–2000 for a number of stations with high-quality daily data (**Figure 2.5**) in the south Pacific[62]. Total annual rainfall (**Figure 2.21a**) shows a general decrease over the south Pacific southwest of the SPCZ and a general increase to the northeast of the SPCZ. The largest changes occurred east of 160°W.

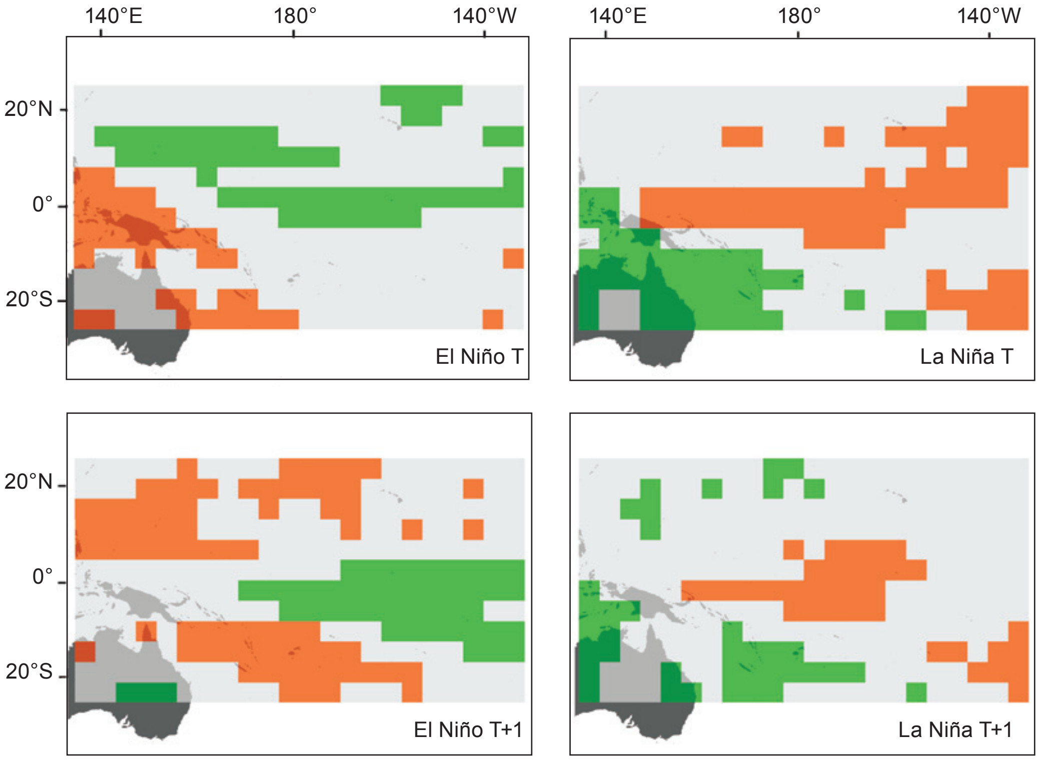


Figure 2.15 Areas with significantly wetter (green) or drier (orange) conditions during El Niño events (left) and La Niña events (right) for year T (top panels) and year T+1 (lower panels). Annual rainfall anomalies were averaged for two years spanning 16 El Niño and 14 La Niña events and tested for significant difference from similar averages for 15 ENSO-neutral years within the period 1900–2006 (source: Smith et al. 2008)[109].

Significant increases in total rainfall occurred only at Penrhyn (about 500 mm per decade) and at Atuona (about 250 mm per decade). Small but insignificant decreases in total rainfall occurred in southern Cook Islands, Rapa, Apia, Nuku'alofa, and Raoul Island, but there has been a significant decline of 180 mm per decade at Pitcairn. West of the dateline, trends in total rainfall are small and show no obvious spatial patterns within the island groups of Tonga, Fiji, New Caledonia and Tuvalu. Trends in total rainfall are spatially consistent east of the dateline, with a discontinuity across the diagonal portion of the SPCZ. These results are consistent with the diagonal SPCZ having moved northwards over the analysis period[53].

Trends in the number of days on which rain falls (rain day index ≥ 2 mm per day) are similar to those for total rain, with sites experiencing more rainfall typically having more rain days than sites with less total rainfall (**Figure 2.21b** cf. **Figure 2.21a**). East of 160°W, there was a significant increase in the number of rain days at Penrhyn (by 17 rain days per decade) and Atuona (by 13 days per decade), and a significant decrease at Pitcairn (by 8 days per decade). As for total rainfall, no significant trends or consistent patterns of change were found for the island groups of Tonga, Fiji, New Caledonia and Tuvalu. Again, these observed trends are consistent with the northward movement of the diagonal portion of the SPCZ.

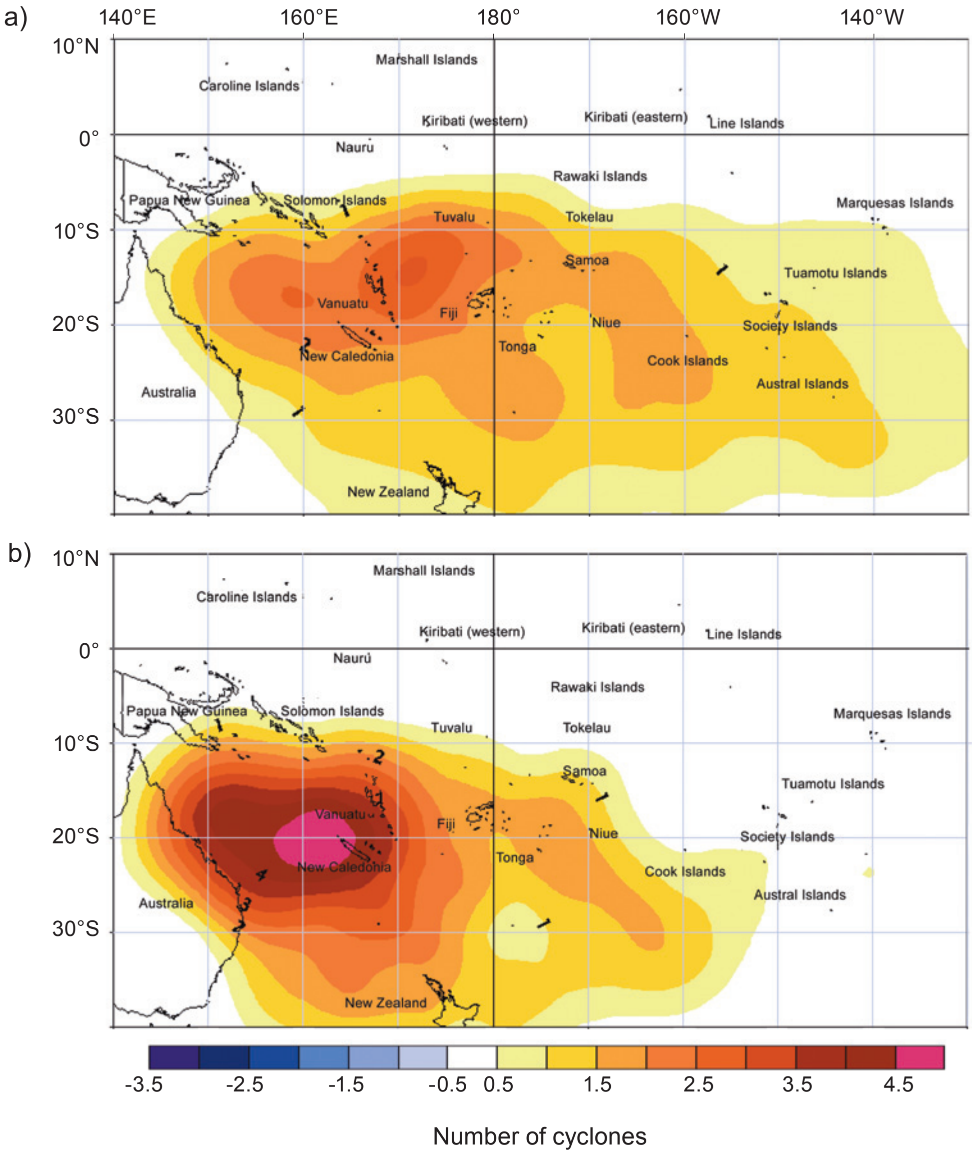


Figure 2.16 Typical spatial patterns of occurrence of tropical cyclones in the southwest Pacific for (a) El Niño and (b) La Niña seasons (source: NIWA)[92].

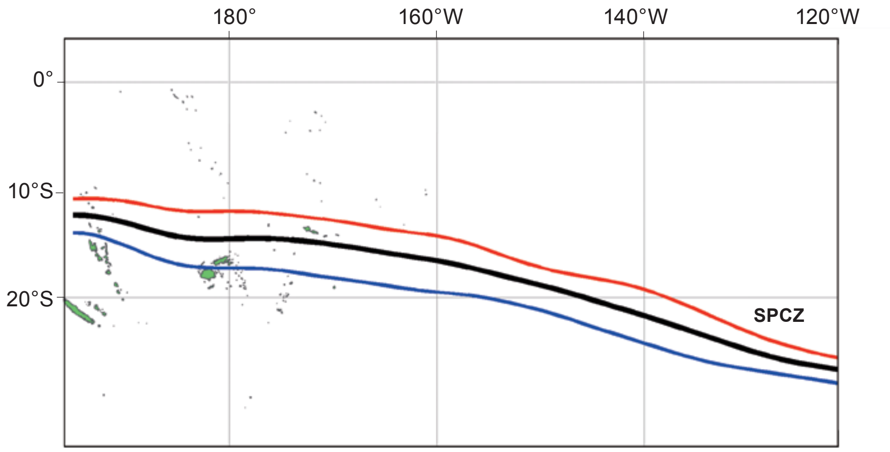


Figure 2.17 Average position of the South Pacific Convergence Zone (SPCZ) from November to April in El Niño (red), ENSO-neutral (black) and La Niña (blue) seasons.

The dry spell index is the maximum number of consecutive dry days each year (days < 1 mm of rain). Stations lying to the southwest of both the zonal and diagonal sections of the SPCZ, or near the SPCZ zone itself, showed an increase in dry spells, corresponding to a lengthening of the maximum dry period duration (**Figure 2.21c**). New Caledonia, Raoul Island, Tonga, Niue, Samoa, Pukapuka, Rarawai, Labasa and Pitcairn all show an increase in dry spells, with the greatest increase of 4.3 days per decade at Koumac. The only significant trend, however, occurs at Aitutaki, with an increase of 1.7 days per decade over the period 1961–2000. Unlike total rain and rain days, the dry spell indices for Tarawa, Penrhyn and Atuona show only small and insignificant decreases. The lengthening of the dry spell southwest of the entire SPCZ is consistent with the diagonal section of the SPCZ having moved northwards. Changes in the SPCZ are associated with an increase in mean sea-level pressure west of the dateline since 1977[67].

Three other indices of daily rainfall were examined: the 5-day rainfall index; the extreme intensity rainfall index (calculated as the average of the highest four rainfall events each year) as an indicator of the typical size of an extreme rainfall event; and the extreme frequency index, which is a count of high rainfall events per year (frequency of daily rainfall exceeding the 1961–1990 mean 99th percentile (days)).

The 5-day maximum rainfall index increased significantly at Penrhyn and decreased significantly at Nanumea (**Figure 2.22a**). Trends in extreme rainfall intensity are again incoherent west of the dateline, and in the island groups of Fiji and the southern Cook Islands (**Figure 2.22b**). More intense extreme rainfall events occur in the vicinity of the diagonal SPCZ near 170°W (e.g. Rotuma to Alofi), and to the north of the diagonal SPCZ east of 170°W. There have been no significant changes in these extreme intensity events in New Caledonia, Niue, Apia and Pukapuka. Significant increases in extreme rainfall intensity are exhibited at Penrhyn and Atuona, which both show an increase of 10 mm per decade. Significant decreases in extreme intensity events occurred at Nanumea, Rapa and Pitcairn (with decreases ranging between 6 and 10 mm). As with total rain, the largest trends in this index occur in the eastern Pacific Ocean, e.g. east of 160°W, with the exception of Nanumea.

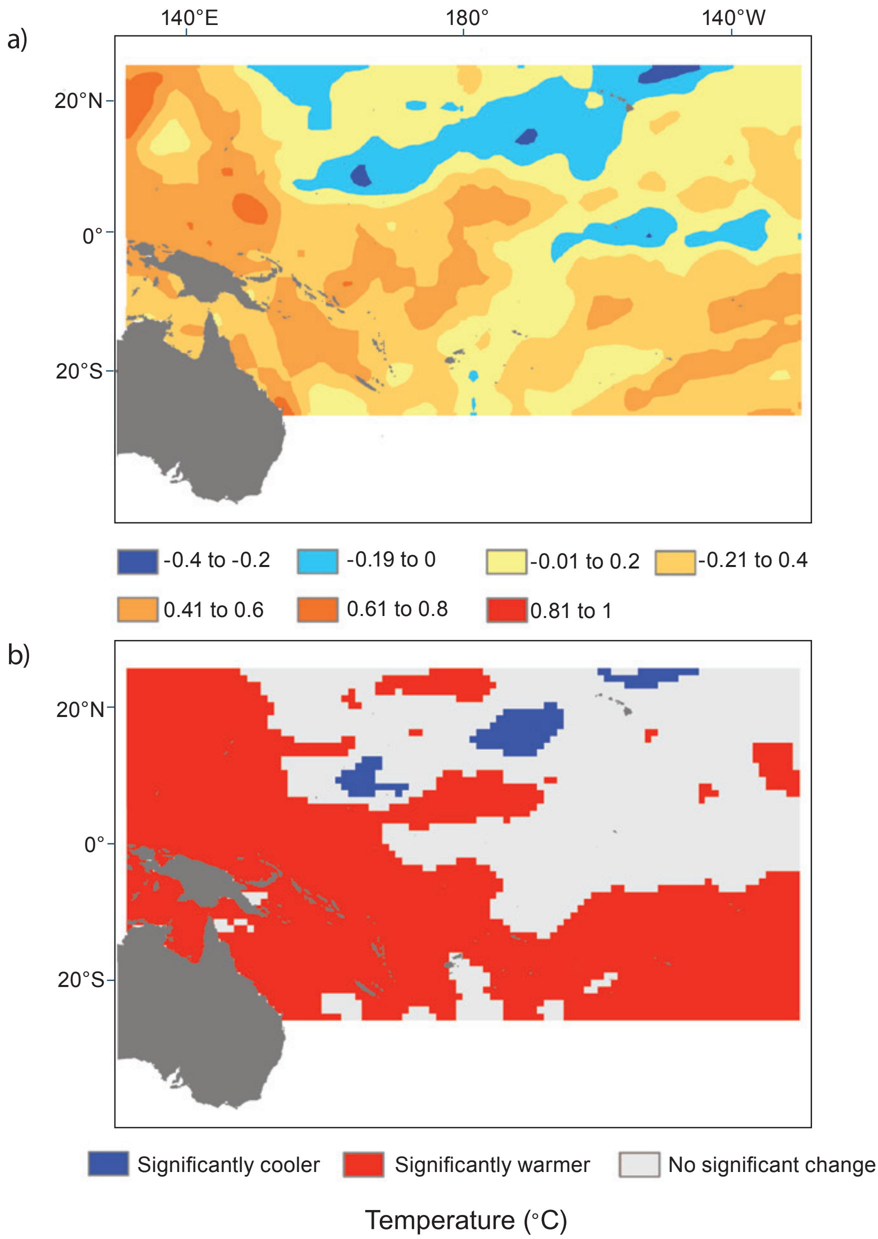


Figure 2.18 (a) Differences in average annual sea surface temperatures between the periods 1988–2007 and 1950–1969; and (b) the significance of differences between these periods (source: Rayner et al. 2003, 2006, HadISST)[97,98,100].

Trends in the extreme frequency index show the same sign and spatial pattern as the trends in extreme intensity. Generally, there has been a decrease in the frequency of extreme events south of the SPCZ (**Figure 2.22c**) and an increase to the north and in the vicinity of the SPCZ near 170°W. West of the dateline, trends are incoherent and small. As with total rain, significant increases in the extreme frequency occur only east of 160°W at Penrhyn and Atuona (1.5 and 1.2 days per decade, respectively). Significant decreases are seen at Rapa and Pitcairn (0.5 and 0.8 days per decade, respectively).

In summary, there have been some significant changes in the observed annual and daily rainfall climate of the southwest tropical Pacific, primarily on islands east of 160°W, with changes to the west tending to be small and incoherent. Generally, over the period 1961–2000, there has been more rainfall and more intense rainfall northeast of the SPCZ and less rainfall to the southwest of the SPCZ.

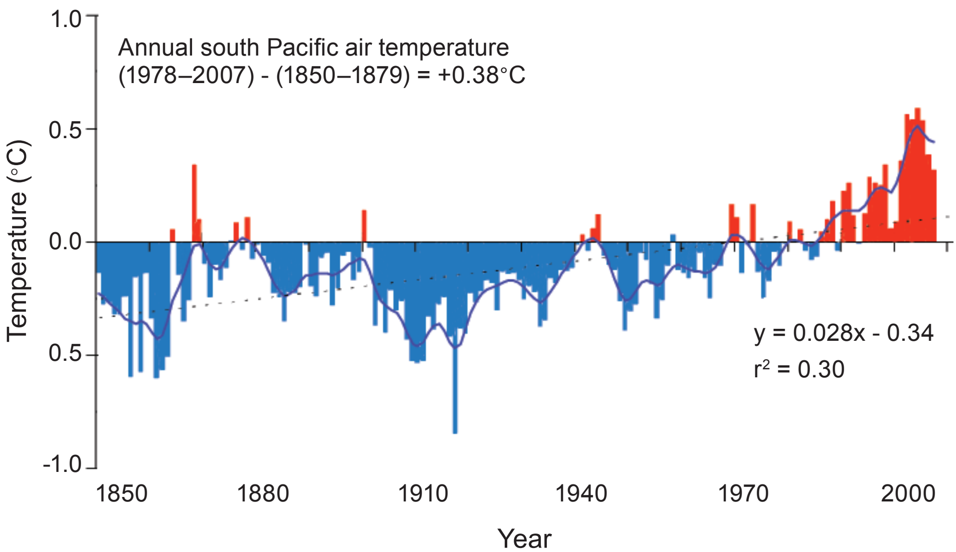


Figure 2.19 Annual average southern tropical Pacific surface air temperature anomalies (from 1971–2000 mean) for the period 1850–2007. Thick line is 10-year Gaussian filter emphasising decadal variability; dashed line is linear trend (source: HadCRUTv3)[94].

2.4.3 South Pacific Convergence Zone

The above analyses of observed changes in surface temperature and rainfall all highlight the significant role played by the SPCZ in the climate of the southwest Pacific. To assess whether there have been significant changes in the position of the SPCZ through time, high-quality, mean sea-level pressure data from stations in Suva (18°9′S, 178°26′E) and Apia (13°48′S, 171°47′W) were used to calculate a proxy index of SPCZ position. These two stations with long and reliable data often lie on opposite sides of the mean SPCZ location. Long-term variations (with positive values representing a displacement towards Apia, and negative values a displacement towards Suva) in the position of the SPCZ do not show any significant long-term change in the position over the period 1890–2005 (**Figure 2.23**). There are, however, decadal variations in the position that closely align with the warm and cool phases of the IPO/PDO (warm and cool refer to the relative sign of SST anomalies in the tropical Pacific) (**Figure 2.13c**).

During the cool phases of the IPO/PDO (e.g. mid-1940s to mid-1970s), the mean position of the SPCZ is displaced to the southwest. Conversely, during the warm phases of the IPO/PDO since the late 1970s, and during the 1920s–1930s, the SPCZ was northeast of its average position. The marked shift that occurred in the mid-1970s, when the IPO/PDO transitioned from relatively cool to warm, has been attributed to a combination of external anthropogenic forcing and internally generated decadal variability[68]. Since the late 1990s, the position of the SPCZ has been displaced southwest. Thus, the SPCZ, one of the most extensive features of the global atmospheric circulation, varies its location according to both the polarity of ENSO, and of the PDO. This significantly affects rainfall patterns throughout the south Pacific.

2.4.4 Tropical cyclones

There are, on average, nine tropical cyclones in the southwest Pacific per season over the period of reliable monitoring from 1969/70 to 2007/08 (**Figure 2.24**). About four tropical cyclones reach at least category 4 each season, with mean wind speeds of at least 118 km/h, and at least two will usually reach category 5, with mean speeds in excess of 167 km/h. There is no discernible trend in the frequency of tropical cyclones in the southwest Pacific over the 30-year period. There is also no evidence, as yet, for any significant change in the intensity of those tropical cyclones that do occur in the southern Pacific Ocean. This contrasts with evidence for stronger tropical cyclones in the Atlantic Ocean and, to a lesser extent, other regions of tropical cyclone activity[69].

2.5 Projected changes in surface climate

Many of the projections of future climate change for the tropical Pacific region continue trends that have been observed over the past several decades. With ongoing increases in anthropogenic greenhouse gases, and given that most of the global temperature increases over the last half century are very likely to have been due to those greenhouse gas increases[70], we can expect such changes to continue for at least the next several decades. This is because the future emissions scenarios do not diverge much over the short term[71]. There is still, however, an urgent recognised need to both improve current climate models for the tropical oceans[25] and to produce better regional-scale climate projections[72]. It is possible that the relatively coarse resolution and still imperfect generation of the global climate models used here do not adequately capture unanticipated changes in tropical Pacific climate. In addition, as a result of the inertia inherent in the ocean, even with an immediate cessation

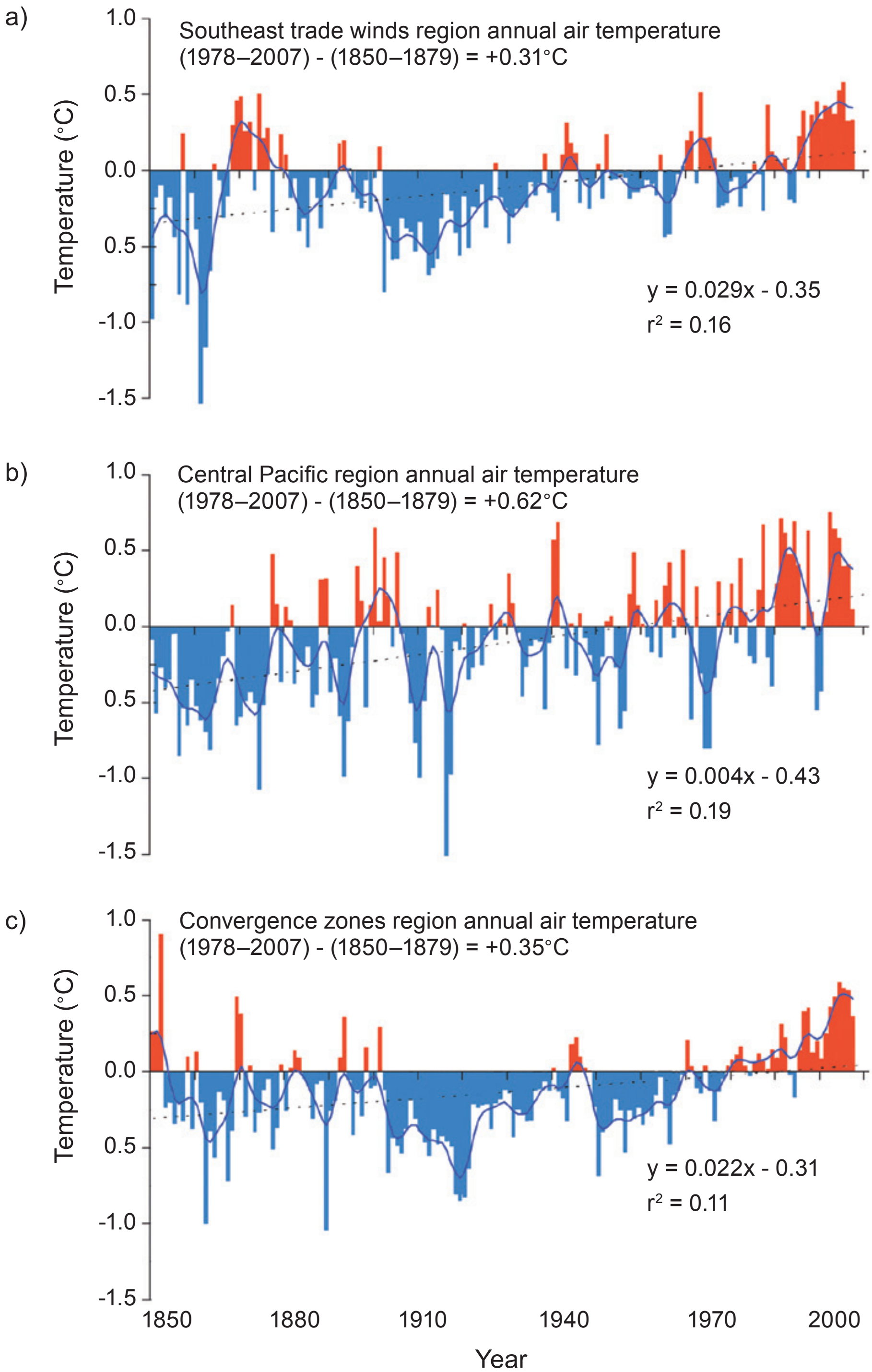


Figure 2.20 Annual average surface air temperature anomalies (from 1971–2000 mean) for the period 1850–2007 for (a) southeast trade winds, (b) central Pacific and (c) convergence zones regions of the tropical Pacific. Thick line is 10-year Gaussian filter emphasising decadal variability; dashed line is linear trend (source: HadCRUTv3)[94].

of greenhouse gas emissions, climate changes will still continue for many decades. Major differences in future climate outcomes due to different emissions scenarios do not become apparent until later in the 21st century.

In this section, we show projections for average temperature and precipitation change for the near-term (2035) and the longer term (2100) for the B1 low emissions scenario and the A2 high emissions scenario. Results from both these scenarios were extensively described in the IPCC-AR4[71]. Here we provide greater detail for the tropical Pacific region.

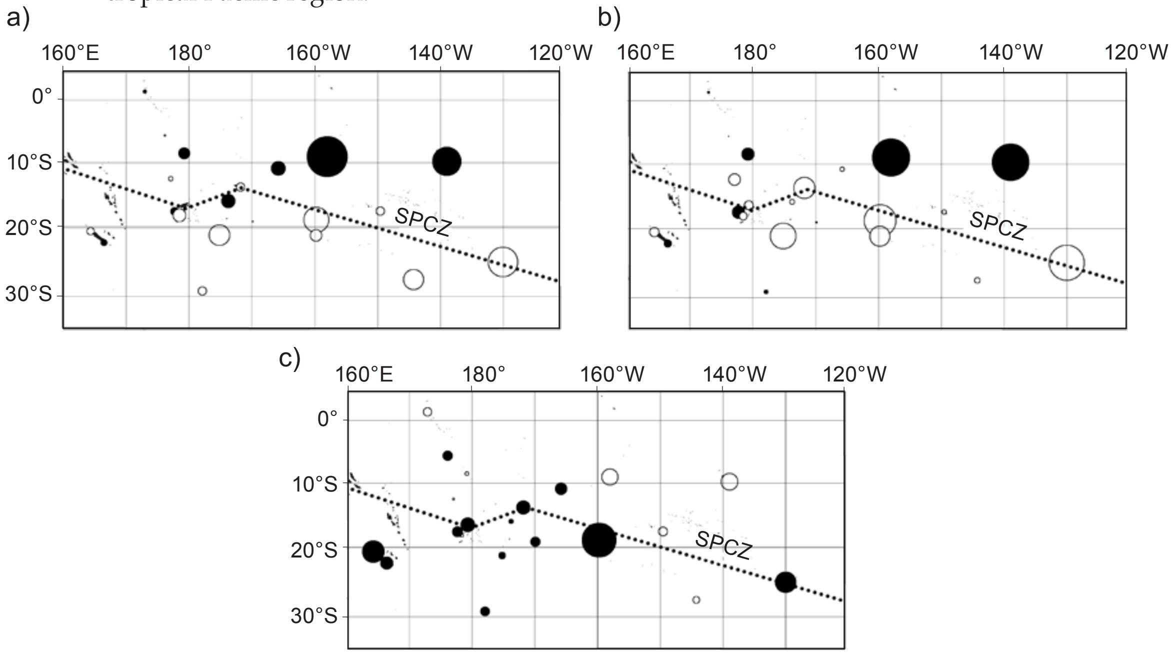


Figure 2.21 Trends in (a) total rainfall, (b) rain day index, and (c) dry spell index from 1961 to 2000, recorded at the meteorological stations shown in Figure 2.5. Black circles represent positive trends and white circles, negative trends. The size of the circle is proportional to the magnitude of the normalised linear trend per decade. SPCZ = mean position of the South Pacific Convergence Zone.

2.5.1 Air and sea surface temperatures

For the near-term (2035), both the low (B1) and high (A2) emissions scenarios show similar amounts of warming across the tropical Pacific, with values ranging from about +0.5°C to +1.0°C. Greater warming occurs in the equatorial regions compared to the subtropics[73], in particular in the eastern equatorial Pacific (**Figure 2.25a,c**).

Some of the warming is from climate change commitment; that is, the warming the climate system is committed to as a result of the additional greenhouse gases already in the system[32]. This is due to the fact that oceans warm more slowly than land and the thermal inertia of the oceans introduces a lag into the system. Averaged globally, climate change commitment amounts to about 0.10°C per decade for the next several decades[71]. Of course, the concentrations of greenhouse gases are also projected to increase in both the B1 and A2 scenarios over this time period. The globally-averaged temperature increase from that forcing, added to the commitment, is about 0.2°C per decade[71]. Therefore, over the next three decades, global warming would be about 0.6°C, close to the average projected for the tropical Pacific.

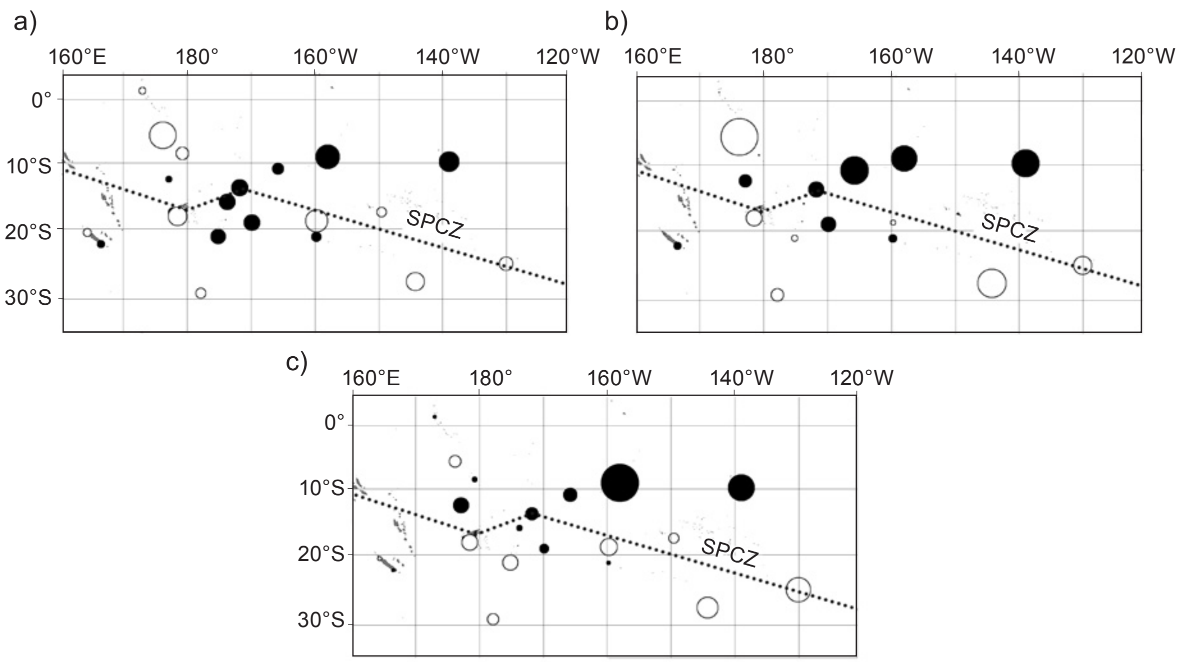


Figure 2.22 Trends in (a) 5-day rainfall, (b) extreme intensity rainfall, and (c) extreme frequency rainfall from 1961 to 2000, recorded at the meteorological stations shown in Figure 2.5. Black circles represent positive trends and white circles, negative trends. The size of the circle is proportional to the magnitude of the normalised linear trend per decade. SPCZ = mean position of the South Pacific Convergence Zone.

This additional warming over the next three decades is similar to the warming already observed (see previous section) and thus continues present-day trends. Future warming will also still be modulated by natural sources of climate variability[74]. These projections for 2035 assume, for example, that no large tropical volcanoes erupt. If a Pinatubo-type volcano were to erupt during this time period, the rate of warming would be reduced as was observed in the years after Pinatubo erupted in the early 1990s. However, after the volcanic aerosols clear from the atmosphere, the warming resumes and catches up (with some lag) in response to the forcing from the ever-increasing greenhouse gases that would continue during and after an eruption.

By the end of the century (**Figure 2.25b,d**), the emissions trajectory clearly makes a difference to projections. The warming of the tropical Pacific in the low B1 scenario (based on the multi-model averages) ranges from 1°C to 1.5°C, while in the high A2 scenario the warming is greater than 2.0°C over much of the tropical Pacific, with largest values in the eastern equatorial Pacific of 3.3°C. The annual mean values shown in **Figure 2.25** are also representative of seasonal mean values. **Table 2.4** illustrates what these short- and long-term temperature projections for the two scenarios might translate into in terms of average annual air temperatures at five Pacific island sites (**Figure 2.9**). Projections for SST are discussed in more detail in Chapter 3.

Table 2.4 Average observed annual air temperature (°C), 1951–1980, and projected ranges of estimated annual air temperatures (°C), for 2035 and 2100 under low emissions (B1) and high emissions (A2) scenarios (Figure 2.25) for five Pacific island stations (Figure 2.9).

Station	1951–1980 observed (°C)	2035		2100	
		B1	A2	B1	A2
Tarawa	28.2	29.0–29.2	28.7–29.0	29.7–30.0	31.2–31.5
Funafuti	28.0	28.5–28.8	28.5–28.8	29.3–29.5	30.8–31.0
Nadi	25.6	26.1–26.4	26.1–26.4	26.9–27.1	28.4–28.6
Raratonga	23.9	24.4–24.7	24.4–24.7	25.2–25.4	26.4–26.7
Pitcairn	20.9	21.4–21.7	21.4–21.7	21.9–22.2	22.6–22.9

* Note that the observed averages are for the period of good observational coverage, 1951–1980, whereas the projections are relative to the 1980–1999 average. Given the global warming observed already, the specific projections presented here may be underestimates.

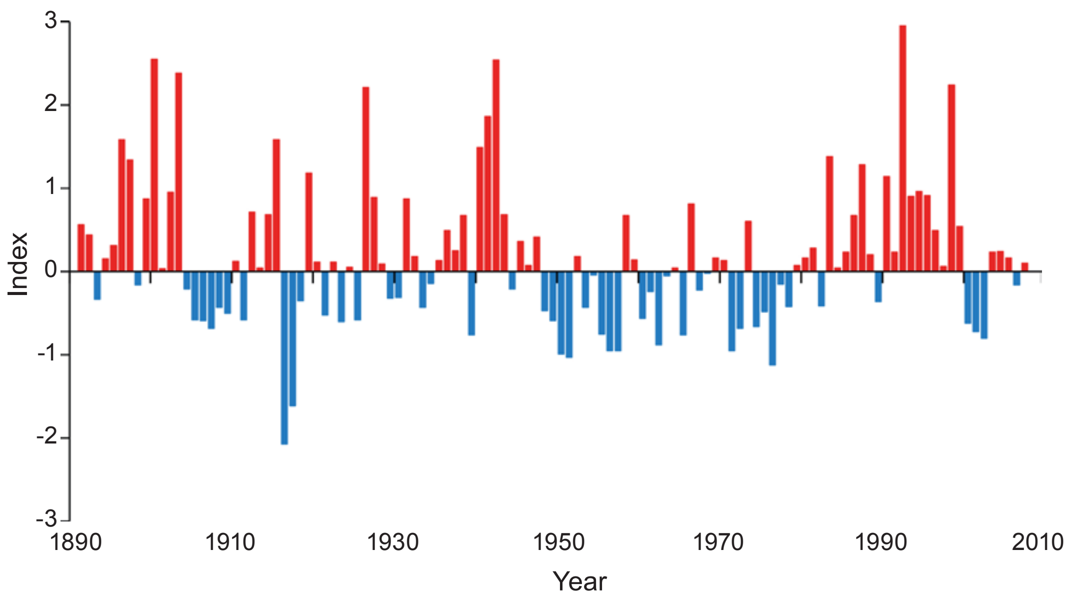


Figure 2.23 Annual index of the position of the South Pacific Convergence Zone (SPCZ) between 1891 and 2008. The index was calculated as the normalised November–April difference in mean sea-level pressure between Suva and Apia, based on the period 1961–1990. It defines the latitude of the SPCZ between longitudes 180° and 170°W.

2.5.2 Rainfall

In general, as tropical Pacific SSTs warm, rainfall increases in both the wet and dry seasons examined here (December–February and June–August) between about 10°N and 10°S, and decreases in the subtropics (**Figure 2.26**). As for temperature, the projected changes in rainfall are not significantly distinguishable between the two scenarios for the near term. However, by the end of the century, even though the pattern is similar, there are larger and more consistent increases of tropical precipitation between about 10°N and 10°S in the A2 scenario compared to the B1 scenario. As tropical SSTs warm, there is greater evaporation and moisture availability for precipitation, so rainfall increases as a consequence. With the increases in tropical precipitation, there is an intensification of the hydrological cycle, an expansion of the Hadley circulation (**Figure 2.4b**), and greater subsidence in the subtropics with reduced rainfall there.

These projected patterns show similarities to observed changes in tropical rainfall between 1991 and 2002, and 1979 and 1990, based on large-scale global precipitation data blended from satellite and surface rain gauge observations[66]. These observations show more total rainfall and more intense rainfall in the ITCZ and SPCZ regions of intense tropical convection, and small to negative changes in the southeastern Pacific. There are also some seasonal variations that could have implications for island water resources[75]. For example, in the A2 scenario, more consistently dry conditions are projected in the southwest Pacific for June–August. **Table 2.5** illustrates what these short- and long-term rainfall projections for the two scenarios might translate into in terms of December–February and June–August rainfall totals at five Pacific island sites.

Table 2.5 Average rainfall totals (mm) for December–February (DJF) and June–August (JJA), 1951–1980, and range of estimated rainfall totals for 2035 and 2100 for the B1 and A2 emissions scenarios (Figure 2.26) for five Pacific island stations (Figure 2.9). Grey shading indicates projected changes are within ± 5% of observed average; dark blue shading indicates wetter conditions, and orange drier conditions.

Station	Season	1951–1980 observed (mm)*	2035		2100	
			B1	A2	B1	A2
Tarawa	DJF	725	761–798	761–798	> 870	798–870
	JJA	397	> 476	437–476	> 476	> 476
Funafuti	DJF	1164	1106–1222	1106–1222	1222–1280	1280–1397
	JJA	748	785–823	785–823	823–898	> 898
Nadi	DJF	785	746–824	746–824	746–824	864–942
	JJA	182	173–191	173–191	173–191	191–200
Raratonga	DJF	426	405–447	405–447	447–469	469–511
	JJA	439	417–461	417–461	417–461	461–483
Pitcairn	DJF	377	358–396	339–358	339–358	302–339
	JJA	471	447–495	447–495	446–495	447–495

* Note that observed averages are for the period of good observational coverage, 1951–1980, whereas the projections are relative to the 1980–1999 average. Specific projections presented here may be underestimates.

2.5.3 El Niño-Southern Oscillation events

ENSO events (see Section 2.3.5) represent the largest interannual climate fluctuations in the tropical Pacific. These events have major effects on regional temperatures, tropical cyclone activity, SPCZ position and rainfall. Given the magnitude of climate anomalies associated with ENSO events and the disruption they can cause, it is of great interest to determine what the possible future changes in El Niño might be.

A number of factors combine to make such estimation difficult. The historical data show that there are multi-decadal changes of ENSO amplitude and frequency associated with the IPO/PDO (see Section 2.3.5). Such multi-decadal variations of ENSO have also been documented in long climate model simulations with no changes in greenhouse gases[76,77]. Given such large, natural multi-decadal variability, it is difficult to assess whether there have been any changes in ENSO frequency and intensity that can be attributed to human activities in the observed record.

A similar problem applies to future climate model simulations of ENSO. Models show a variety of ENSO changes in response to increasing greenhouse gases, but some could simply be sampling each model's inherent multi-decadal variability without any systematic change in ENSO amplitude or frequency. In fact, in an assessment of projected ENSO events from about 20 global coupled climate models there is no consistent change in ENSO behaviour in a future warmer climate (**Figure 2.27**).

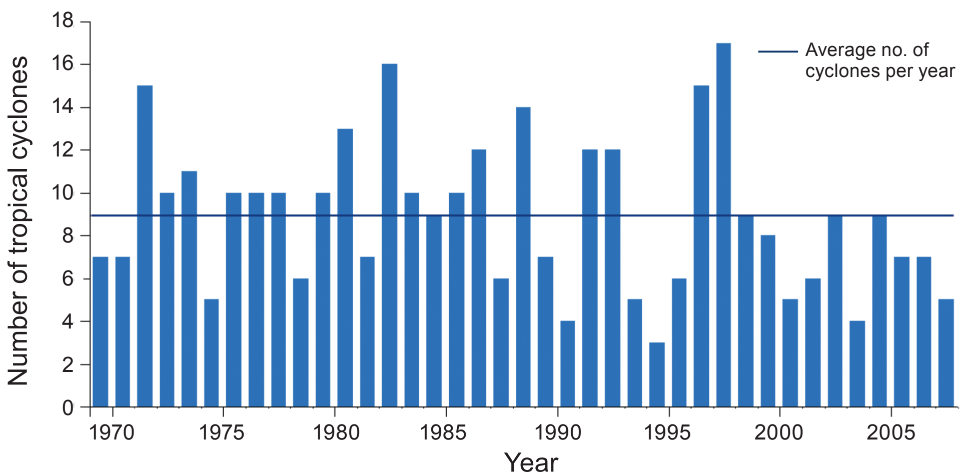


Figure 2.24 Annual number of tropical cyclones in the southwest Pacific, 1969/70–2007/08.

What is more common across the models is the tendency for an 'El Niño-like' response[78] in the mean climate change. That is, the majority of the models tend to show somewhat greater average warming in the eastern equatorial Pacific compared to the western equatorial Pacific (**Figures 2.25** and **2.27**). This tendency does not, however, suggest there will be a more permanent El Niño-like state in the future. In any case, all models show El Niño and La Niña events continuing in the future no matter which scenario is followed and thus they should be considered as an ongoing source of interannual climate variability and extremes in the tropical Pacific.

2.5.4 Tropical cyclones

As with any future climate change projection, we must rely on climate models to provide estimates of how tropical cyclones and their devastating consequences may change with continued global warming. The current generation of climate models discussed here has grid points (where the equations calculate winds, temperature, pressure, etc.) spaced about every 150 km in the atmosphere, and about every 100 km in the ocean. Near the equatorial Pacific, the grid spacing is closer, at about 50 km, to provide better representation of the dynamical mechanisms in the ocean associated with ENSO events. Given this resolution, El Niño and La Niña events are indeed simulated in the models, and large-scale temperature and rainfall changes (such as those depicted in **Figures 2.25** and **2.26**) are relatively well represented. However, it has been estimated that to simulate the most intense tropical cyclones, resolution of about 5 km is required. The ultimate climate model for studying the possible future changes to tropical cyclones in detail would be a global coupled atmosphere-ocean-land-sea ice model with an atmospheric model resolution of about 5 km. However, the expense of running such a climate model at such high resolution is prohibitive with present-day computing capabilities.

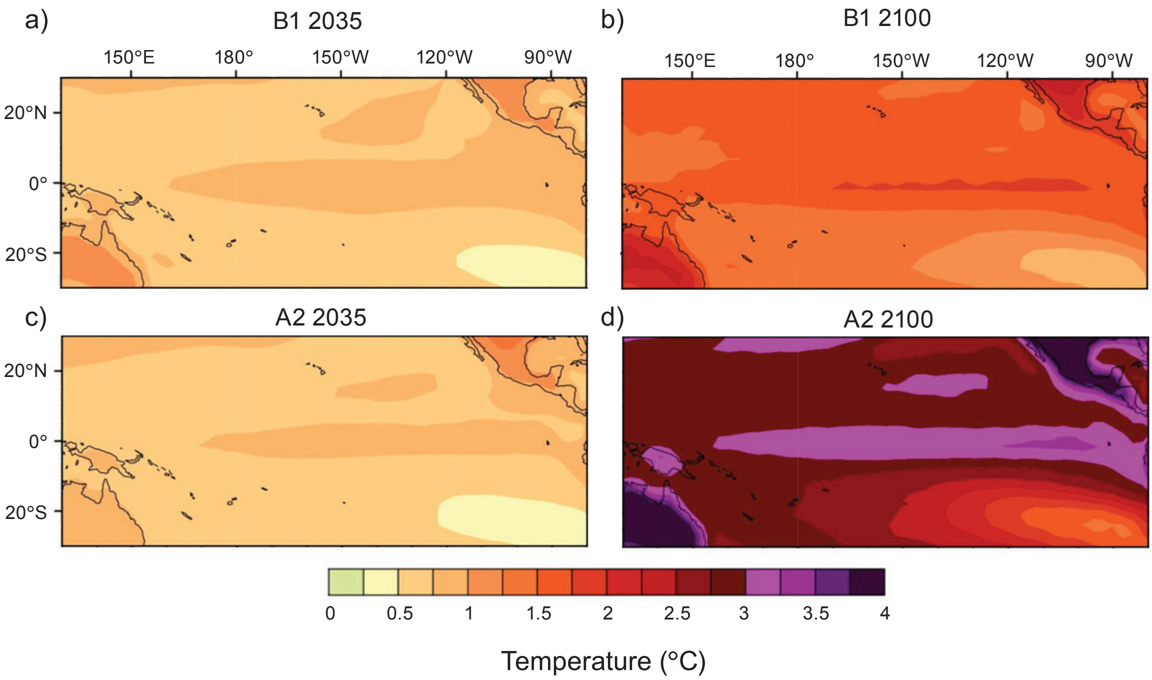


Figure 2.25 Differences in multi-model annual mean surface temperature relative to 1980–1999 for the low emissions (B1) scenario for (a) 2035 and (b) 2100; and for the high emissions (A2) scenario for (c) 2035 and (d) 2100.

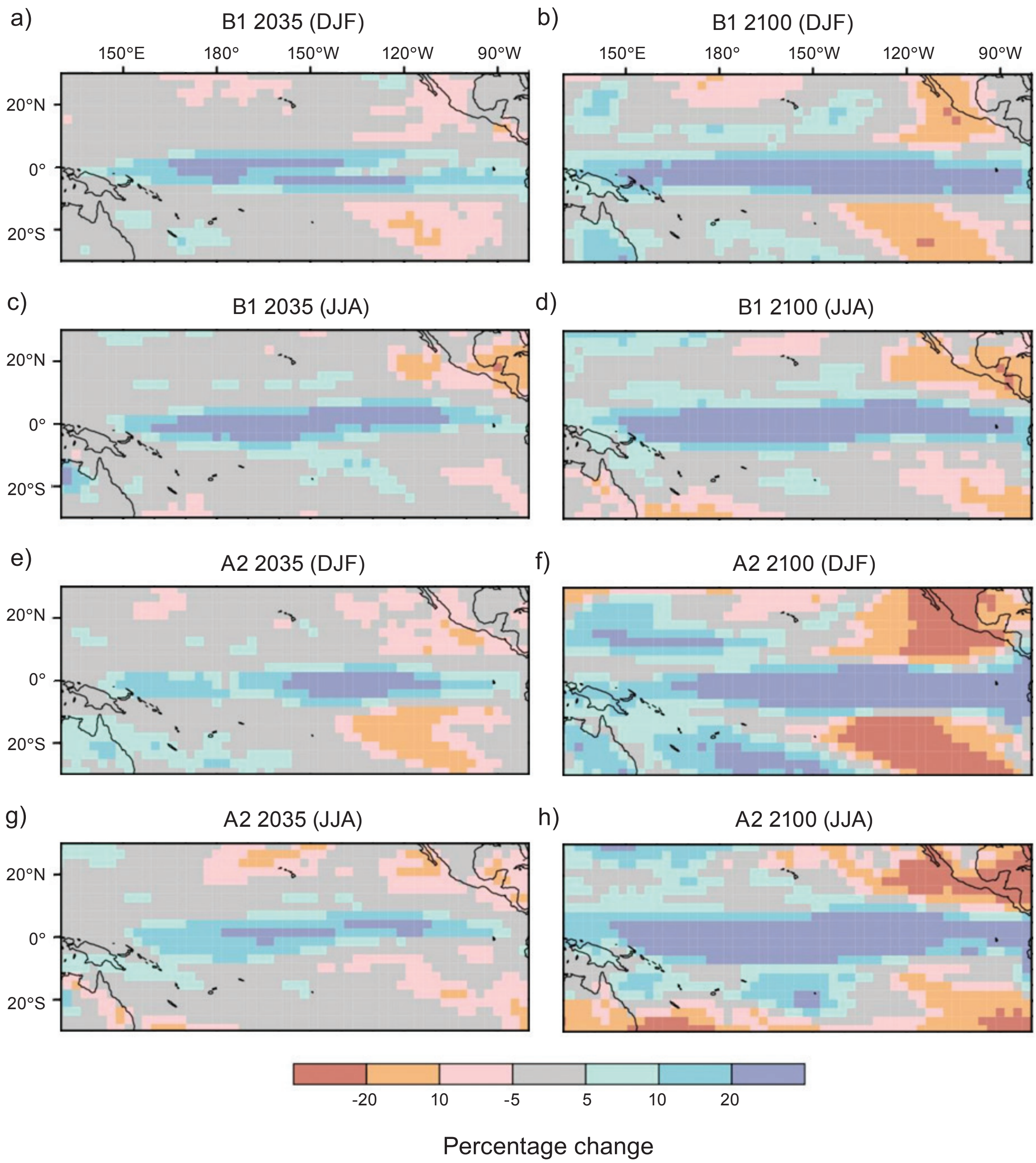


Figure 2.26 Differences in multi-model seasonal mean rainfall, relative to 1980–1999, for the low emissions B1 scenario for (a) December–February (DJF) 2035, (b) DJF 2100, (c) June–August (JJA) 2035, and (d) JJA 2100; and for the high emissions A2 scenario for (e) DJF 2035, (f) DJF 2100, (g) JJA 2035, and (h) JJA 2100.

Since the appropriate climate model to quantify possible future changes in tropical cyclones is well beyond what is currently feasible, there has, by necessity, been a reliance on other climate modelling tools. High-resolution (closer grid spacing) regional models have been embedded in the global climate models to better represent tropical cyclones and estimate future changes in a warmer climate[79–81]. Global atmospheric models with much higher resolution have been run for short periods of time (without being coupled to ocean models) to estimate possible future changes in tropical cyclones.

Overall, the results from these various modelling studies have shown that in a warmer climate, it is likely there will be an increase in the intensity of tropical cyclones. It has been estimated that for every 1°C of tropical SST increase, core rainfall rates would increase by 6–18%, and surface wind speeds of the strongest tropical cyclones would increase by about 1–8%[82]. These relatively small percentage changes in wind speed actually translate into quite large increases in the destructive potential of a tropical cyclone. There is less certainty with regard to the future number of tropical cyclones; several studies have suggested there could be fewer in a warmer climate, but those that form would be more intense[71,82,83].

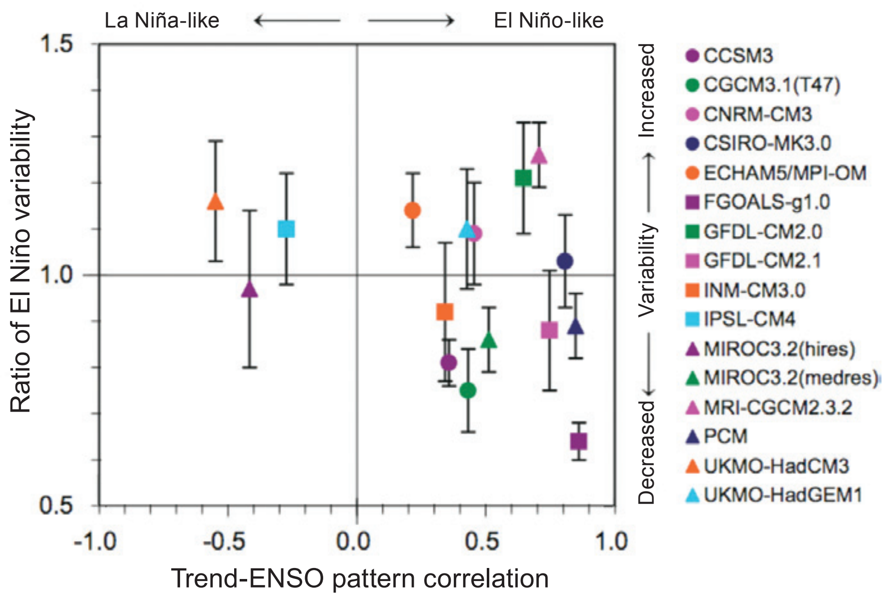


Figure 2.27 Variability among 16 global climate models in projecting future ENSO activity in the tropical Pacific, based on models producing more El Niño- and La Niña-like conditions (horizontal axis) and more or less interannual variability (vertical axis). Error bars indicate 95% confidence limits. If all models were agreeing about future ENSO changes, then values would cluster within one quadrant[71,110,111].

An example of one such study that showed this result used a global atmospheric model with 20-km resolution that could simulate up to strong category 3 or weak category 4 cyclones[84] (**Figure 2.28**). The atmospheric model was forced with observed time series of SSTs for a 10-year period at the end of the 20th century to generate the model's present-day climatology of tropical cyclones (**Figure 2.28b**). This was then compared to observed tropical cyclones for that same period (**Figure 2.28a**). Future ocean temperatures were derived from a lower resolution, global coupled climate model, and those SST anomalies were added to the 20th century SSTs to produce a future 10-year time series of warmer SSTs. The atmospheric model was run with those SSTs along with double the atmospheric concentration of CO_2 and 2.5°C global warming by the end of the 21st century. The resulting simulation (**Figure 2.28c**) shows fewer cyclones in the tropical Pacific, but more intense winds in those that do form (**Figure 2.29**).

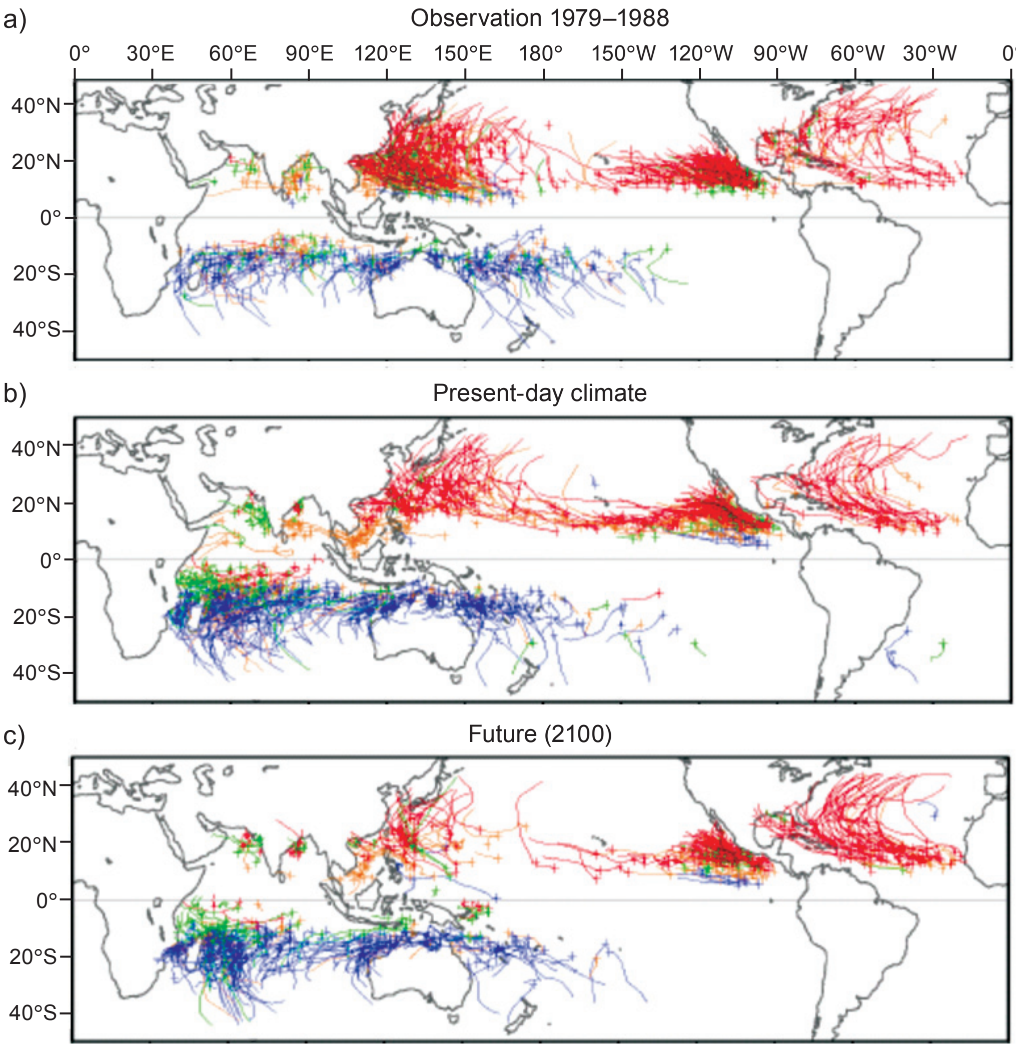


Figure 2.28 Tropical cyclone tracks worldwide from (a) observations (1979–1988), (b) simulations from a 20-km resolution atmospheric model for a 10-year present-day climate, and (c) simultations from a future 10-year period around 2100. Red tracks are for the northern summer, and blue tracks are for storms occurring during the southern summer (source: Oouchi et al. 2006)[84].

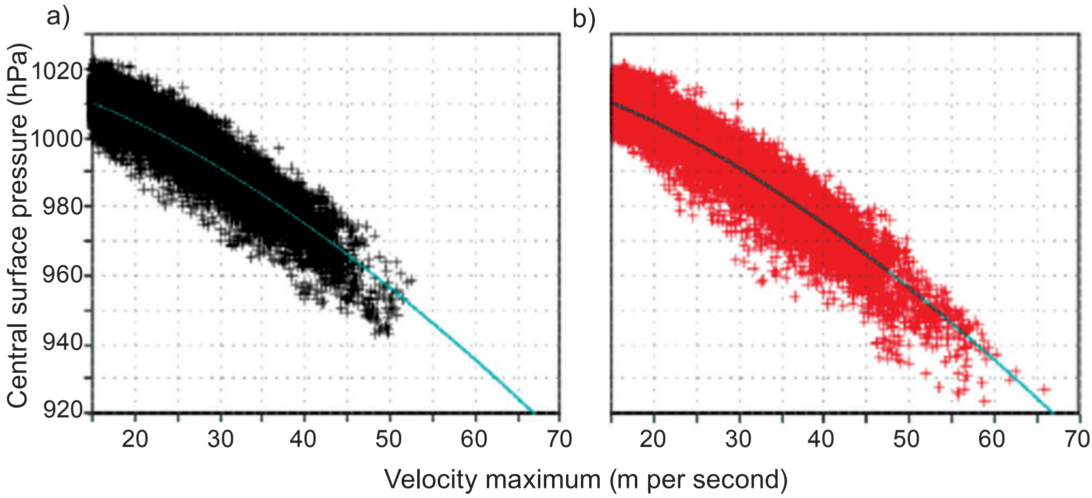


Figure 2.29 Relationship between central surface pressure (hectopascals, hPa) and maximum tropical cyclone winds (m per second) simulated for (a) present-day climate and (b) future climate. The 20-km resolution atmospheric model is able to capture the observed non-linear relationship between central pressure and wind for present-day and future climates, with an increase in deeper central pressures and strongest winds in the future warmer climate (source: Oouchi et al. 2006)[84].

2.6 Summary

The surface climates of PICTs are dominated by the vast Pacific Ocean surrounding them. Until recently, the size of the area has severely limited observational coverage of the surface climates of the region, i.e. collection of the long-term information necessary to detect changes in present-day climatic conditions. The surface climate of the tropical Pacific Ocean is significantly modulated by the major source of interannual global climate variability, ENSO. The two phases of ENSO produce distinct and different temperature and rainfall anomalies that affect Pacific islands. Tropical cyclones, which are the most destructive of the world's weather systems, significantly impact island communities north and south of ~ 5–10° latitude of the equator through high winds, heavy rainfall and storm surges. The usual locations of tropical cyclones are also modulated by ENSO events. Superimposed on these interannual sources of climate variability are slower (decadal time scale) modulations of Pacific climate and the strength of ENSO climate anomalies associated with the IPO/PDO.

The surface climate of the tropical Pacific Ocean is already showing evidence of significant changes with the relatively modest global warming observed to date. Temperatures are significantly warmer in large parts of the tropical Pacific, although there is considerable spatial variability in the observed pattern of warming and some regions have not warmed significantly. There have also been some significant observed changes in rainfall that are related mainly to changes in the position of the SPCZ. Unlike in other ocean basins, there is as yet no evidence for significant changes in either the frequency or intensity of tropical cyclones in the Pacific.

There are several limitations in projecting how the surface climate of the tropical Pacific will continue to change with continued global warming. These limitations include the inadequacies of the existing climate models for the tropical Pacific Ocean region, the lack of reliable model projections at the small scale required for Pacific island communities and, of course, uncertainty in projecting how future concentrations of greenhouse gases will change. The latter depends on the now extremely urgent requirement for global leadership and responses to drastically reduce the atmospheric concentrations of greenhouse gases to levels that will curtail the magnitude of future global warming within manageable limits[85].

Even with rapid and drastic actions, the world – including Pacific island communities – is committed to having to cope with a rapidly changing climate for the foreseeable future. Some components, such as sea level, will continue to rise for centuries (Chapter 3). Indeed, it is not simply a question of a change to a new climate regime, to which we could adapt, but that climate will continue to change for the coming centuries at least. The contributions of PICTs to the present global climate predicament are small, yet they will have to deal with many of the worst of the consequences.

Model projections for future climates of the tropical Pacific show that the emissions trajectory does not make much difference to the magnitude of climate changes in the near-term (to 2035), but the trajectory that is followed makes a big difference by the end of the century. Average changes in temperature for the tropical Pacific range from less than 1°C near 2035, to > 3°C by 2100. On average, precipitation in the tropical Pacific increases in proportion to the temperature increase, with decreases in the subtropical Pacific and some areas of the tropics depending on the season.

The large, natural multi-decadal variability of ENSO events in observations and models contributes to difficulties in attributing past ENSO fluctuations to a particular cause and in making credible future projections to changes in ENSO from the models. However, all global climate models show ENSO events continuing in the future, superimposed on the warmer average tropical Pacific SSTs, which the models project will increase more in the eastern equatorial Pacific than in the west. Using a combination of modelling tools that simulate some aspects of tropical cyclones, the most recent science suggests that while there may be fewer tropical cyclones in a warmer climate, those storms that do form are likely to be more intense.

The main projections from the analyses reported in this chapter are summarised below (see also **Table 2.6**).

- Temperatures are expected to continue to warm.
- Rainfall is likely to increase in the convergence zones near the equator and decrease in the subtropics. As the hydrological cycle intensifies with continued warming, the intensity of extreme rainfall events is also very likely to increase.

Even without significant changes in average rainfall, the intensity of future droughts associated with a given rainfall deficit are very likely to be greater than at present due to warmer temperatures and greater evaporation.

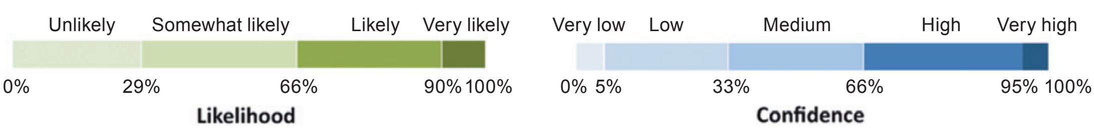
- There may be fewer tropical cyclones, but those that do occur are likely to be stronger and more destructive.
- Significant interannual surface climate anomalies associated with the El Niño and La Niña phases of ENSO are expected to continue.
- Although the magnitude of anthropogenic greenhouse gas emissions is tracking above the A2 emissions scenario, the differences in the climatic consequences between the two scenarios considered here are not expected to emerge until towards the end of the 21st century.
- Even with the 'low emissions trajectory' (B1), significant changes to the tropical Pacific climate are inevitable and may have devastating consequences when superimposed on other threats to Pacific livelihoods[86].
- Even the B1 trajectory may be too high to avoid dangerous climate change[87].
- Most of these projections are likely or very likely to occur, and can be made with high confidence (**Table 2.6**). The exceptions are for changes in the frequency and intensity of ENSO events and variability in the PDO, for which projections are made with low confidence. More detailed projections of future climate changes at the scale of individual PICTs requires both improvements in global climate models for the tropical oceans and more reliable downscaling of such models to finer spatial scales[v].
- Planning is already underway for the IPCC's Fifth Assessment Report (AR5) due to be published in early 2014[88]. This will be based on a new set of four time-dependent scenarios known as Representative Concentration Pathways (RCPs) and a new generation of global climate models that will be compared and integrated through the Coupled Model Intercomparison Project phase 5[89]. In addition, compared to AR4, AR5 will have an increased focus on the impacts of a changing climate on oceanic ecosystems. All these activities will, over the next few years, provide greater certainty about the consequences and impacts of a rapidly changing global climate for PICTs.

v This work is now being done for the tropical Pacific by the Australian Bureau of Meteorology, CSIRO and partners, under the Pacific Climate Change Science Programme; see www.cawcr.gov.au/projects/PCCSP

Table 2.6 Summary of projected changes to key Pacific surface climate variables relative to 1980–1999 values. Estimates of likelihood and confidence are provided for each projection, as defined in Chapter 1 (see key below).

Climate variable	2035 B1	2035 A2	2100 B1	2100 A2
Surface temperature (°C)	0.5–1.0	0.5–1.0	1.0–1.5	2.5–3.0
Sea surface temperature*	➢ SST changes similar to those for surface temperature, although slightly smaller in magnitude ➢ Spatial variation occurs in projected SST warming, with greater warming in the eastern than western equatorial Pacific and less warming in the southeast Pacific			
Rainfall	5–15% increase in equatorial regions	5–20% increase in equatorial regions	10–20% increase in equatorial regions	10–20% increase in wider region equatorial
	5–10% decrease in subtropics	5–20% decrease in subtropics	5–20% decrease in subtropics	5–20% decrease in subtropics
	➢ Extremes in wet and dry periods become more extreme ➢ Drought associated with decreases in rainfall become more intense due to warmer temperatures			
Tropical cyclones	➢ Total number of tropical cyclones may decrease ➢ No changes in usual locations ➢ Cyclones that do occur likely to be more intense			
ENSO events	➢ ENSO events continue as a source of interannual climate variability ➢ Unclear as to whether changes in frequency and intensity of ENSO will occur			
PDO-decadal variability	➢ PDO continues as source of decadal modulation in Pacific basin climate and ENSO events ➢ Unclear as to whether this will change			
Prevailing winds and circulation	➢ More vigorous hydrological cycle ➢ Enhanced Hadley circulation ➢ Expansion of area encompassed by 'tropics'			

* See Chapter 3 for more detailed projections of sea surface temperature.



2.7 Recommendations to reduce uncertainties in projecting the future climate

There has been a shortage of peer-reviewed science relating to small islands and climate change in recent IPCC assessment reports[90]. The measures that need to be taken to address this problem and reduce the uncertainties involved in projecting the future climate of PICTs are set out below[vi]:

- Commitment and support for new, high-quality surface weather observations for PICTs, and maintenance of existing facilities, to allow:
    - detection of the nature and significance of changing climates;
    - responses to these changes in natural and managed ecosystems to be characterised;
    - linking of island-scale climates of most relevance to PICTs to larger-scale climate observations now available through, for example, remote sensing; and
    - linking of changes in rainfall to variations in the river flow and groundwater regimes of islands.
- Improved high-resolution paleoclimatic reconstructions for the Pacific from, for example, annually banded corals, to allow:
    - detection of current changes and their significance; and
    - improved climate models.
- Identification of PICTs as a regional focus for climate change observations and modelling, including downscaling to island-specific scales to allow more rigorous assessment of local sensitivity and vulnerability to a changing climate[91] and specific inclusion of PICTs as a regional focus for the Coordinated Regional Climate Downscaling Experiment (CORDEX)[72].
- Improved climate modelling for the tropical ocean regions and, in particular, the capacity to realistically assess how the frequency and intensity of ENSO events and tropical cyclones are likely to change[25].
- Ongoing commitment to, and support for, improved weather forecasting and short-term seasonal climate outlooks for PICTs (e.g. tropical cyclones, ENSO), and appropriate, accessible warning systems for severe weather events with associated support for disaster-recovery strategies[92].

vi The Pacific Climate Change Science Program CCiP report ('Climate Change in the Pacific: Scientific Assessment and New Research') now covers, in detail, many of the aspects of tropical Pacific climate change summarised in this chapter.

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