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Chapter 2
Climate of the Western Tropical Pacific and East Timor

Summary

- The main features of the climate of the Pacific and East Timor are detailed using climate data from weather and ocean observing stations in the PCCSP region, together with regional and global gridded data based on in-situ and satellite data. In some instances observational data have been used in conjunction with computer models to produce spatially and temporally complete datasets.
- Many high-quality, multi-decadal observations have been recorded at sites across the region. However, many data have not yet been digitised and there are significant gaps in coverage. The quality of the digitised data has been carefully controlled and these data have been made accessible through new tools developed by the PCCSP. There are few multi-decadal sea-level records in the region.
- In broad terms, there are significant regional variations in mean climate, with different temperature and rainfall regimes in the eastern and western PCCSP region, with further differences between countries near the equator and those closer to the sub-tropics.
- Regions of strong convective rainfall exist in the equatorial Pacific in two bands: the Intertropical Convergence Zone (ITCZ), just north of the equator and the South Pacific Convergence Zone (SPCZ) from near the Solomon Islands to east of the Cook Islands.
- The ocean and atmosphere interact strongly in the region. Large-scale atmospheric circulation patterns drive both the ocean currents and temperature patterns; the ocean in turn also affects atmospheric winds, temperatures and rainfall.
- Most Partner Countries have weak seasonal variations in temperature but strong seasonal cycles in rainfall, with pronounced wet and dry seasons. Many are also affected by tropical cyclones, storm surges and other extreme events which can have devastating impacts on coasts and low lying regions.
- Typical seasonal variations of each country are described in terms of the main large-scale climate features of the region that exist throughout the year (e.g. the SPCZ and the ITCZ) or re-occur in certain seasons each year (West Pacific Monsoon), as well as other influences, such as sub-tropical weather systems and local effects on islands that are related to topographic features.
- Ocean currents driven by atmospheric circulation result in significantly warmer waters in the western Pacific than in the eastern Pacific.
- Regional differences in the mean sea-level distribution in the PCCSP region is largely a result of surface winds, with higher sea level in the western Pacific, particularly at latitudes 20–30°N and 20–30°S.

2.1 Introduction

This chapter describes the main features of the climate of the atmosphere and ocean of the region covering all PCCSP Partner Countries (Figure 1.1). Firstly, available observational data that are used throughout this publication to describe the climate of the region are summarised. The large-scale features of the atmospheric circulation and the ocean are then presented. The seasonal cycles of the climate in Partner Countries are discussed, as well as the climate features that determine this seasonality (e.g. the West Pacific Monsoon). Extreme events affecting the region are also described (e.g. tropical cyclones).

The Pacific Ocean is the largest ocean in the world, covering almost one third of the Earth's surface. It plays a dominant role in shaping the climate of not just the PCCSP region but also the entire planet. There are regional variations due to the presence of:

- Large land masses (e.g. the Maritime Continent, consisting of parts of Southeast Asia and the islands of Indonesia and the Philippines on the western equatorial edge of the Pacific) affecting climate on the regional scale.
- Mountainous regions in some countries that have large influences on their local climate.

Also, because some Partner Countries consist of small islands scattered across huge areas of ocean, very different climate conditions may exist within the one country.

The Pacific and the Maritime Continent are at the heart of some of the most prominent phenomena of the global climate system. The region is also characterised by extensive bands of large-scale wind convergence and associated rainfall features, such as the Intertropical Convergence Zone. Some regions experience very high seasonal rainfall variations associated with the West Pacific Monsoon. The waters in the western tropical Pacific are the world's warmest, and this and other unique features give rise to tropical cyclones that can produce storm surges and flooding. As is typical in tropical regions, many countries experience prominent and marked seasonal rainfall variations but little seasonal variation in temperature, and they are at significant risk of extreme events.

2.2 Data Record

Analysis of observed climate relies on long, complete, spatially extensive observational records of meteorological and oceanographic data. In the PCCSP region, land-based monitoring sites are sparse due to the small size of islands and the large distances between them. The Pacific Ocean covers a huge area. However only a small portion of it is routinely monitored. To overcome this, a significant effort has been made to collect all land and ocean observational records and, in the case of land-based records, digitise and homogenise them (apply quality control checks, Section 2.2.1.2). These records have been supplemented with satellite observations and combined with computer models of the climate system to provide a more complete spatial coverage over the region.

2.2.1 Atmospheric Data

2.2.1.1 Data from Observing Stations in the Pacific and East Timor Region

Many observed surface meteorological data (such as maximum and minimum air temperatures, rainfall, wind, air pressure, and sunshine measurements) from PCCSP Partner Countries are now available for analysis. These data have been recorded over long periods and significant efforts have been made to source, digitise, and quality-control these records. In the region, there are no station data available in the large ocean areas between most islands, and this represents a significant limitation to the data coverage. This gap can be filled with data from non-station data sources and such data are described in Section 2.2.2.

The earliest known observations began in 1851 at Malua, Samoa (Turner, 1861) and 1861 at Levuka, Fiji (Smythe, 1864). These observations were taken by European settlers, initially out of meteorological interest, but also to understand the weather and climate of the islands for the purposes of economic development. Over the next few decades, there was gradual growth in the numbers of observation stations as industry expanded in the new colonies.

The 1937 International Meteorological Conference in Wellington, New Zealand established the 'South-West Pacific' as the Fifth International Meteorological Organization region. Subsequent World War II developments provided the catalyst for rapid development of meteorological services in the region. Civil aviation, weather forecasting, tropical cyclone warnings and further industry development (e.g. agriculture, forestry, and mining) provided the main impetus for sustained meteorological services development in the region (Krishna, 2009). In the former British colonies, formal meteorological practices were established by the New Zealand Meteorological Service and the Australian Bureau of Meteorology.

Summaries of observation records were generally sent to the home country of previous administrations, however many of the original paper records and books were retained in the country of origin. A key present-day issue for the Partner Countries is determining how many historical data remain in colonial archives in paper form. Recovering and digitising all available data is critical for understanding the past climate in the Partner Countries.

Many of the electronic records sourced by PCCSP scientists have been obtained from five datasets (see below). These include the Australian Bureau of Meteorology's ADAM, New Zealand's National Institute of Water and Atmospheric Research (NIWA) CLIDB database, the National Oceanographic and Atmospheric Agency (NOAA) Global Summary of the Day, and Partner Country CliCom databases. Data have also been collected from Partner Country spreadsheets.

Australian and New Zealand datasets containing Partner Country data are sourced from original observation registers designed in those countries. In some cases monthly summaries, which have been transcribed by hand, have been used in lieu of the original registers. Office procedures in place at the time ensured that the summaries provide a reliable facsimile of the original record. In most Partner Countries these office procedures have continued to this day. NOAA datasets are captured from the World Meteorological Organization Global Telecommunications Network. This network transmits meteorological reports worldwide immediately after the observation is made.

Pre-independence paper records from Papua New Guinea, Nauru and the Solomon Islands, which were transferred to Australia, have been entered into the Australian Bureau of Meteorology ADAM database. These observation booklets and rainfall registers are kept in the National Archives of Australia in Victoria. ADAM contains the most complete pre-independence archive for Papua New Guinea, Nauru and the Solomon Islands. Likewise, the NIWA CLIDB database is the most extensive pre-independence archive for the Cook Islands, Fiji, Kiribati, Niue, Samoa, Tonga, Tuvalu and Vanuatu. These countries were formerly part of the New Zealand region of responsibility. Pre-independence data stored in CLIDB have been observed and recorded under New Zealand meteorological practices. An exception to this is where more recent Partner Country data have been captured from the Global Telecommunications Network and have not passed through a quality control process. NOAA datasets have been sourced from Network message formats and have only passed a coarse level of quality control. The PCCSP has sourced NOAA data for Palau, Federated States of Micronesia, Marshall Islands and East Timor.

In the 1990s, Fiji, Solomon Islands, Papua New Guinea and Vanuatu began digitising records into their CliCom databases. The Fijian and Papua New Guinean Meteorological Services have excellent paper archives which contain all known observation booklets and registers which remained in-country.

The CliCom Climate Data Management System[1] was introduced in 1985 by the World Meteorological Organization (WMO), with the assistance of NOAA to assist small countries with climate data management. Over many decades CliCom began to lose reliability as hardware improved and operating systems changed. Many backup files were damaged and the loss of historical electronic data was notable in a number of Partner Countries. Only Solomon Islands and Fiji still maintain operational CliCom systems.

Following the breakdown of the CliCom database, many Partner Countries began using electronic spreadsheets. Spreadsheets are a user-friendly application to record data and produce reports and graphs; however they are poor data storage systems. The variety of spreadsheet designs across and within countries makes retrieval and transfer of data difficult as does the fact that spreadsheets are normally held on individual computers. They have little or no data validation tools, a high probability of data duplication, and are normally accessed by a single user. Databases, on the other hand, have built in data validation and checking, can be accessed by multiple users and modifications are made using proper programming techniques. In addition, many different reports can be generated from the same data.

The PCCSP has developed a climate database management system known as CliDE (Climate Data for the Environment). During 2010 and 2011, CliDE was offered to Partner Country national meteorological services and installed in all the Partner Countries except Nauru (which does not have a national meteorological service). CliDE provides quality controlled data and reports. The Partner Country national meteorological services now have a robust, secure and structured climate database management system to assist in the provision of services to government, researchers and consumers.

For most Partner Countries, there remains a large volume of non-electronic records of observations stored in-country. In addition there are a number of overseas repositories containing images of records. Online repositories include the British Atmospheric Data Centre[2] and NOAA[3]. Repositories holding paper records include the Portuguese archives in Lisbon, East Timor archives in Dili and the Noel Butlin Archives in Australia[4]. Many of the Partner Countries would benefit from a digitising program to rescue the paper records. Table 2.1 shows the known extent of digitised and non-digitised data.

Table 2.1: Meteorological station data available in PCCSP Partner Countries for digitised climate data and rainfall only stations, and known data in manuscript form not yet digitised (climate and rainfall only). For each country the total number of available stations is given, as is the range of record length (in years) at these stations, and the total number of years of data available is in parentheses. For example, the Cook Islands has digitised climate data from 17 stations with record lengths from 1 to 71 years, with 471 total years of data.

Country	Climate digitised[5]		Climate paper		Rainfall digitised[6]		Rainfall paper	
	Stations	Range of years (total years)	Stations	Range of years (total years)	Stations	Range of years (total years)	Stations	Range of years (total years)
Cook Islands	17	1 - 71 (471)	43	1 - 11 (103)	42	1 - 89 (988)	13	1 - 22 (211)
East Timor	21	1 - 53 (139)	48	2 - 28 (770)	22	1 - 53 (236)	73	2 - 26 (963)
Fiji	72	1 - 80 (1959)	276	1 - 103 (2936)	163	1 - 80 (5276)	146	1 - 80 (5015)
Federated States of Micronesia	10	31 - 65 (375)			11	1 - 65 (396)		
Kiribati	23	1 - 61 (710)	65	1 - 6 (93)	68	1 - 62 (1555)	92	1 - 32 (265)
Nauru	3	6 - 26 (39)			3	1 - 37 (64)		
Niue	5	5 - 92 (147)	31	1 - 1 (31)	10	2 - 92 (194)		
Palau	1	31 - 31 (31)			2	1 - 63 (64)		
Papua New Guinea	98	1 - 34 (818)	402	1 - 32 (934)	677	1 - 75 (8767)	2636	1 - 37 (3425)
Marshall Islands	7	34 - 66 (325)			7	34 - 37 (254)		
Samoa	3	1 - 58 (83)	119	1 - 35 (460)	28	1 - 60 (508)	150	1 - 59 (1823)
Solomon Islands	36	1 - 36 (512)	64	1 - 36 (420)	80	1 - 53 (994)	51	1 - 42 (460)
Tonga	14	3 - 76 (464)	106	1 - 2 (107)	19	2 - 78 (538)	1	89 - 89 (89)
Tuvalu	5	4 - 60 (217)	39	1 - 33 (353)	10	4 - 83 (525)	11	8 - 38 (272)
Vanuatu	15	6 - 29 (309)	22	1 - 38 (265)	31	2 - 61 (703)	68	1 - 66 (908)

[1] http://www.wmo.int/pages/prog/wcp/wcdmp/clicom/index_en.html

[2] http://badc.nerc.ac.uk/browse/badc/corral/images/metobs/pacific

[3] http://docs.lib.noaa.gov/rescue/data_rescue_portuguese.html

[4] http://www.archives.anu.edu.au/nbac/html/index.php

[5] 'Climate' means those weather stations which record temperature and rainfall observations. Note: the definition of 'Climate' differs within the Partner Countries. In the former New Zealand region of responsibility 'Climate' refers to stations where one observation within a 24-hour period, normally at 0900 hrs local time is conducted. Stations where multiple observations are conducted are referred to as 'Synoptic-climate' or 'Synoptic'. In Australia and in the former Australian region of responsibility 'Climate' refers to both single and multiple observation stations.

[6] 'Rainfall' means those weather stations that record rainfall only.

2.2.1.2 Homogenised Observed Station Data

Observed climate variables sometimes show sudden shifts in the average values or variability. Not all of these shifts are caused by real changes in climate. Non-climate related shifts can be due to changes in instrumentation, observation site, surrounding environment and observation practices, or other factors. The aim of climate data homogenisation is to adjust data if necessary, so that all variations in the data series are caused by real changes in the climate, and not due to changes in the way the data have been recorded.

The most recent effort at homogenising Partner Country records was undertaken by the PCCSP and respective national meteorological services who received training in data homogenisation at a PCCSP workshop in June 2010. The homogenisation methodology employed followed practices of the Australian Bureau of Meteorology for Australian data. The software included RHtest (Wang et al., 2007; Wang, 2008a; Wang, 2008b; Wang et al., 2010). RHtest is an application run in 'R', a programming language and environment for statistical computing and graphics. R is freely available via the Internet, therefore ideal for use in developing countries. RHtest includes functions to detect potential shifts in a candidate (or base) climate series. Further information on RHtest is available at http://cccma.seos.uvic.ca/ETCCDMI/software.shtml.

Approximately 60 monthly rainfall, 30 maximum and minimum air temperature and six mean sea-level pressure station records were homogenised between 2010 and 2011 by the PCCSP.

2.2.1.3 Pacific Climate Change Data Portal

A data portal (http://www.bom.gov.au/climate/pccsp/) has been developed to present historical raw and homogenised data at the monthly time scale for rainfall, maximum and minimum air temperature and mean sea-level pressure. PCCSP Partner Country data are presented in the portal along with data from neighbouring Australian, New Zealand, Tokelau, Pitcairn Islands, French and US Territories to provide as detailed coverage as possible for the region.

Within the portal (Figure 2.1), users are able to view trends in historical climate data on annual and seasonal time scales over any period of interest where data are available. Users are also able to view data averages and running averages over three to 15 year periods.

2.2.1.4 Datasets on Global or Regional Grids

While data from individual observing stations can be used for detailed studies on local climate, the spatial inconsistency and the relatively small number of stations across the huge PCCSP region pose significant problems for many types of analysis. Many spatially and temporally uniform datasets have been created to complement the in-situ observations. These gridded data are easier to analyse and display, and overcome many of these problems. However, these gridded data usually lose much of the small-scale local information available from individual station data. Outputs from global climate models are similarly gridded and as a result they can be more easily compared to gridded observational datasets.

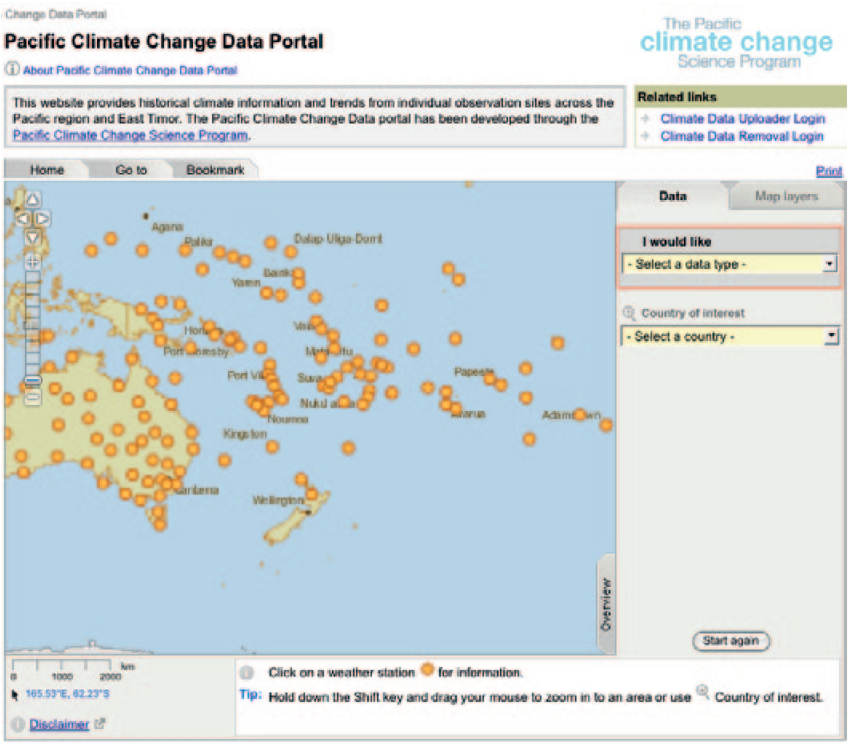


Figure 2.1: Locations of surface observed data stations in the Pacific Climate Change Data Portal http://www.bom.gov.au/climate/pccsp/.

A variety of gridded data products (Table 2.2) have been used for the analysis of past and present climate in the PCCSP region. It is important to distinguish between two types of gridded data products. The first type statistically interpolate the available in-situ observations to provide a dataset that has no gaps in space and time. Examples of this are the precipitation datasets described below. The second type, is a hybrid of numerical weather prediction model and observations known as reanalysis datasets.

Reanalysis products are produced by numerical weather prediction models that are constrained by available observations and where the interpolation is a result of the physical and dynamical processes embedded within the model. For surface air temperature, mean sea-level pressure and the surface winds the following reanalysis products were used:

- European Centre for Median-Range Weather Forecast (ECMWF) 40-year reanalysis (ERA-40; Uppala et al., 2005)
- Joint National Centres for Environmental Prediction and Department of Energy reanalysis (NCEP/DOE R-2; Kanamitsu et al., 2002)
- Japanese 25-year reanalysis (JRA25; Onogi et al., 2007).

The ERA-40, NCEP/DOE R-2 and JRA25 projects use state-of-the-art analysis/forecast systems to perform data assimilation using past data.

ERA-40 consists of a set of global analyses describing the state of the atmosphere and land and ocean-wave conditions from mid-1957 to mid-2002. ERA-Interim is an interim reanalysis of the period 1989–present in preparation for the next-generation extended reanalysis to replace ERA-40. The ERA-Interim archive is more extensive than that for ERA-40, e.g. the number of pressure levels in the atmosphere is increased from 23 to 37 and additional cloud parameters are included. ERA-Interim products are also available, including several products that were not available for ERA-40. It is therefore particularly useful for more high resolution applications.

The Global Precipitation Climatology Project (GPCP; Adler et al., 2003) and Climate Prediction Centre Merged Analysis of Precipitation (CMAP; Xie and Arkin, 1997) are generally considered to provide a more accurate representation of global rainfall than current reanalysis products (Beranger et al., 2006; Bosilovich et al., 2008), and these were used to represent the observational precipitation record.

Both GPCP and CMAP datasets merge satellite data from a number of satellite sources, and use rain gauge data over land. Details on the component datasets as well as the method used to merge these data are provided by Huffman et al. (1997) and Adler et al. (2003) for GPCP, and Xie and Arkin (1997) for CMAP.

Table 2.2: Overview of the gridded atmospheric datasets used in the PCCSP climate analysis

Climatological fields	Gridded dataset		Source	Reference
	Name	Origin		
Precipitation	CMAP version 0802	CPC Merged Analysis of Precipitation	http://www.cpc.ncep.noaa.gov/products/global_precip/html/wpage.cmap.html	Xie and Arkin (1997)
	GPCP version 2	Global Precipitation Climatology Project	http://precip.gsfc.nasa.gov	Adler et al. (2003) and Huffman et al. (1997)
	TRMM	Tropical Rainfall Measuring Mission	http://trmm.gsfc.nasa.gov/	Kummerow et al. (1998 and 2000)
	CRU	Climate Research Unit Rainfall Climatology	http://www.cru.uea.ac.uk/cru/data/	New et al. (1999)
Winds, mean sea-level pressure and surface air temperature	ERA-40	ECMWF 40-year reanalysis	http://www.ecmwf.int/products/data/archive/description/e4/index.html	Uppala et al. (2005)
	NCEP-DOE version 2	National Centre for Environmental Prediction and Department of Energy Reanalysis	http://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis2.html	Kanamitsu et al. (2002)
	JRA25	Japanese 25-year Reanalysis	http://www.jreap.org/indexe.html	Onogi et al. (2007)
	ERA-Interim	ECMWF interim Reanalysis	http://www.ecmwf.int/research/era/do/get/era-interim	Dee et al. (2011)

Differences between GPCP and CMAP datasets are primarily due to differences in the input data and in the methodologies by which the various input data are merged. Over large land areas, these differences are negligible.

There are large-scale regional differences over the oceans. Over many of the oceanic tropical areas, CMAP has 20–50% more rain when compared to GPCP. Possible explanations are:

1. Differences in the use and manipulation of input data between CMAP and GPCP. Both CMAP and GPCP estimate rainfall over the oceans using microwave emissions calculated from brightness temperature measurements in various frequencies. CMAP also uses the Special Sensor Microwave Imager (SSM/I) scattering-based estimates over oceans while GPCP does not.
2. GPCP adjusts the Geostationary Operational Environmental Satellite precipitation index to the SSM/I, while CMAP does not.
3. The satellite estimates throughout the tropical oceans are adjusted by the atoll rain gauge reports in CMAP, while no such adjustment is made in GPCP.

Different analysis methods attempt to produce the most accurate precipitation data possible. However, there are still errors in all products, and the differences between their values reflect the uncertainties in each. Despite the differences in the means as discussed, CMAP and GPCP have considerably better agreement in departures from the climatological monthly means (anomalies), and temporal correlations between the two datasets are high.

Two other rainfall gridded datasets are used. The Tropical Rainfall Measuring Mission (TRMM) is a joint satellite mission between the National Aeronautics and Space Administration (NASA) and the Japan Aerospace Exploration Agency (JAXA) designed to monitor and study tropical rainfall. TRMM, during its mission and broad sampling footprint between latitudes 35°N and 35°S, is providing some of the first detailed and comprehensive data on the distribution of rainfall over vastly under-sampled oceanic and tropical continental regimes. It produces maps and gridded rainfall products on fairly fine spatial resolution (0.25° x 0.25°) and frequency (near-real-time; 1 to 3 hourly snapshots). Version 6 of the TRMM product used in this report covers a time period from 1998 until 2011.

The Climate Research Unit of the University of East Anglia, UK, has assembled a number of global gridded datasets for rainfall and temperature. All of these are land-only since they are based on station observations. One of their most recent products is the CRU TS2.1 dataset which includes a global land-only precipitation dataset on fairly fine spatial resolution (0.5° x 0.5°) covering the time period 1901–2006 in monthly time steps. This dataset is often used in the validation of global climate models and downscaling simulations.

2.2.2 Ocean Data

2.2.2.1 Datasets on Global or Regional Grids

Similar to atmospheric data from observing stations, ocean data are generally measured at irregularly spaced points in space and time. In order to create a more consistent coverage and to make these data easier to analyse, several systems have been used to generate datasets on regular grids and at regular time intervals.

The World Ocean Atlas 2005 (Locarnini et al., 2006; Antonov et al., 2006; Garcia et al., 2006) provides objectively analysed (i.e. based on the synthesis of multiple data sources) monthly, seasonally or annually averaged ocean climatologies displayed on a 1° x 1° geographic grid (Table 2.3). Variables include temperature, salinity, oxygen and nutrients at various depths. The CARS2006 from CSIRO (Ridgeway et al., 2002) provides surface and sub-surface temperature and salinity data for the tropical Pacific and Southern Hemisphere regions. The dataset uses all available historical sub-surface ocean property measurements (pre 2006) from ships and autonomous floats to provide climatology of various ocean properties.

Several sea-surface temperature (SST) datasets are used in this report (Table 2.3). Using different datasets allows the robustness of trends and variability to be evaluated. The UK Met Office Hadley Centre's dataset HadISST (Rayner et al., 2003) provides a 1° x 1° SST and sea-ice reconstruction (i.e. where modern measurements are used to extend sparse historical observations over a global domain) from 1870 to the present. Source data for SST come from the UK Met Office Marine Data Bank and the Comprehensive Ocean-Atmosphere Data Set (ICOADS). Interpolation provides a spatially continuous dataset. The related dataset HadSST2 (Rayner et al., 2006) product provides SST averaged in 5° x 5° grid cells. Unlike HadISST, HadSST2 does not perform any spatial interpolation, thus no SST estimates are provided in areas with poor data coverage. Also, based on the ICOADS dataset, the ERSST provides 2° x 2° resolution interpolated gridded output (Smith et al., 2008). The Kaplan et al. (1998) SST dataset is based on slightly different base datasets to the previously noted reconstructions with spatially interpolated data available at a 5° x 5° resolution.

QuikSCAT (NASA Quick Scatterometer, 2000) uses satellite scatterometer data to produce a high resolution database of ocean winds at 10m height. The NOAA Climate Forecast System Reanalysis (Wang et al., 2010) provides hourly surface wind and atmospheric pressure fields for 1979–2010. The NOAA Wave Watch III model (Tolman, 2002) provides surface wave field information from 1997 to the present.

Two new datasets using historical and Argo float data provide information on the temporal evolution of ocean temperature (1960–2007), and salinity (1950–2008), (Durack and Wijffels, 2010) down to about 2000 m depth. The multi-parametric analysis technique provides an estimate of the long-term trend, seasonal cycle and components of the temporal evolution related to the El Niño-Southern Oscillation and other patterns of climate variability.

The LDEOv2009 CO_2 partial pressure (pCO_2) dataset includes almost five million measurements of in-situ pCO_2, SST and sea-surface salinity spanning from 1957 to 2009. Data are available at 4° x 5° resolution (Takahashi et al., 2010).

2.2.2.2 Sea Level

Sea level varies on a wide range of time and space scales. While comprehensive sea-level observations to fully characterise this variability are not available, there has been a significant improvement in sea-level data from in situ and satellite observations over recent decades.

High quality satellite altimeter observations (from the TOPEX/Poseidon, Jason-1 and OSTM/Jason-2 satellites) of sea level between latitudes 66°S and 66°N are available from late 1992 to the present. The data are readily available (Table 2.3).

In situ sea-level observations made with tide gauges are also available at a number of locations in the PCCSP region. During the 1970s and 1980s, the number of gauges installed in the Pacific increased as part of efforts to understand the evolution of El Niño-Southern Oscillation events. With an increasing focus on sea-level rise in the late 1980s, projects like the South Pacific Sea Level and Climate Monitoring Project (http://www.bom.gov.au/pacificsealevel/) have resulted in an improved quality of individual gauges and of the network as a whole, using modern instrumentation with rigorous datum control (most recently with continuous global positioning system instruments). The publicly available in-situ sea-level datasets are detailed in Table 2.3.

The near global coverage of the altimeter data sets available since 1993 has provided information about the large-scale spatial distributions of sea level over the oceans. This information has been combined with coastal and island measurements of sea level over a longer time period using statistical techniques to estimate pre-altimeter period sea level from tide-gauge measurements alone on a 1° x 1° grid since 1950 (Church et al., 2004, 2006; called reconstructed sea level) and global averaged sea level since the late 19th century (Church and White, 2006, in press).

2.2.3 Climatological Periods

A period of 30 years is recommended for calculating climatological averages (Guttman, 1989; WMO, 1984), and the WMO's reference period is 1961–1990. This period is generally used here for calculating climate averages from Partner Country sites. However, 1971–2000 is used for many gridded analysis products, as there is greater availability of satellite data after the late 1970s, and this period is covered by most of the datasets listed in Tables 2.2 and 2.3. This period also closely matches the period used in Chapter 5 for evaluating climate models. It is important to note that in some instances the shorter time span of 1980–1999 had to be used when the data (e.g. the CMAP rainfall analysis) did not cover the full 30 years.

For some variables with small seasonal variations, only the annual averages are shown. In other instances, different seasonal averages are given, most often the four seasons (defined as December to February, March to May, June to August and September to November), as well as the tropical wet and dry seasons (November to April and May to October, the respective season depending on the hemisphere).

Table 2.3: Overview of the ocean datasets used in this publication (see text for details).

Climatological fields	Gridded dataset		Source	Reference
Sea-surface temperature, sea-surface salinity, nitrate concentration	WOA05	World Ocean Atlas 2005	http://www.nodc.noaa.gov/OC5/WOA05/pubwoa05.html	Locarnini et al. (2006), Antonov et al. (2006), Garcia et al. (2006)
Wind stress	QuikSCAT		ftp.ifremer.fr/products/gridded/mwf_quiskcat/data	NASA Quick Scatterometer (2000)
Wind stress, surface pressure	NOAA Climate Forecast System Reanalysis (CFSR)		http://cfs.ncep.noaa.gov/cfsr/	Wang et al. (2010)
Surface gravity wave height, direction, period	NOAA Wave Watch III		http://polar.ncep.noaa.gov/waves/	Tolman (2002)
Sea-surface temperature	HadISST	Hadley Centre SST dataset	http://badc.nerc.ac.uk/data/hadisst/	Rayner et al. (2003)
	HadSST2		http://hadobs.metoffice.com/hadsst2/	Rayner et al. (2006)
	ERSST	NOAA Extended Reconstructed SST	http://www.esrl.noaa.gov/psd/data/gridded/data.noaa.ersst.html	Smith et al. (2008)
	Kaplan Extended SST V2		http://www.esrl.noaa.gov/psd/data/gridded/data.kaplan_sst.html	Kaplan et al. (1998)
Temperature, salinity	CARS2006	CSIRO Atlas of Regional Seas	http://www.marine.csiro.au/~dunn/cars2006/	Ridgeway et al. (2002), Dunn and Ridgeway (2002)
			Contact authors for data availability	Durack and Wijffels (2010)
pCO_2	LDEOv2009	Lamont-Doherty Earth Observatory (LDEO) database	http://cdiac.ornl.gov/oceans/LDEO_Underway_Database/	Takahashi et al. (2010)
	Description/Institution		**Source/Reference**	
Sea-level altimetry	Measured along satellite ground tracks, from NASA		http://podaac.jpl.nasa.gov/DATA_CATALOG/topexPoseidoninfo.html; http://podaac.jpl.nasa.gov/DATA_CATALOG/jason1info.html; http://www.nodc.noaa.gov/SatelliteData/Jason2 or ftp://avisoftp.cnes.fr/AVISO/pub/jason-2/gdr_t/	
	Processed versions of this dataset		http://www.cmar.csiro.au/sealevel/sl_data_cmar_alt.html	
	Gridded products of this and other data		ftp://ftp.aviso.oceanobs.com; http://www.cmar.csiro.au/sealevel/sl_data_cmar_alt.html	
In situ sea-level data	Australian National Centre Facility		http://www.bom.gov.au/oceanography/projects/spslcmp/spslcmp.shtml	
	Permanent Service for Mean Sea Level		http://www.psmsl.org/; Woodworth and Player, (2003)	
	University of Hawaii Sea-level Centre		http://uhslc.soest.hawaii.edu/uhslc/	

2.3 Current Climate in the Region: Atmosphere

This section describes the features of the average climate in the region. One aspect that clearly emerges is the close connection between patterns in temperature, the atmospheric circulation and rainfall. The oceans play a vital role in these interactions, as described in Section 2.6.

2.3.1 Surface Air Temperature

Surface air temperatures observed in the PCCSP region are dominated by several important geographical features. Firstly, in many Partner Countries air temperatures are largely determined by the surface temperatures of the surrounding Pacific Ocean because the relatively small land areas of most islands have little influence on the local energy budget. Secondly, seasonal temperature variations can be driven by the large land masses on the western edge of the Pacific (Australia, Asia and the Maritime Continent) that heat and cool at different rates to the oceans. The resulting land-ocean temperature contrasts drive the West Pacific Monsoon and other atmospheric and ocean circulations.

In the tropics, with the sun being almost directly overhead, the amount of incoming solar radiation (heat) is high and changes only very slightly throughout the year. As a result, air temperatures remain high and relatively constant throughout the year. Temperature variations between the seasons increase with distance from the equator, as does the impact of extra-tropical air masses on temperatures.

Whilst the sun is the main driver of temperatures in the PCCSP region, interactions between the atmospheric circulation and the ocean result in regional variations in air temperatures. These temperature variations affect the patterns of rainfall across the region and closely match the ocean temperatures. Air temperatures in the western tropical Pacific and Maritime Continent are on average several degrees warmer than in the eastern tropical Pacific due to winds affecting ocean temperatures (Section 2.6). The warm waters in the west are known as the West Pacific Warm Pool, and the cool waters in the east are called the Equatorial Cold Tongue (Figure 2.2). Over the West Pacific Warm Pool there is virtually no seasonal variation in air temperature and over much of the tropics temperatures in the coldest months are only a few degrees lower than the warmest months (Figure 2.3). Along the equator, year-to-year variations in temperature can be larger than the average variations between the seasons (Chapter 3).

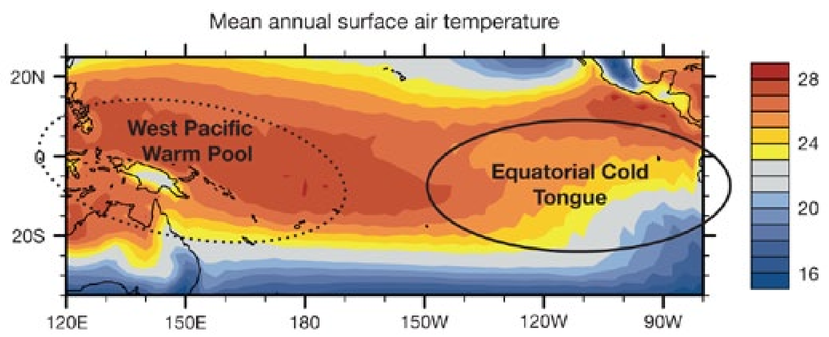


Figure 2.2: Annual average surface air temperature (ºC) in the equatorial Pacific. Approximate positions of two important features are indicated: the West Pacific Warm Pool and the Equatorial Cold Tongue in the east. Data from the NCEP2 reanalysis for the period 1971–2000.

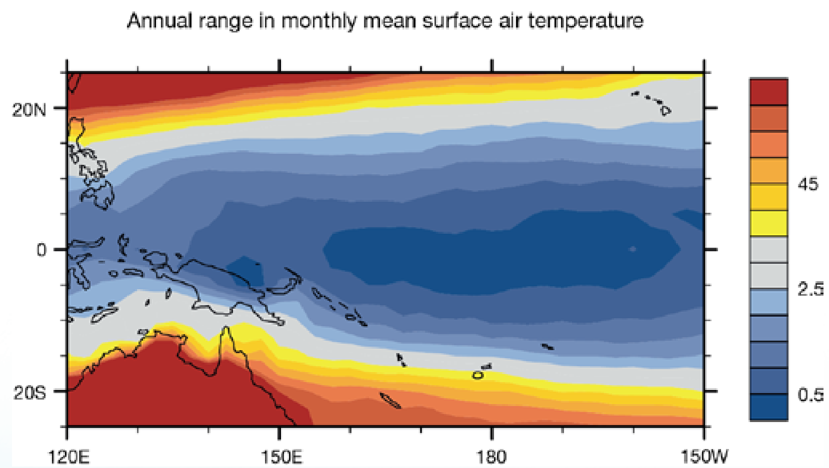


Figure 2.3: Range in surface air temperature (ºC) between average hottest and coldest months. Data from the NCEP2 reanalysis for the period 1971–2000.

2.3.2 Atmospheric Circulation

The imbalance of higher amounts of incoming solar radiation (heat) in tropical regions, compared to higher latitudes, results in the Earth's atmosphere and ocean circulations transporting heat toward the poles. The Earth's rotation causes changes in the direction of the air movement (the Coriolis effect), giving rise to complex atmospheric circulation patterns. Large land masses and topographic features also modify regional and local circulations. In the tropics, the major features of this atmospheric circulation include the Hadley Circulation (the major vertical movement of heated equatorial air and its north-south transfer into the mid latitudes), the Walker Circulation (the east-west circulation of air across the Pacific region), the trade winds, large-scale convection over the West Pacific Warm Pool and convergence zones, the West Pacific Monsoon, tropical cyclones and sub-tropical high pressure systems. Some of these can be seen in Figure 2.4. Others are shown in Figure 2.5 and Box 2.2.

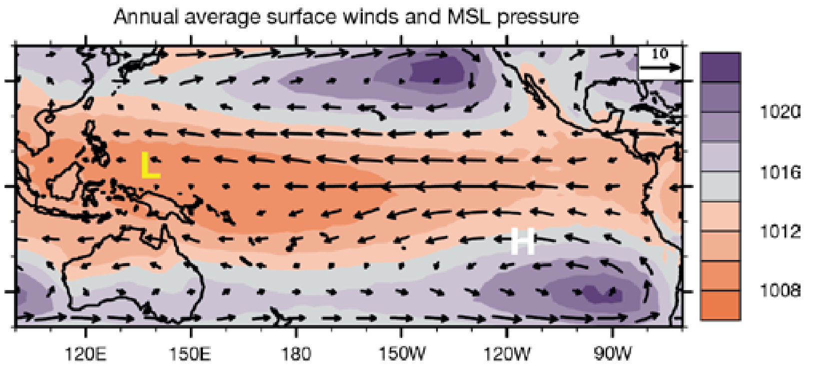


Figure 2.4: Annual average mean sea-level pressure (in hPa, purple=high pressure, red=low pressure), with semi-permanent pressure systems (H=high, L=low) and surface winds (in metres per second) indicated by the length of the arrows (the 10 metres per second reference arrow is shown in the top-right corner). Data from the NCEP2 reanalysis for 1971–2000.

Over areas where the ocean is relatively warm, rising air (convection) leads to the formation of clouds and produces rainfall. Surface winds tend to converge towards and feed into these convection areas. Strong convection occurs over the West Pacific Warm Pool so the surface trade winds blow from east to west across the Pacific along the equator, feeding into this large region of convection. Air rises high in the atmosphere over the Warm Pool and flows back from west to east in the upper atmosphere (the upper troposphere). This air descends in the eastern Pacific and over the Equatorial Cold Tongue, feeding the surface trade winds, thus completing the circulation. This is known as the Walker Circulation (Figure 2.5), and is one of the most prominent and important of the atmospheric circulations (Gill, 1982; Power and Smith, 2007).

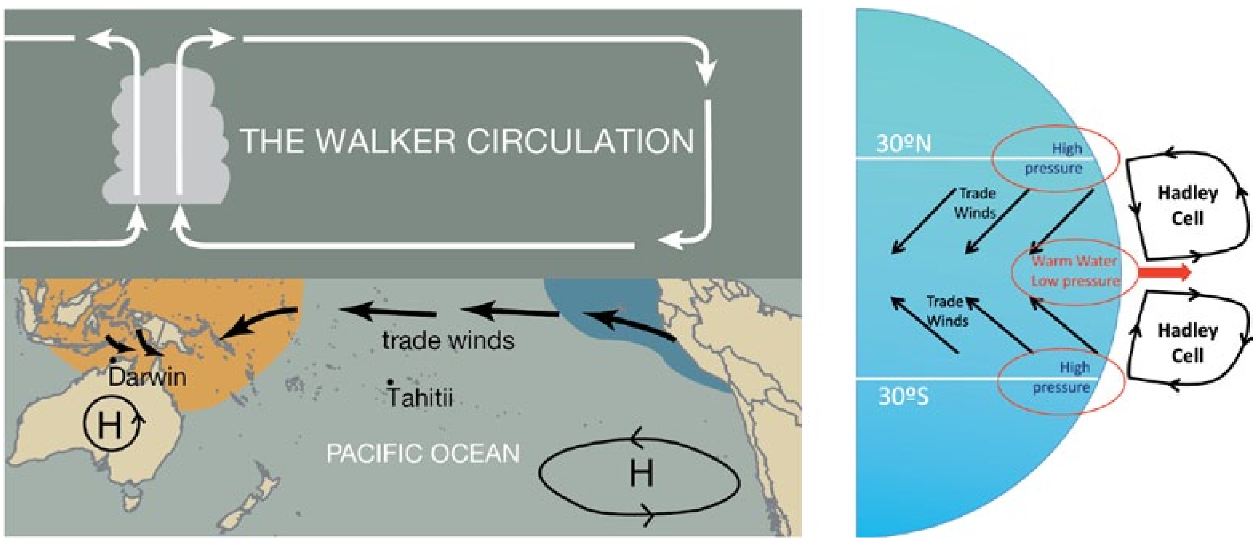


Figure 2.5: (Left) A schematic of the atmospheric Walker Circulation. Surface trade winds from the cold waters of the Equatorial Cold Tongue (blue) in the east blow across the Pacific, feeding into convection over the West Pacific Warm Pool (orange). Air rises and flows back to the east in the upper atmosphere to complete the circulation (Source: Australian Bureau of Meteorology). (Right) A schematic of the Hadley Circulation, with rising air at the equator, flowing poleward in the upper atmosphere, descending over the sub-tropics and flowing back to the equator as trade winds.

The Walker Circulation weakens during El Niño events and strengthens during La Niña events (Chapter 3), but in normal situations the equatorial trade winds are quite constant in both direction and strength. The Walker Circulation's rising air over the West Pacific Warm Pool creates a region of low pressure, while the descending air over the eastern Pacific (east of about 150°W) creates a region of higher pressure (Figure 2.4). When the Walker Circulation and equatorial trade winds change strength, the pressure in both regions usually changes at the same time. For example, if the Walker Circulation slows down, convection over the West Pacific Warm Pool decreases and so pressure increases, and subsidence over the eastern Pacific weakens and so pressure there tends to fall. This oscillation in air pressure across the region is known as the Southern Oscillation.

In addition to the Walker Circulation, rising convective air also feeds a north-south circulation, known as the Hadley Circulation. When the Walker Circulation strengthens, the Hadley Circulation tends to weaken and vice versa (Oort and Yienger, 1996). In the Hadley Circulation, the air that rises near the equator flows towards both poles in the upper troposphere, descending over the sub-tropics. The mean positions of this descending, dry air in both hemispheres determine the location of the more arid regions of the Earth.

This descending air leads to regions of high pressure around latitudes 30ºS and 30ºN (Figures 2.4 and 2.5). Within these regions lie the permanent South-east Pacific High and North-west Pacific High. These high pressure systems drive surface winds towards the equator and from the east, producing the predominantly south-east trade winds in the Southern Hemisphere, and north-east trade winds in the Northern Hemisphere. To the west of these permanent highs, high pressure systems migrate from west to east, modifying the local wind direction as they pass (Steiner, 1980). The intensity of these semi-permanent and migrating high-pressure systems determines the strength and direction of the predominantly north-east and south-east trade winds. Cold fronts and extra-tropical low pressure systems lie south of the high pressure zone and they can influence local weather when they approach.

The north and south descending branches of the Hadley Circulation and the resulting high-pressure systems generally move equatorward during the winter and spring and poleward during summer and autumn in response to changes in north-south temperature gradients with the seasons (Barry and Chorley, 1992). When the high-pressure systems move closer to the equator they can bring cooler conditions to a country as the more north-south winds bring cooler, extra-tropical air. They also bring drier conditions as the descending air associated with them is dry.

2.3.3 Rainfall

Many tropical areas receive very high annual rainfall as a result of moisture-laden winds. Evaporation and convection occur where the ocean surface is warm, and when the winds converge this convection is enhanced. In the Pacific and around the Maritime Continent there are several regions where these factors act together.

Patterns of average rainfall show that there are some differences between the December to February and June to August periods (Figure 2.6). However, some large-scale features remain year round where ocean temperatures remain warm and winds converge throughout the year. For example, over the West Pacific Warm Pool rainfall persists throughout the year. This warm reservoir of water provides the energy source for the extensive convective activity that drives the Walker Circulation and much of the Hadley Circulation in this region. This is a major moisture source for the heavy rainfall in the region, which averages more than 2–3 m per year (Delcroix et al., 1996). The South Pacific Convergence Zone (SPCZ) in the south-west Pacific, (Section 2.4.1) and Intertropical Convergence Zone (ITCZ) just north of the equator, (Section 2.4.2) are two other prominent features associated with high rainfall.

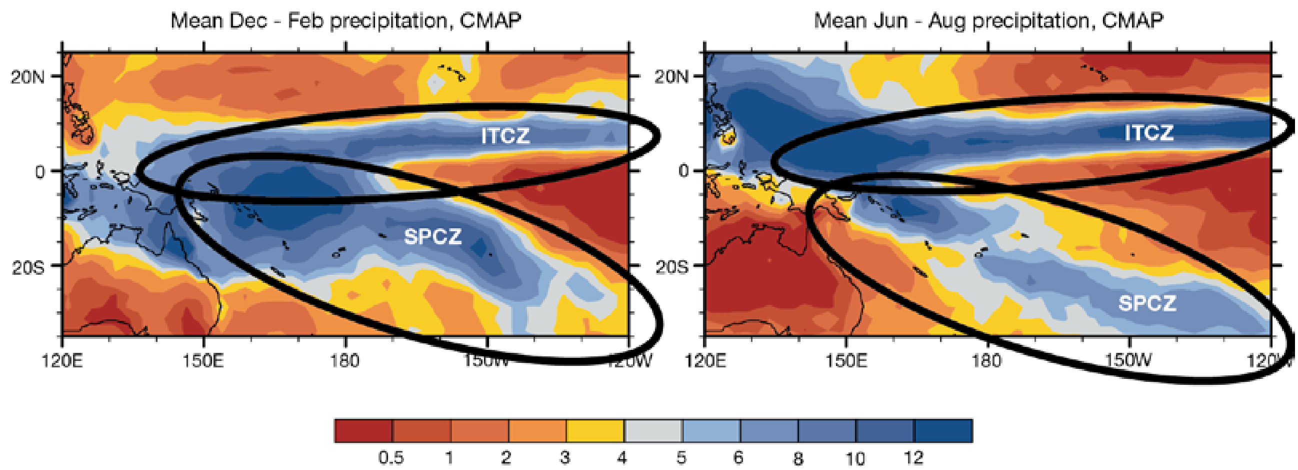


Figure 2.6: Average rainfall over the PCCSP region for December to February (left) and June to August (right), in mm per day (blue = wet and red = dry). Data from the Climate Prediction Center Merged Analysis of Precipitation dataset for years 1980–1999. The average ITCZ and SPCZ positions are indicated.

In the tropics, significant rainfall also comes from tropical cyclones and storms. Tropical cyclones are mostly confined to the latitudes 10º-25º in both hemispheres, where the ocean is warm enough to provide the energy for them to form and the Coriolis effect (related to the Earth's rotation) is large enough for them to spin-up. The contribution of tropical cyclones to Partner Country annual rainfall is generally around 5–10% in those countries affected by tropical cyclones, and up to around 15% in Vanuatu (Jiang and Zipser, 2010).

On some islands (including quite small ones) with significant topography the moist winds (such as the trade winds) are forced upwards, leading to condensation and enhanced rainfall. This can lead to very different local climates, depending on the location relative to the prevailing winds and the topography. In the sub-tropics, rainfall also comes from cold fronts that traverse the region, usually in the winter months.

The largest seasonal variations in rainfall are seen in regions strongly affected by the West Pacific Monsoon (Box 2.2 and Section 2.4.3) in the far west of the PCCSP region. In these countries the difference in rainfall between the wet season (during the monsoon) and the dry season is significant. For example, in Dili in East Timor, almost 80% of the total annual rainfall falls during the wet season of November to April.

In other countries there are some seasonal variations in rainfall, but smaller than in the region affected by the West Pacific Monsoon. Rainfall seasonality is associated with variations in the tropical cyclone seasons, the strength and position of the ITCZ and SPCZ, the position and strength of sub-tropical high pressure systems and cold fronts, and variations in the amount of available moisture as ocean temperatures vary. In those countries that have significant topography, rainfall will vary during the year depending on the strength and direction of winds, particularly the trade winds.

In all countries in the PCCSP region there is a significant difference in the regular rainfall that comes in the wet and dry seasons. The regularity of these seasons has a large impact on life in the region. This has led to long-held traditional knowledge of the seasons, reflected in the language of each of the peoples of the Pacific and East Timor. Each group has one or several names for the two seasons (Box 2.1), which usually refer to the wet and dry periods, the time of planting and harvesting crops, or the changes in the prevailing winds between the seasons.

Box 2.1: Indigenous Weather and Climate Terminology

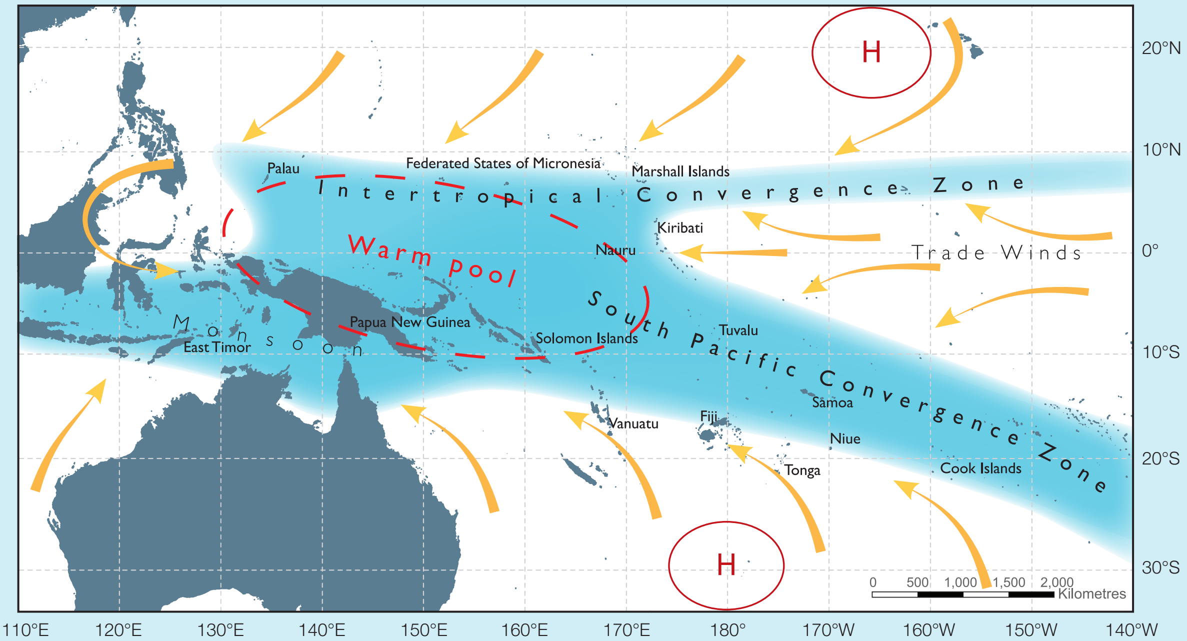
Having lived with the extremes of weather and climate for many generations, the indigenous peoples of the Pacific and East Timor have intimate knowledge of climate phenomena. Most indigenous languages have terms for these climate phenomena that affect their everyday lives. This has led to terminology for the wet and dry seasons in local languages of the peoples in the Pacific and East Timor.

Country		Wet season	Dry season
Cook Islands		Tuatau Mauu	Tuatau Maro
East Timor		Udanben	Bailoron
Federated States of Micronesia	Chuuk	Nukuchochun	Pwas
	Yap	Nuw	Yal
Fiji		Suasua	Mamaca
Kiribati		Ameang	Aumaiaki
Marshall Islands		Mejleb	Anńeńeańń
Nauru		Luai	Aré
Niue		Vahä Mafana	Vaha Mokomoko
Palau		Kemtimt	Sechal El Ongos
Papua New Guinea	Pidgin	Taim Bilong Ren	Taim bilong San
	Motu	Medu Ena Nega	Dina Ena Nega
Samoa		Vaipalolo / Taū Sūsū	Vaitoelau / Tau Mugalā
Solomon Islands		Komburu	Ara
Tonga		Fa'ahita'u 'Uha	Fa'ahita'u La'ala'a
Tuvalu		'Tau 'Moko	Tau Yela
Vanuatu (based on planting season)		Nowa mate (Harvesting)	Nowa Mourii (Planting)

2.4 Large-Scale Climate Features

As discussed in preceeding sections, there are several very important features of the climate system in the Pacific and East Timor region that influence the mean climate and seasonal climate variations in Partner Countries (Box 2.2). Some of these features exist throughout the year while others have pronounced and regular seasonal cycles. They are described here, and the average seasonal cycles of the climate in each Partner Country are briefly described and explained in terms of these climate features.

Box 2.2: Main features influencing the climate of the PCCSP region



The average positions of the climate features in November to April. The yellow arrows show near-surface winds, the blue shading represents the bands of rainfall (convergence zones with relatively low pressure), and the red dashed oval indicates the West Pacific Warm Pool. H represents the typical positions of moving high pressure systems.

West Pacific Monsoon

This moves north to mainland Asia during the Northern Hemisphere summer and south to Australia in the Southern Hemisphere summer. The seasonal arrival of the monsoon usually brings a switch from very dry to very wet conditions. It affects countries in the far western Pacific and the Maritime Continent.

Intertropical Convergence Zone

This band of high rainfall stretches across the Pacific just north of the equator and is strongest in the Northern Hemisphere summer. It affects most countries on, or north of, the equator.

South Pacific Convergence Zone

This band of high rainfall stretches approximately from the Solomon Islands to east of the Cook Islands. It is strongest in the Southern Hemisphere summer and affects most countries in the South Pacific.

Sub-Tropical and High Latitude Influences

These include sub-tropical high pressure systems and associated south-east and north-east trade winds, and cold fronts.

2.4.1 South Pacific Convergence Zone

The SPCZ extends northwest-southeast in a diagonal line from near the Solomon Islands (0°, 150°E) to the south-eastern Pacific (around 30°S, 120°W) (Figure 2.6). It is composed of two parts (Trenberth, 1976; Kiladis et al., 1989; Vincent, 1994): the west-east-oriented western part attached to the West Pacific Warm Pool, which interacts with the West Pacific Monsoon; and the diagonally-oriented portion further to the east that extends into the sub-tropics. The Partner Countries affected by the SPCZ are the Solomon Islands, Tuvalu, Vanuatu, Fiji, Tonga, Samoa, Niue and the Cook Islands, and in some years Nauru and Kiribati. See Table 2.4 for further details.

The SPCZ forms in the region of convergence between the north-east trade winds and the south-easterly circulation ahead of anticyclones from the Australian region (Trenberth, 1976; Streten and Zillman, 1984). The western, tropical portion of the SPCZ lies over the warmest sea-surface temperatures of the West Pacific Warm Pool, while the eastern portion is dominated by interactions with troughs in the mid-latitude circulation, which contributes to its diagonal orientation (Kiladis et al., 1989; Vincent, 1994).

The seasonal cycle of rainfall in the SPCZ region can be seen in time-latitude plots of zonal mean (i.e. east-west) rainfall across the region (Figure 2.7). The SPCZ is most clearly defined with strongest rainfall in December-February when the monsoon trough (Section 2.4.3) is closest to the SPCZ region and winds feed more moisture into the SPCZ. It is weaker and less well defined in June-August (Vincent, 1994) as the monsoon trough moves north and the drier zone on the equator expands as the ITCZ (Section 2.4.2) moves north, feeding less moisture into the SPCZ.

Cloudiness and rainfall in the SPCZ region vary on a daily basis with the passage and merging of mid-latitude fronts and low-pressure systems. Tropical troughs and surface low-pressure systems also develop in the SPCZ region and influence rainfall, particularly in the dry season (Thompson, 1986). Therefore, the image of a continuous SPCZ with a line of high rainfall and strong convection represents the seasonal or long-term average conditions, rather than the conditions on any single day. There is evidence that the Madden-Julian Oscillation (Section 2.4.4) influences the SPCZ on intra-seasonal time scales (Vincent et al., 1994; Matthews et al., 1996). The interannual variability of the SPCZ is dominated by the impact of El Nino Southern Osclillation (ENSO) events, with the SPCZ moving north and east during El Niño events and south and west during La Niña events (Trenberth, 1976; Vincent, 1994; Folland et al., 2002; Vincent et al., 2011. See Chapter 3 for more information about ENSO.

2.4.2 Intertropical Convergence Zone

The ITCZ is a persistent east-west band of low-level wind convergence, cloudiness, and rainfall, located close to the equator and extending across the tropical North Pacific (Figure 2.6). These features are also co-located with high-level wind divergence and maxima in upward motion at an altitude equivalent to 500 hPa. The ITCZ is one of the major features determining the climate of the tropical North Pacific, marked by the presence of a surface pressure trough, and formed by the convergence of moisture and heat-laden Northern and Southern Hemisphere trade winds. The upward branch of the Hadley Circulation cell in the Pacific sits over the ITCZ. In the eastern Pacific its north-south position is located where sea-surface temperatures are warmest – to the north of the equator. In the west, in the region of the West Pacific Warm Pool, sea-surface temperature gradients are weaker, and the ITCZ is not necessarily located over the very warmest waters. The Partner Countries affected by the ITCZ are Papua New Guinea, Palau, Federated States of Micronesia, Marshall Islands, Nauru and Kiribati. See Table 2.4 for further details.

In the central and eastern Pacific, the ITCZ is narrow, whereas in the western Pacific the ITCZ becomes broad, due to strong monsoon flows, and the breadth of the West Pacific Warm Pool to the north and south. Seasonal variation in the position of the ITCZ in the central and eastern Pacific shows an amplitude of approximately 5° of latitude (Figure 2.7). The ITCZ is closest to the equator in March-May, and furthest north during September-November, when it becomes broader, expanding

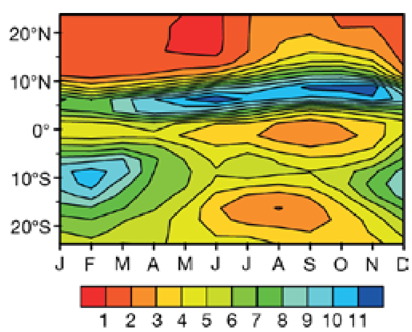


Figure 2.7: Variation of average rainfall with month (horizontal axis) and latitude (vertical axis) averaged over longitude band 155°E–140°W in mm per day. Data from the Climate Prediction Center Merged Analysis of Precipitation dataset for years 1979–1999. The regions of strong rainfall (blue/green) to the south and north of the equator indicate positions of the SPCZ and ITCZ respectively.

both to the north and south (Waliser and Gautier, 1993). Rainfall totals in the ITCZ region (defined here as 160°E–120°W, 0°-15°N) peak in September-November at values around 50% higher than those of the (late) southern summer, January-March. These results are based on CMAP precipitation analysis, with a smaller seasonality indicated by GPCP analysis.

A weak ITCZ-type feature, distinct from the SPCZ, forms in the eastern South Pacific at around latitude 5°S during March-May when northern rainfall is at its weakest with the rainfall pattern in the region symmetric across the equator (De Szoeke and Xie, 2008). As this South Pacific feature of the ITCZ is weak and only appears for a few months each year, the term ITCZ is almost always used to refer to the much more prominent Northern Hemisphere ITCZ. The same convention is used in this publication.

2.4.3 West Pacific Monsoon

The WPM is the eastern edge of the Maritime Continent Monsoon, and is the southern extension of the larger Asian-Australian Monsoon system. This system moves seasonally from the Northern Hemisphere across the equator into the tropical regions of the Southern Hemisphere during December-February. The WPM therefore extends into the transition region of both the SPCZ and the ITCZ.

The WPM can be characterised by the seasonal reversal of the prevailing winds (Figure 2.8; Kim et al., 2008). This reversal of the cross-equatorial low-level wind originates from a reversal in the cross-equatorial atmospheric pressure that results from changes in the land-sea temperature contrast (Wang, 2006). The timing of the onset of the monsoon is often tied to the arrival of the Madden-Julian Oscillation (Section 2.4.4; Hendon and Liebmann, 1990a, b) and intrusion of mid-latitude troughs (Davidson et al., 1983).

The right panel of Figure 2.8 indicates regions which experience a reversal in the east-west wind component between December-February and June-August and therefore receive rainfall from the WPM. In the Southern Hemisphere, these include East Timor, Papua New Guinea, Solomon Islands and, on the outer edges, Tuvalu and Vanuatu. In the Northern Hemisphere, a seasonal wind reversal is experienced in Palau and parts of the Federated States of Micronesia, and in some years as far east as the Marshall Islands.

Along with the reversal in low-level winds comes the typical intra-seasonal variability in monsoon rainfall (Figure 2.9). Monsoonal rainfall has significant daily to weekly variability, with a monsoon season having several burst (wet) and break (dry) periods between initial onset and final end of the monsoon season.

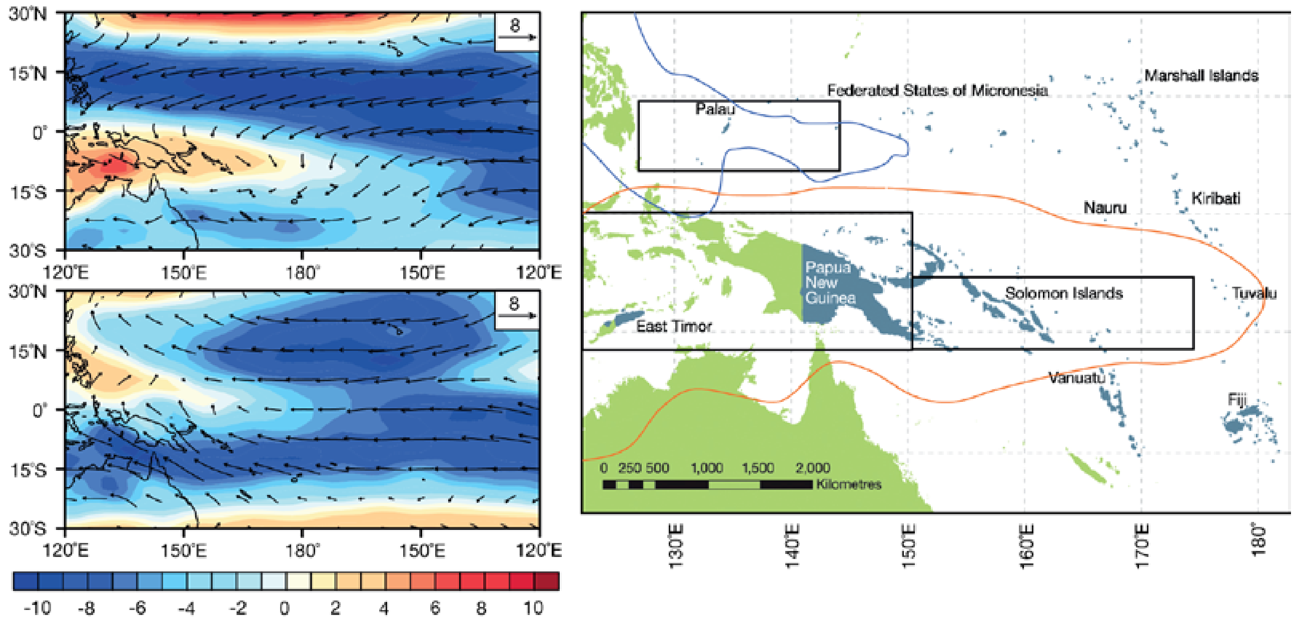


Figure 2.8: (Left) The magnitude of the low level (925 hPa) zonal wind components for the months of January (upper panel) and July (lower panel). The white shading indicates the transition from westerly (yellow to red) to easterly (blue) components. (Right) Regions affected by westerly wind components are found within (to the left of) the orange contour line during December-February and the blue contour line during June-August. Area averages have been calculated for the boxed regions Northern Hemisphere box (125–145°E; 5–10°N), Southern Hemisphere box-1 (120–150°E; 0–10°S), and Southern Hemisphere box-2 (150–175°E, 5°S–10°S). (Source: NCEP reanalysis data (Kalnay et al., 1996)).

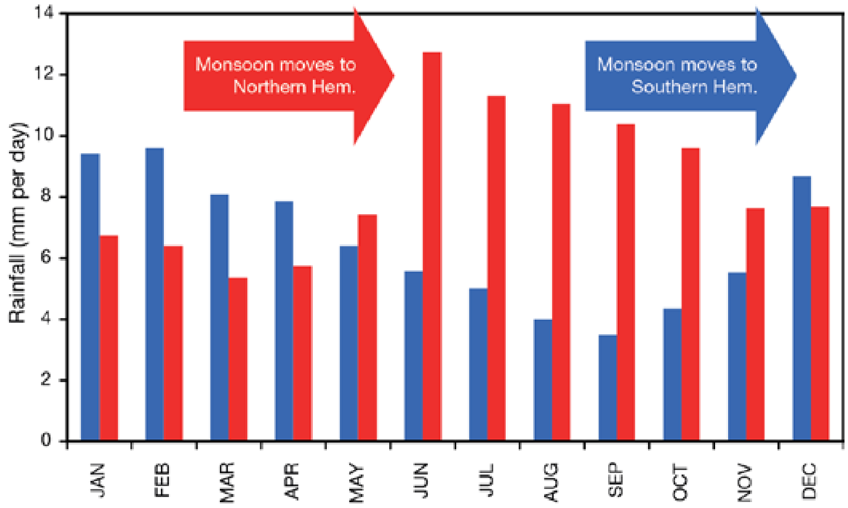


Figure 2.9: Annual cycle of monthly rainfall (in mm per day) for the Southern Hemisphere monsoon westerly region (Southern Hemisphere box-1, blue bars) and the Northern Hemisphere monsoon westerly region (Northern Hemisphere box, red bars) as indicated in Figure 2.8. Arrows show the build up to onset of the Northern (red) and Southern (blue) Hemisphere monsoons.

The onset of the monsoon in the Northern Hemisphere in May-June is much more sudden than its retreat in October-November. The reason for this lies in the different thermal inertia of land compared with ocean, which facilitates the south-eastward move but hinders the north-westward move during the retreat (Wang and Chang, 2008), leaving large monsoon convective regimes, mainly south of latitude 5°N, until the end May. For the Southern Hemisphere this feature is not present, showing a smoother transition for monsoon onset and retreat.

2.4.4 Other Climate Features

Whilst the large-scale climate features described previously are largely responsible for setting the mean state and seasonal variations in the climate of most PCCSP countries, there are other phenomena that affect the climate on shorter time scales. The Madden-Julian Oscillation (MJO) mentioned earlier is a tropical circulation feature moving west to east along the equator with a frequency of 40–60 days (Madden and Julian, 1994). The MJO brings significant variations in convection and is one of the dominant drivers of intra-seasonal rainfall variations in the tropics. It is most active in December-February, often resulting in short wet and dry periods within a single monsoon season (Wheeler and McBride, 2005). Active phases of the MJO usually develop in the western Indian Ocean and move eastward into the western Pacific Ocean. In the PCCSP region they can influence the climate variability within a season particularly in East Timor, Papua New Guinea and the Solomon Islands.

The Southern Annular Mode is another important feature of extra-tropical climate variability which varies on short, synoptic time scales (up to 10 days), as well as longer periods. It influences the sub-tropical high pressure systems that affect the climate of many countries in the South Pacific sub-tropics. It is described in more detail in Section 3.4.7.

2.5 Mean Climate at Key Locations Across the Region

Average seasonal climate variations are given here for one site in every Partner Country (Figure 2.10) to demonstrate climatic conditions across the region. More detailed descriptions for each country are given in the individual country reports in Volume 2.

Islands between latitudes 15°N and 15°S experience a relatively weak seasonal variation in air and sea-surface temperature (between 1–2°C). This variation is slightly stronger for countries in the far western Pacific south of the equator where the continental land masses have an effect (East Timor and Papua New Guinea). Near the equator some locations (Solomon Islands, Palau, Nauru, Kiribati and Tuvalu) experience two peaks in their annual air temperature cycle (around May and November).

Average monthly sea-surface temperature values are between average monthly mean and average monthly maximum air temperature values at most locations throughout the year. This reflects the small size of many islands and their small impact on local temperatures (Papua New Guinea is the main exception), and indicates the influence ocean temperatures have on the climate experienced on islands.

Compared to air and sea temperature there is greater variation in rainfall. The wettest months in countries in the north Pacific region, Palau, the Federated States of Micronesia and the Marshall Islands, occur when the ITCZ is strongest and displaced furthest north between May and August. Nauru, Kiribati and Tuvalu, all near or just south of the equator, experience their wettest months between December and March, when the ITCZ and the monsoon trough are furthest south. These countries are also influenced by the SPCZ, especially Tuvalu as it is further south. Rainfall is therefore higher in Tuvalu, which is also a consequence of the influence of the WPM.

Partner Countries south of 15°S experience a greater annual air and sea-surface temperature range compared to those further north. For example, at Nuku'alofa, Tonga, the summer to winter maximum and minimum air temperature ranges are more than 6°C. At these locations the air and sea temperature maximum is experienced at most sites between December and March. These are the wettest months and the period when the SPCZ is most active and furthest south. Peak tropical cyclone and Madden-Julian Oscillation activity also occurs during this period. In countries farthest from the equator, such as Niue and Tonga, the difference between wet season and dry season rainfall is reduced as more rainfall in the winter is received from passing cold fronts.

On small low-lying islands, the climatology of the key location may in fact represent that of the rest of the island and nearby islands. However, on larger elevated islands topography can have a significant effect, so the climate of the key location may not be applicable several kilometres away. For example, in Fiji rainfall is higher in Suva, which is exposed to the south-east trade winds, than in Nadi (not shown in Figure 2.10) which lies on what is predominantly the lee side of the same island. There can also be notable differences in climate within countries that are spread over a large area, e.g. the northern and southern Cook Islands.

Table 2.4: List of the Partner Countries and the main features that influence their climate. Also included is whether the country is regularly impacted by tropical cyclones and if it has significant topography leading to climatic effects.

Country	Main climate features and influences
Cook Islands	SPCZ, sub-tropical highs, trade winds, tropical cyclones, topography
East Timor	WPM, topography
Federated States of Micronesia	ITCZ, WPM, trade winds
Fiji	SPCZ, trade winds, sub-tropical highs, tropical cyclones, topography
Kiribati	ITCZ, SPCZ, trade winds
Marshall Islands	ITCZ, WPM (in some years), tropical cyclones
Nauru	ITCZ, SPCZ, trade winds
Niue	SPCZ, trade winds, sub-tropical highs, tropical cyclones
Palau	WPM, ITCZ, trade winds
Papua New Guinea	WPM, ITCZ, topography
Samoa	SPCZ, trade winds, sub-tropical highs, tropical cyclones, topography
Solomon Islands	SPCZ, WPM, tropical cyclones
Tonga	SPCZ, trade winds, sub-tropical highs, tropical cyclones, topography
Tuvalu	WPM, SPCZ, trade winds, sub-tropical highs, tropical cyclones
Vanuatu	SPCZ, trade winds, sub-tropical highs, tropical cyclones, topography

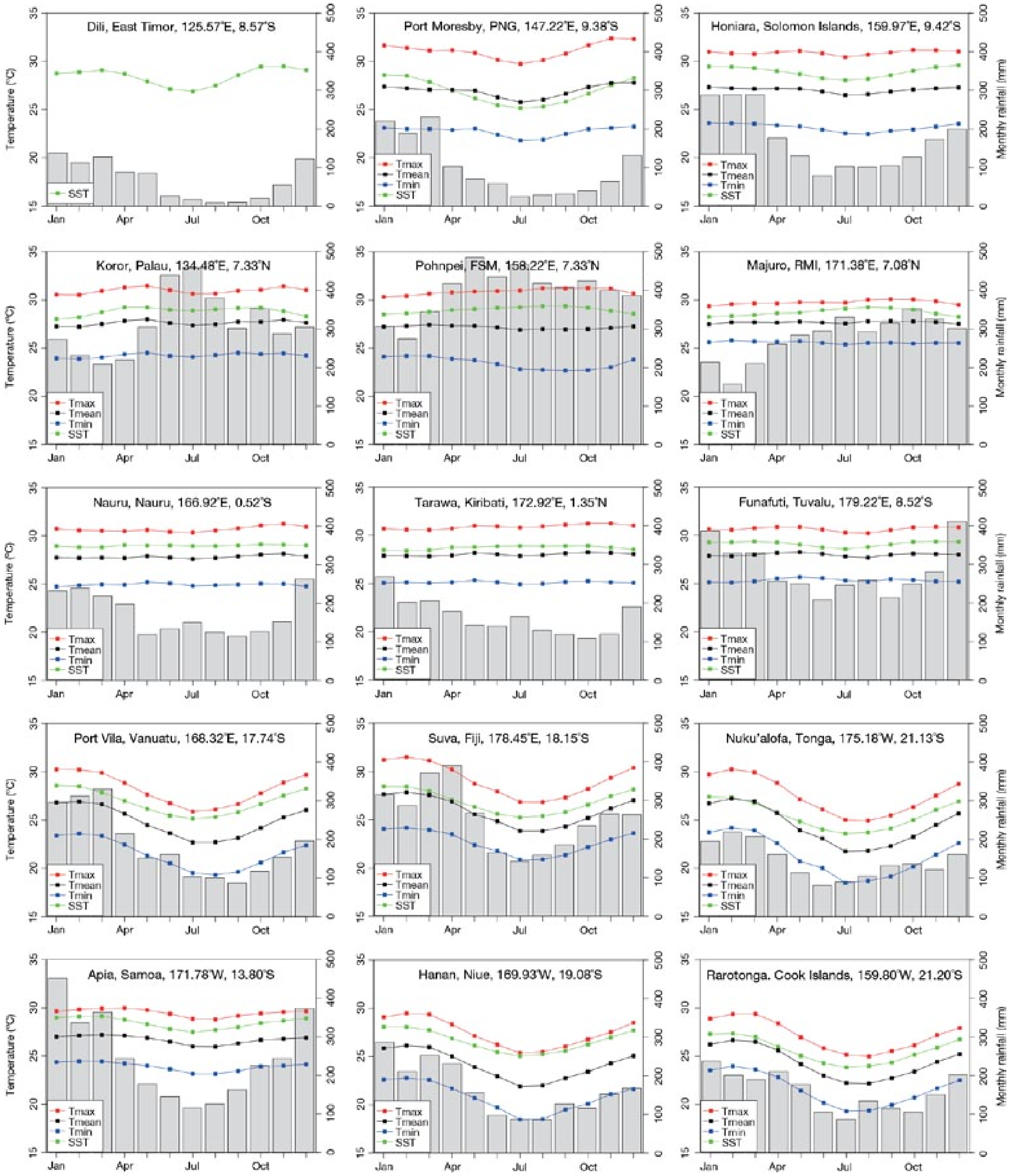


Figure 2.10: Monthly average maximum air temperature (red), mean air temperature (black), minimum air temperature (blue), sea-surface temperature (green) and average monthly rainfall (grey bars) at key locations in each of the Partner Countries. Station data over 1961–1990 were used.

2.6 Current Climate in the Region: Oceans

The ocean and atmosphere interact strongly in the tropical Pacific through the exchange of heat, water and momentum via surface winds. Surface winds affect the upper ocean in the first few tens of metres, driving surface currents and causing water to pile up in certain regions and to drop in others. This in turn generates pressure differences that can create currents extending from the surface to hundreds, or even thousands, of metres below. Surface winds also generate waves at the surface of the ocean which increase in size with the strength, duration and distance over which the wind blows.

While the atmosphere can drive changes in ocean circulation, temperature and salinity, the ocean's surface temperature also shapes the properties and circulation of the atmosphere, as described in previous sections. Movement of heat within the ocean also controls changes in atmospheric circulation and rainfall on interannual time scales. The most dramatic example of this is the wind and rainfall changes driven by shifts in the position of the warmest tropical Pacific waters as part of ENSO (Chapter 3).

The ocean in the PCCSP region has a number of important and unique temperature, salinity and circulation features that influence the overlying atmosphere and play a role in marine productivity.

2.6.1 Sea-Surface Temperature

In general, sea-surface temperature decreases consistently from the equator towards the poles. This is a consequence of the greater amount of solar energy entering the ocean at lower latitudes compared to higher latitudes. There are important regional deviations. Along the equator, easterly trade winds push the warm surface tropical waters towards the west. This forms an extensive pool of the world's warmest waters, the West Pacific Warm Pool, with temperatures exceeding 28–29ºC and covering an area of 15 million km^2 (Wyrtki, 1989; Figure 2.11). The Warm Pool stretches from the central Pacific to the far eastern Indian Ocean. The Indian Ocean portion of the Warm Pool is fed via a warm current (called the Indonesian Throughflow) from the Pacific Ocean (Hautala et al., 2001) via the Indonesian archipelago. The warm water abutting the western margin of the Pacific basin is forced downwards and as a result the very warm waters extends to a depth of up to 100 m (Figure 2.12).

While water is pushed westward in the direction of the winds directly at the equator, the easterly trade winds drive surface currents (in the top 10–50 m) that flow away from the equator in both hemispheres (Figure 2.11, top panel, blue arrows). This deviation of surface current direction from the prevailing wind direction is caused by the Coriolis effect, which is a consequence of the rotation of the Earth. These surface flows are termed Ekman currents. This causes a divergence of surface water away from the equator, draws sub-surface water to the surface in a process known as equatorial upwelling. A similar mechanism occurs adjacent to the South American continent. Here the south-easterly alongshore winds (in combination with the Coriolis effect) drive an offshore Ekman current that again causes coastal upwelling of deep water to the surface.

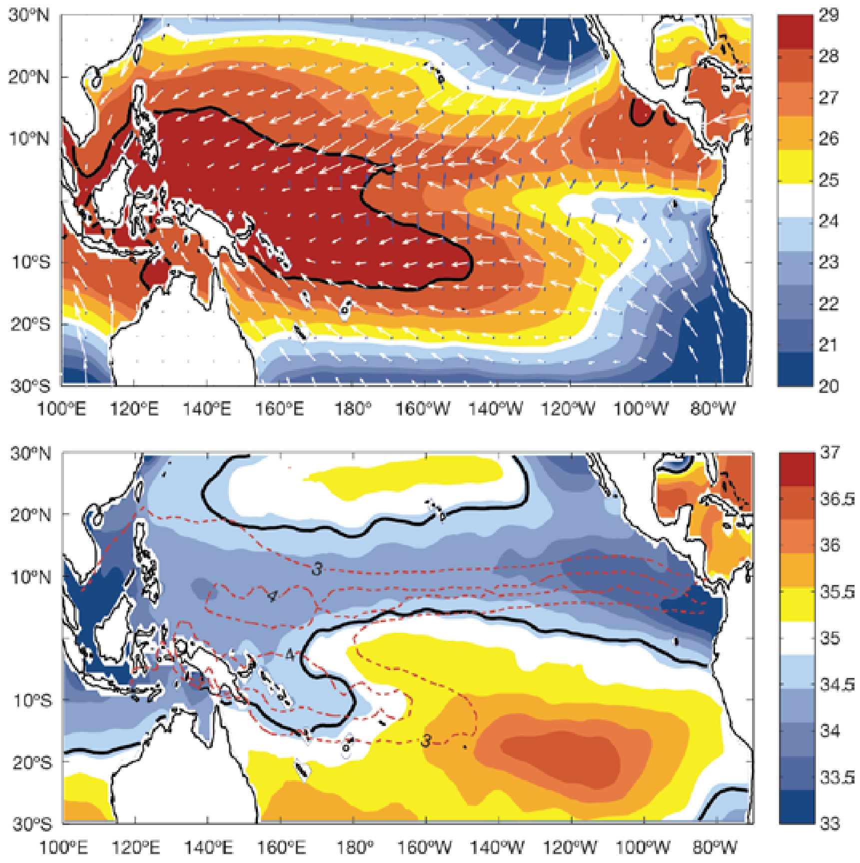


Figure 2.11: Long-term annual-average sea-surface temperature (˚C) (top) and sea-surface salinity (parts per thousand, ppt) (bottom) from the World Ocean Atlas 2005 Database. Thick black lines show the 28.5°C and 34.8 ppt contours, commonly used for the boundary of the Warm Pool. Also shown on the upper panel are the Quikcat wind stress vectors (white arrows) and surface Ekman currents (blue arrows, computed from windstress). Precipitation minus evaporation from ERA-40 reanalysis is shown on the lower panel (mm per day, red dashed contours). See text for explanations of these terms and Tables 2.2 and 2.3 for data sources.

The effect of the coastal and equatorial upwelling is that deeper, cooler and more nutrient-rich waters are pulled to the surface. This leaves a clear signature in the sea-surface temperatures with relatively cool water along the west coast of South America and along the equator in the eastern Pacific, known as the Equatorial Cold Tongue (Figure 2.11). While there is also equatorial upwelling in the western Pacific, because the Warm Pool extends to considerable depth, the upwelled water is still quite warm and does not produce surface cooling as it does in the eastern part of the Pacific basin.

2.6.2 Salinity and Nutrients

The pattern of sea-surface salinity (Figure 2.11) is strongly influenced by rainfall and evaporation patterns. In regions where rainfall exceeds evaporation (as it does in the tropical western Pacific), the salt concentration at the ocean surface is diluted, reducing salinity. The warm waters of the West Pacific Warm Pool and two bands of relatively warm water that straddle the equator centred near latitudes 10ºS and 7ºN (Figure 2.11) are the regions of high convective activity, surface wind convergence, and rainfall associated with the South Pacific and Intertropical Convergence Zones. As a result, they are also regions of relatively low surface salinity. The high rainfall over the Warm Pool relative to surrounding regions leads to the Warm Pool being bounded by a sharp salinity contrast (Maes et al., 2004).

In the tropical Pacific a large amount of heat and fresh water from rainfall enters the surface ocean and thus surface waters tend to be much warmer and fresher than sub-surface water. In the top few tens of metres of the ocean, strong vertical mixing occurs through the action of winds, waves and convective activity. This gives rise to a region where temperature and salinity, and therefore density, are well mixed and are almost constant with depth. This layer of the ocean is called the surface mixed layer.

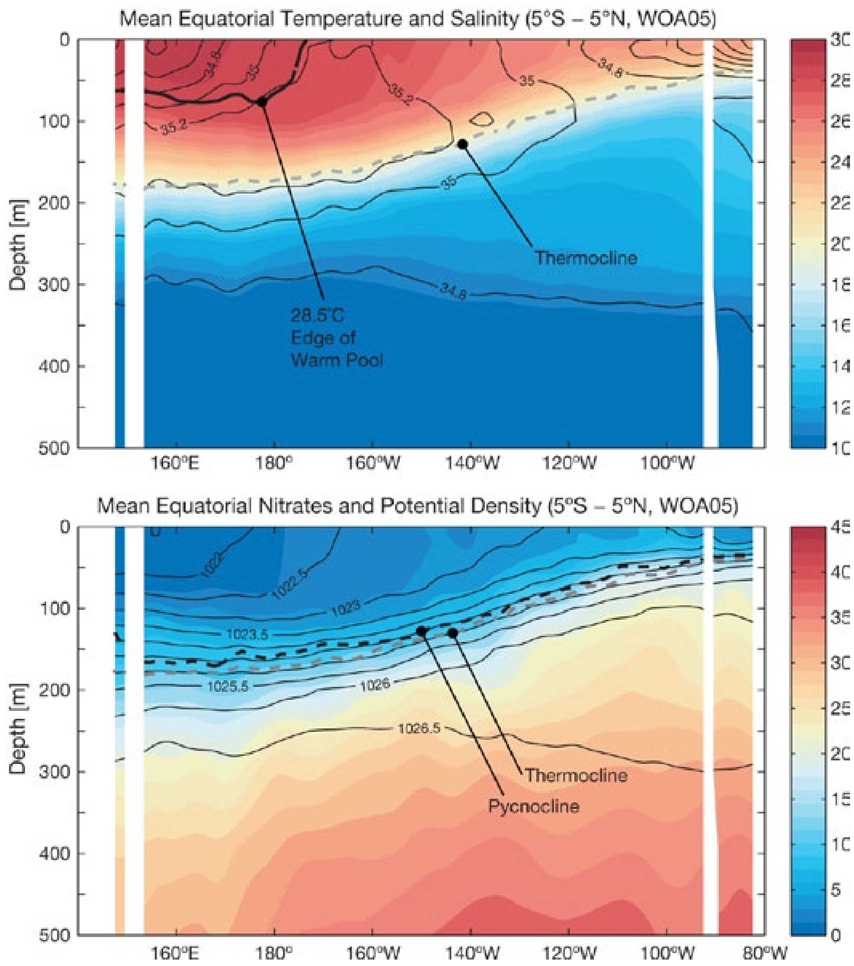


Figure 2.12: Ocean properties in the upper 500 m of the ocean near the equator. (Top) Average temperature (colours; °C) and salinity (line contours; parts per trillion) near the equator (5°S – 5°N). Superimposed are the Warm Pool location, commonly defined as temperatures greater than 28.5°C (black solid line), and the average thermocline depth (the depth of strongest vertical temperature change, grey dashed line). (Bottom) Average nitrate concentration (colours, micromoles per litre) and potential density (line contours, kg per cubic metre). Superimposed are average thermocline (grey dashed line) and pycnocline (the depth of strongest vertical density change, black dashed line) depths. Calculated from World Ocean Atlas 2005.

As noted, both temperature and salinity affect water density: warmer water is less dense than colder water, and fresher water is less dense than saltier water. As temperature decreases and salinity increases rapidly below the mixed layer, there is a sharp density increase below the mixed layer. The region of strongest vertical density change is called the pycnocline. The pycnocline separates an upper layer of relatively light (low density) water from the deep heavy (high density) water. The pycnocline acts as a barrier against mixing between the layers just as 'light' oil sitting on top of 'heavier' water is hard to mix together. The pycnocline gets shallower from west to east (Figure 2.12).

The depth at which the vertical temperature gradient is strongest is called the thermocline. Because density is determined to a large extent by temperature in tropical surface waters, the pycnocline and thermocline are at very similar depths over much of the ocean, and are commonly used to mean the same thing. An important exception occurs in regions of high rainfall (e.g. in the vicinity of the Warm

Pool and the convergence zones) where surface density is strongly reduced by the lower salinity, so the pycnocline sits at a shallower depth than the thermocline. This can have important consequences for interactions between the ocean and atmosphere.

This layering of ocean properties is known as stratification. Stratification strongly affects marine biological activity. Primary productivity (growth of marine plants/phytoplankton) only occurs in the surface ocean where sufficient light is available for photosynthesis. However photosynthesis and other metabolic activity require certain nutrients. The free nutrients are quickly consumed and depleted in the sun-lit (photic) surface ocean as a result of biological activity, but become much more abundant below these depths. This is because biological material falling downwards from the surface into the deep ocean is reprocessed by bacteria, releasing the locked up nutrients such as nitrate (Figure 2.12). Surface nutrients increase from west to east because of coastal and equatorial upwelling that supply the surface ocean with nutrient-rich colder water.

Ocean processes, such as the vertical currents associated with coastal and equatorial upwelling, work against the inherent ocean stratification and facilitate the transfer of deep nutrients up into the photic zone where they are used for biological production. Other processes that affect stratification include ocean mixing via strong winds or storms and convective activity, both of which can act to increase the depth of the mixed layer and entrain nutrient-rich deeper water. In the interior ocean, internal waves, often generated by tidal currents, move through the water and interact with the bottom of the ocean in shallow regions, causing vertical mixing. Ocean eddies (whirlpools of water that are many kilometres in diameter) are also important for mixing and are associated with strong vertical movement of water in their cores.

2.6.3 Ocean Currents

In addition to the Ekman currents that are confined to the first few tens of metres of the surface ocean, and the associated upwelling currents at the equator and along the South American coastline, there are a number of other important ocean currents in the Pacific Ocean (Figure 2.13). At higher latitudes the sub-tropical gyres (vast rotating masses of water extending downwards over a thousand metres) span the north and south extra-tropical regions. Their broad eastern equatorward flowing limbs extend across the central and eastern parts of the basin. In contrast, their western, poleward flowing limbs form narrow (100–200 km wide), fast flowing (sometimes exceeding 1 metre per second) jets (or western boundary currents) that hug the continental margins. The South Pacific western boundary current is the East Australian Current (EAC) (Figure 2.13). Its Northern Hemisphere counterpart is the Kurushio Current. The westward flow of the low-latitude limbs of the sub-tropical gyres extends across much of the tropical ocean in the form of the South and North Equatorial Currents in the Southern and Northern Hemispheres, respectively. The sub-tropical and tropical circulation is indirectly driven by the surface winds, in particular the south-east and north-east trade winds. Wind conditions are dramatically altered within the convergence zones, and as a result the prevailing westward currents of the South and North Equatorial Currents are interrupted in these areas. In fact, beneath the Intertropical Convergence Zone (Section 2.4.2) the prevailing current (the North Equatorial Counter Current) is reversed across most of the basin (Johnson et al., 2002). The corresponding current reversal beneath the less extensive South Pacific Convergence Zone (Section 2.4.1), the South Equatorial Counter Current, is confined to the western basin in the Southern Hemisphere (Gouriou and Toole, 1993).

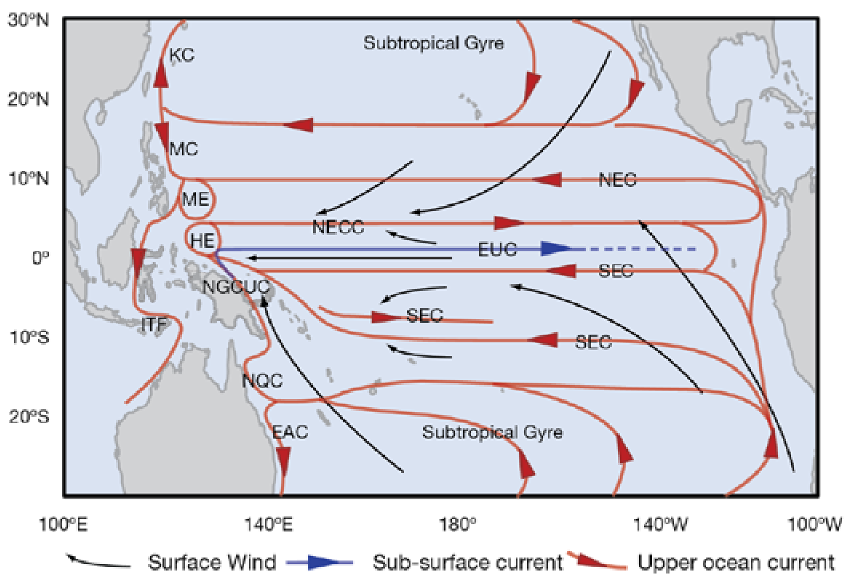


Figure 2.13: Schematic of upper ocean circulation and surface winds. Upper ocean currents range in depth from a less than 100 m to 2000–3000 m in the case of the sub-tropical gyres. Legend: East Australian Current (EAC); Equatorial Undercurrent (EUC); Halmahera Eddy (HE); Indonesian Throughflow (ITF); Kurushio Current (KC); Mindanao Eddy (ME); North Equatorial Current (NEC); North Equatorial Counter Current (NECC); New Guinea Coastal Current (NGCC); New Guinea Coastal Undercurrent (NGUC); North Queensland Current (NQC); South Equatorial Current (SEC); South Equatorial Counter Current (SECC); Modified from Tomczak and Godfrey (2003, Figure 8.6).

Water transported westward in the northern limb of the southern sub-tropical gyre divides at the Australian continental margin. Some of the water flows southwards as part of the East Australian Current, while the remainder flows northward as part of the North Queensland Current, and later the New Guinea Coastal Undercurrent. Water runoff from Papua New Guinea is thought to supply the New Guinea Coastal Undercurrent with high concentrations of dissolved iron (Mackey et al., 2002). At the equator this current feeds the Equatorial Undercurrent, a sub-surface current, which sits directly under the equator at a depth of more than 200 m in the west and about 50 m in the east. This narrow jet of water supplies the iron essential for biological productivity to the eastern Pacific.

In addition to the broad-scale circulations described, the ocean is densely populated with fine-scale currents that are associated with ocean eddies and fronts. Furthermore, as broad currents interact with the ocean floor or islands, they are deflected or channelled, forming local features including jets, island boundary currents and island eddies that have important biological implications. As the South Equatorial Current interacts with Vanuatu and New Caledonia, for example, the North and South Caledonia Jets and the North Vanuatu Jet are formed (Webb, 2000; Gourdeau et al., 2008; Ganachaud et al., 2008). In addition, high islands can also alter the strength and direction of the prevailing wind on the lee side, which can also drive local currents and upwelling.

2.6.4 Sea Level

The observed mean sea level for the period 1993–2002 derived from satellite altimeter, in situ measurements and a geoid model (Maximenko et al., 2009, data available from http://apdrc.soest.hawaii.edu/projects/DOT; Rio and Hernandez 2004, data available from http://www.aviso.oceanobs.com) reveals information about ocean currents as the upper ocean circulation approximately follows contours of constant sea-level height in the ocean interior (Figure 2.14). Along the equator, there is a strong zonal (east-to-west) sea-level slope, with sea level west of the International Date Line (180° longitude) about a half metre higher than that in the cold tongue region of the eastern equatorial Pacific and South American coastal regions. This zonal tilting of sea level on the equator is required to balance the trade winds pushing surface water westward. Higher sea level can also be found in the centres of sub-tropical gyres (about 20° to 40° N and S), with relatively strong (weak) zonal slope in the west (east), which is consistent with strong and narrow poleward western boundary currents and slow and wide equatorward return flow in the interior of the basin. There is a sea-level ridge around 5°N and a sea-level trough around 10°N, which is associated with the eastward-flowing North Equatorial Counter Current between these latitudes.

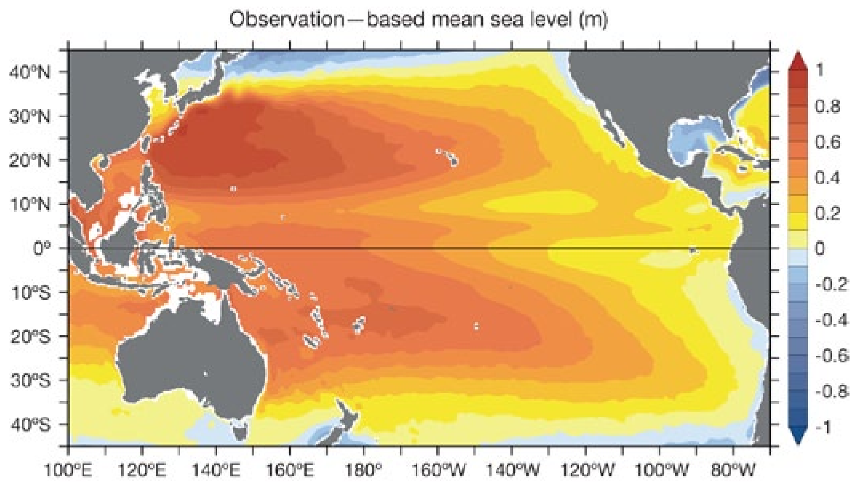


Figure 2.14: Mean sea level (in metres) from observations. The global mean has been removed. (From Maximenko et al., 2009).

2.7 Atmospheric and Oceanic Extremes

The seasons bring regular changes to climate in most Partner Countries. However, occasionally events occur that result in extreme rainfall, winds and temperatures, as well as flooding from storm surges and waves. The following sections describe the main short-term extremes that occur in the region. Year-to-year variability in extreme events, as well as the factors that cause this variability, are described in Chapter 3.

2.7.1 Tropical Cyclones

Each year, approximately 80 tropical cyclones form around the globe (Gray, 1979), with about one-third of them occurring in the Southern Hemisphere (Kuleshov, 2003). In the western North Pacific, about 30 tropical cyclones occur on average per year (Chan, 2005) which is more than for any other ocean basin in the world. Tropical cyclones dramatically affect maritime navigation and communities in many of the PCCSP Partner Countries and other island nations in the Pacific.

The Southern Hemisphere experiences most of its tropical cyclones during November-April, with a maximum in tropical cyclone frequencies during the January-March period when there is an average presence of one to two cyclones per day (Figure 2.15). On average about 2.2 cyclones are present on any given day in the Southern Hemisphere, with about 1 per day in the South Pacific around the end of February to the beginning of March. Very occasionally a cyclone may occur in the Southern Hemisphere on either side of the main cyclone season, in May and October.

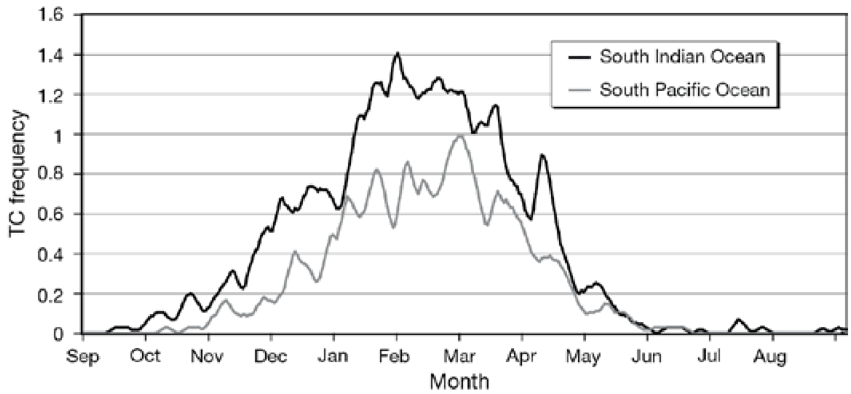


Figure 2.15: Seasonal profiles (1969/70–1998/99) of the average daily number of active tropical cyclones (TC) for the South Indian and the South Pacific Oceans (Source: Kuleshov, 2003).

The spatial distribution of the average number of tropical cyclones per year in the Southern Hemisphere (Figure 2.16) shows that the most tropical cyclones occur in the South Pacific in the area between the Australian coast and the International Date Line, from about 12°S to 22°S.

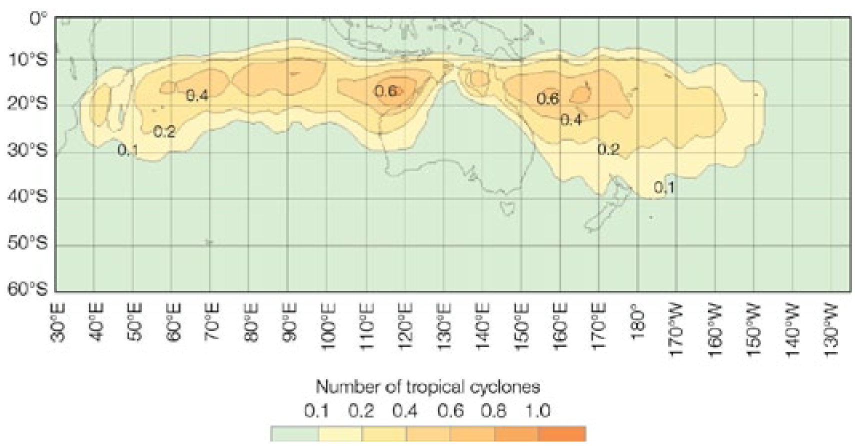


Figure 2.16: Average annual number of tropical cyclones per year in the Southern Hemisphere in an area 2° x 2° over the period, 1969/70 to 2005/06.
Source: http://www.bom.gov.au/jsp/ncc/climate_averages/tropical-cyclones.

In the western North Pacific, the high number of tropical cyclones is due to the high frequency of favourable conditions for their development (Gray, 1979). For example, of all of the world's ocean basins, the western North Pacific has the highest values of an index representing the potential for cyclone genesis (based on relative humidity, vertical wind shear, vorticity, sea-surface temperature and convective available potential energy; Camargo et al., 2007a). Cyclones in this region can form at any time throughout the year, with August having the highest frequency and January-March having the lowest frequency (Chan, 2005). There is also considerable interannual and inter-decadal variation in the number of tropical cyclones that occur in this region (Chan, 1985; Chan and Shi, 1996) and in the South Pacific (Callaghan and Power, 2010; Section 3.5.4). In the western North Pacific cyclones are referred to as typhoons.

2.7.2 Extreme Sea Levels

Extreme sea levels can cause significant coastal impacts, including inundation of low-lying coastal terrain, erosion of beaches, ecosystem loss, damage or destruction of coastal infrastructure, damage to crops and water supplies, and injury or loss of life. They are caused by a combination of three components: tides; seasonal or longer-term fluctuations due to changing wind, pressure and ocean temperature patterns such as ENSO; and short-term events due to weather, such as storm surge, ocean waves and tsunamis. Although tsunamis can cause the most devastating extreme sea-level events, they are not affected by climate and so will not be discussed further here. The contributions of tides, storm surge and ocean waves to sea-level extremes are illustrated in Figure 2.17. The forces that influence these different components of extreme sea level vary and will be discussed in turn in this section.

Astronomical tides are predictable events that are caused by the gravitational forces that arise from the movement of the Moon, Sun and other planetary bodies relative to the Earth. These motions cause the daily rise and fall of waters and the monthly variation in tidal heights (spring and neap tides; Figure 2.18a). In addition to variations on monthly time scales, there is also yearly variation in tidal heights, such that the highest spring tides tend to occur around the same time each year. Although the motions of the sun and moon cause different tides in different locations and no two years, even at the same location, are exactly the same, in most case the highest tides each year tend to occur either when the sun and the moon are closest to the equator (at the equinoxes, in March and October; Figure 2.18b) or when the declination of the sun is at its greatest (at the solstices, in December and June).

As discussed in detail in Chapter 3 (Section 3.6.3.2), ENSO has a major influence on sea levels across the Pacific and this can influence the occurrence of extreme sea levels. During La Niña events, strengthened trade winds cause higher than normal sea levels in the western tropical Pacific, and lower than normal levels in the east. Conversely, during El Niño events, weakened trade winds are unable to maintain the normal gradient of sea level across the tropical Pacific, leading to a drop in sea level in the west and a rise in the east. Pacific islands within about 10° of the equator are most strongly affected by ENSO-related sea-level variations.

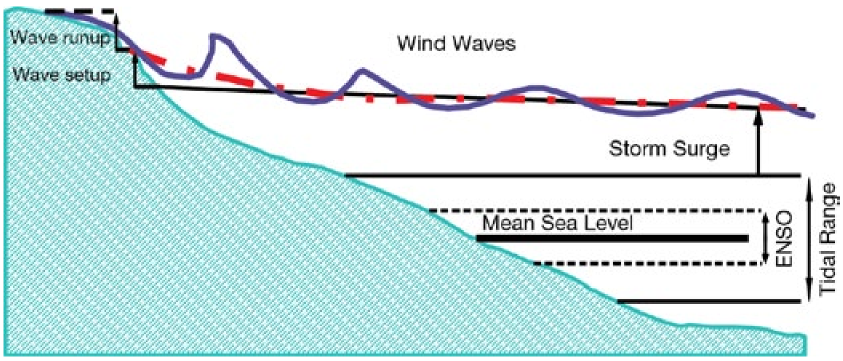


Figure 2.17: Diagram illustrating the contributions to sea-level extremes due to tides, ENSO, storm surge and wind-generated waves. The maximum sea level from the first three of these contributions is given by the thin black line at the top. Also shown are wave height (blue line) and wave set-up (red line).

Tropical cyclones can cause severe short-term sea-level extremes in the Pacific due to storm surges and/or ocean waves. Falling atmospheric pressures associated with cyclones draw the ocean surface upwards at a rate of 1 cm for each hPa drop in pressure, and onshore winds can build up water levels against the coast (wind setup). While cyclone intensity (measured by the minimum central pressure or maximum wind speed) strongly influences the severity of waves and storm surge, other tropical cyclone attributes such as size, direction and speed of movement also play an important role, as does the geomorphology of the coast itself.

Tropical cyclone-induced storm surges tend to be localised, and concentrated in the region of maximum onshore winds close to the cyclone centre. Whilst their impacts are potentially devastating, they are rare at any given location. On the other hand, ocean waves produced by such systems can propagate long distances in the deep ocean as swell, with little loss of energy, and can therefore impact a larger number of more distant coastlines. As the waves encounter shallow coastal waters, they steepen and break, progressively losing energy and producing an increase in coastal sea levels known as wave setup. Atolls with steep shelf margins may be particularly affected by remotely generated swell causing wave setup in lagoons, thereby contributing to sea-level extremes (Callaghan et al., 2006). As a wave breaks at the coast the maximum vertical extent of the wave uprush on a beach or structure above the still water level is known as wave runup (Figure 2.17).

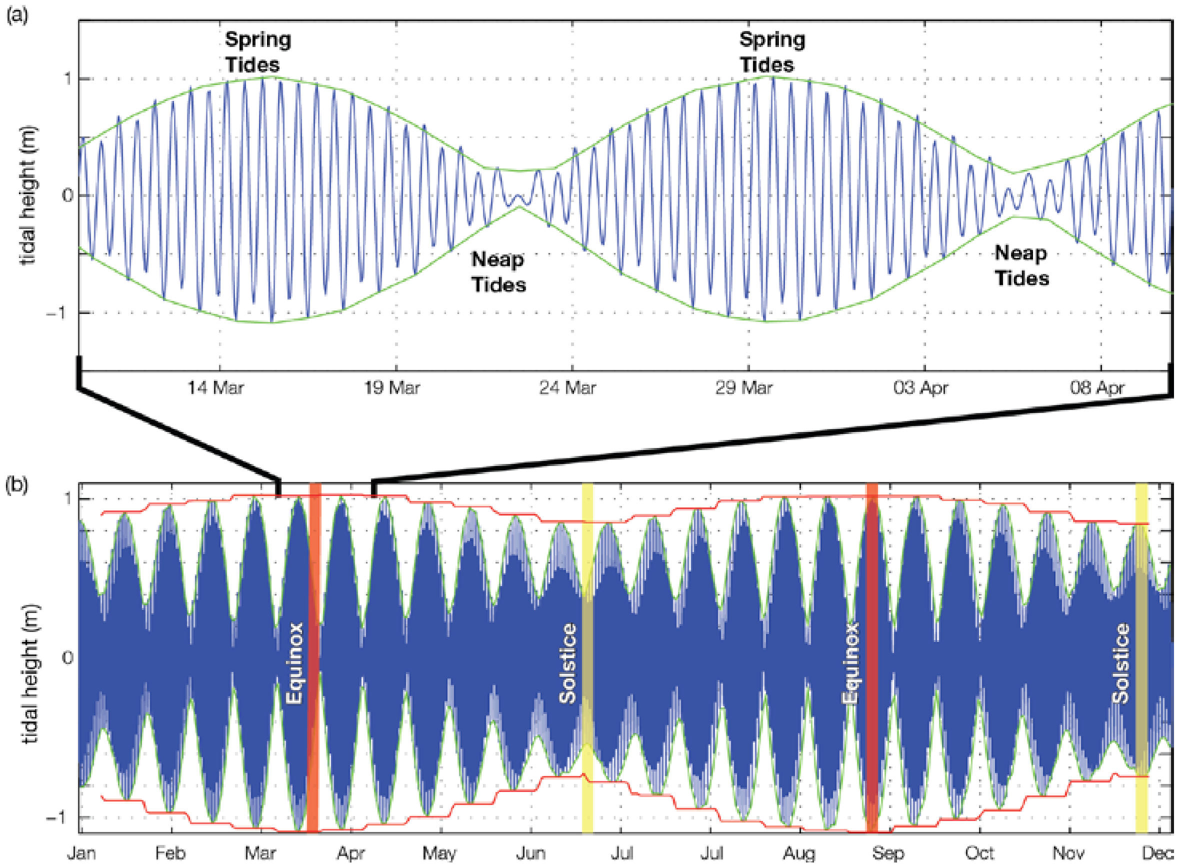


Figure 2.18: A tidal prediction for Tarawa, Kiribati using the five largest astronomic constituents to illustrate (a) spring and neap tides and (b) the semi-annual range in tides; (a) is over a one-month period centered on the March 2006 equinox, indicating times of spring and neap tides and (b) shows the entire year of 2006, indicating the times of the equinoxes and the solstices. In (a) and (b) the green lines indicate the maximum daily tidal range envelope and in (b) the red lines indicate the maximum fortnightly (two-week) tidal range.

The coastal impacts resulting from the cyclone-induced storm surges and waves are also a function of various coastal attributes, such as the shape of the sea bed (bathymetry), the shape of the coastline in relation to the cyclone path, and the landform (geomorphology). Wide and shallow continental shelves amplify the storm surge while bays and channels can funnel and increase the storm surge height. The presence of reefs will cause waves to break and lose energy before they reach coastal areas.

In summary, while weather events such as tropical cyclones may cause short-term sea-level extremes due to storm surges and/or high waves, it should be noted that the impacts of these events can be strongly moderated by other factors such as tides, seasonal variations in sea level and ENSO (Figure 2.17). The tidal range can experience large variations throughout the year so that the potential for a short-term extreme event coinciding with higher than normal background sea level is greater in some months in the year compared to others. Seasonal variations in sea level arise in some locations from variations in atmospheric circulation patterns and/or ocean currents. On interannual time scales, ENSO variability can affect both the weather events that cause short-term sea-level extremes and the background mean sea level (Chapter 3). The relative roles of these factors are investigated for PCCSP countries in Volume 2.