ANALYSIS OF THE KYOTO PROTOCOL ON PACIFIC ISLAND COUNTRIES

Part I
Identification of Latent Sea-Level Rise within the Climate System at 1995 and 2020

Part II
Regional Climate Change Scenarios and Risk Assessment Methods

South Pacific Regional Environment Programme (SPREP)

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PART I
IDENTIFICATION OF LATENT SEA-LEVEL RISE WITHIN THE CLIMATE SYSTEM AT 1995 AND 2020

PART II
REGIONAL CLIMATE CHANGE SCENARIOS AND RISK ASSESSMENT METHODS

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PICCAP
our islands, our lives...

Contents

Part One: Identification of Latent Sea-Level Rise within the Climate System at 1995 and 2020

Part Two: Regional Climate Change Scenarios and Risk Assessment Methods

Executive summary

This report addresses the following questions as specified by the South Pacific Regional Environment Programme (SPREP):

What is the latent sea-level rise due to:

- *anthropogenic emissions of greenhouse gases from pre-industrial times to the present;*
- *projected anthropogenic emissions of greenhouse gases from pre-industrial times to 2020 given that the Kyoto Protocol emission targets are met?*

Latent sea-level rise is defined here as the sea-level rise ultimately likely to occur due to emissions of greenhouse gases already in the atmosphere, i.e. if all anthropogenic emissions of greenhouse gases were to cease at a particular time, various global systems would continue to change in response to the gases remaining in the atmosphere until equilibrium was reached. The systems that most impact on latent sea-level rise are:

- *the ocean*, that expands as surface waters warm and penetrate into the deeper ocean;
- *glaciers and small ice-sheets*, that contract when melting exceeds their snow supply;
- *the large ice-sheets of the Antarctic and Greenland* which are complex systems in their own right, and can react to global warming and precipitation changes over short to long time periods.

Two scenarios, based on 5-yearly intervals from 1990–2100, were constructed to investigate these questions:

- ***Scenario 1*** applied estimated emissions of the major greenhouse gases in 1990, estimated CO_2 from fossil fuel emissions and IS92a estimates for the other greenhouse gases in 1995, then natural emissions (after Pepper et al., 1993) from 2000–2100.
- ***Scenario 2*** applied greenhouse gas emissions based on the IS92a scenario, with reductions to anthropogenic emissions based on the Kyoto protocol, to 2020, then natural emissions from 2025–2100.

These scenarios were constructed in order to determine latent sea-level rise at a specified date, without reference to continuing anthropogenic greenhouse emissions beyond that date. They are both physically impossible, as they rely on the assumption that anthropogenic emissions drop to zero from 2000 and 2025 respectively.

The projections of latent sea-level rise for both scenarios are:

	Emissions to	*Range of sea-level rise*	*Time of peak*
Scenario 1	1995	5– 12 cm	2020– 2025
Scenario 2	2020	14– 32 cm	2050– 2100+

These projections also show that as the volume of greenhouse gases emitted into the atmosphere increases, the inertia of systems contributing to sea-level rise also increases, due to greater disequilibrium between atmospheric forcing and the state of the oceans and cryosphere. Both the height and duration of latent sea-level rise increase over time due to projected increases in anthropogenic greenhouse gas emissions.

Latent sea-level rise committed to by past emissions has the potential to threaten all regions where coastal impacts are currently marginal to severe. By 2020 the amount of latent sea-level rise may be sufficient to increase the vulnerability of regions where impacts are currently infrequent or not particularly severe, and to increase the severity of impacts in areas currently under threat.

1. Introduction

Projected sea-level rises accompanying climate change due to the enhanced greenhouse effect are expected to adversely affect almost all small island states, including those in the South Pacific (Nurse et al., 1998). The Intergovernmental Panel on Climate Change (IPCC) Second Assessment Report anticipated sea-level rise over the next one hundred years of about 5 mm/yr with an uncertainty range of 2–9 mm/yr (Warrick et al., 1996). In contrast, the estimated historical sea-level rise since 1880 has been about 1.8 mm/yr with an uncertainty range of 1.0–2.5 mm/yr (Warrick et al., 1996).

In December 1997, the Third Session of the Conference of Parties to the United Nations Framework Convention on Climate Change (UNFCCC) was held in Japan. At this meeting, the Kyoto Protocol was adopted. A major feature of the Protocol is the agreement of Annex 1 parties to undertake mitigation measures which aim to reduce emissions of the major greenhouse gases of the industrialised countries by an estimated 5.2% below 1990 levels in the period 2008–2012 (>5%, Article 3.1, Kyoto Protocol).

Within the general framework outlined by the UNFCCC and associated agreements, countries identified as vulnerable to climate change require the best possible scientific advice on projected changes. As part of this process, such countries also require estimates of the impacts they are already committed to by previous and existing activities.

The aim of this brief study is to identify latent sea-level rise within the global climate system due to:

- anthropogenic emissions of greenhouse gases from pre-industrial times to the present;
- projected anthropogenic emissions of greenhouse gases from pre-industrial times to 2020 given that the Kyoto Protocol emission targets are met.

Latent sea-level rise is defined here as the sea-level rise ultimately likely to occur due to emissions of greenhouse gases already in the atmosphere, i.e. if all anthropogenic emissions of greenhouse gases were to cease at a particular time, various global systems would continue to change in response to the gases remaining in the atmosphere until equilibrium was reached. Those systems include the atmosphere; the cryosphere, comprising snowfields, tundra soils, glaciers and ice-caps; the biosphere, including both terrestrial and aquatic ecosystems; and the hydrosphere, incorporating the oceans and terrestrial waters.

2. Method

A Global Climate Model (GCM) is the most appropriate tool for modelling temperature, precipitation and large-scale ice features, all important factors in future sea-level rise. However, due to the computational demands of GCMs, it is not possible to run them for all possible scenarios and physical assumptions in order to gain a comprehensive understanding of future climate change. For this reason, simpler models like MAGICC, the Model for the Assessment of Greenhouse-Gas Induced Climate Change (Wigley and Raper, 1992, 1993, 1995), which have been calibrated using the behaviour of the larger models as a guide, are used to estimate possible changes in global mean sea level.

MAGICC is an upwelling diffusion-energy balance model that takes estimates of greenhouse gas emissions at 5-yearly intervals and calculates projections of atmospheric greenhouse gas concentrations, global mean surface air temperature and sea-level. It has been used in successive IPCC reports, and the version used here is similar to that used in Warrick et al. (1996).

MAGICC utilises historical estimates of greenhouse gas emissions from 1765–1990 to estimate historical sea level, air temperature and processes which carry a memory of past changes, such as the behaviour of ice sheets and oceanic temperature changes. Emission scenarios for 5-yearly intervals from 1990–2100 are then applied to estimate future changes. Two emission scenarios covering 1990–2100 were constructed for the present study:

2.1 Scenario 1

Estimated emissions of the major greenhouse gases were applied for 1990. For 1995, the estimated emissions of CO_2 from fossil fuel and IS92a emissions for CO_2 from deforestation and the other major greenhouse gases were applied. From 2000, all greenhouse gases were reduced to natural emissions, as outlined in Pepper et al. (1993). For example, methane and nitrous oxide emissions were sharply reduced, with halocarbons, and CO_2 emissions from fossil-fuel and deforestation becoming zero (Figure 1).

2.2 Scenario 2

The second emission scenario applied the same data as above to 1995, IS92a data in 2000 and IS92a with Annex 1 total country emissions based on 1990 values reduced by 2% in 2005, 4% in 2010 and 5.2% in 2015 and 2020, consistent with the Kyoto Protocol. This was assumed to be a 'slow compliance' scenario, where full compliance was not attained until 2012, and could not be applied in the model until 2015. From 2025, all greenhouse gases were reduced to natural emissions, as outlined above (Figure 1).

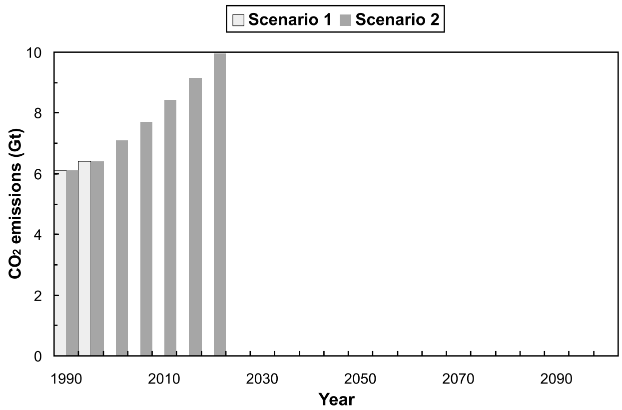

Figure 1: Annual anthropogenic emissions of CO_2 in Gigatonnes as applied in Scenarios 1 and 2.

It must be stressed that these scenarios are totally unrealistic. They are not intended to represent any feasible course of action which could produce an outcome of zero anthropogenic emissions. They have merely been created to demonstrate the inertia of the global climate system to previous atmospheric emissions and consequent changes in radiative forcing, by estimating latent sea-level rise based on existing and past activity, and projections to 2020 arising out of the recently adopted Kyoto Protocol.

Projections of sea-level rise from MAGICC were compared with estimated ranges of historical sea-level rise sea-level based on observations and on factors such as thermal expansion, glaciers and ice-sheets. These comparisons were used to narrow the ranges of sea-level rise produced by MAGICC by removing the low case projections which are not supported by the historical data.

3. Projected sea-level rise

The major factors within MAGICC affecting sea-level rise are thermal expansion of the ocean, contributions from glaciers and small ice caps, and contributions from the large ice sheets of Antarctica and Greenland (Wigley and Raper, 1992, 1993, 1995). The individual influences of these factors change over time. For instance, surface warming of the oceans and changes to small glaciers occur relatively quickly, whereas deep warming of the ocean and large changes in the mass balance of the Antarctic ice-sheet occur more slowly and may have a continuing impact on sea level, lasting for hundreds to thousands of years.

Although temperature is the most significant driver of sea-level rise via thermal expansion and glacial melting, changing precipitation can affect the mass balance of both small and large ice masses. For example, warmer temperatures are expected to lead to an increase in precipitation over Antarctica, leading to an increase in snow accumulation which remains frozen under prevailing temperatures, providing a negative influence on sea level (Smith et al., 1998). On the other hand, if warming occurs in areas close to 0°C, ice may melt, contributing to sea-level rise. This possibility has been raised for the Greenland ice cap (Thompson and Pollard, 1997). Increased breakup or outflow of ice from grounded ice shelves and glaciers may also contribute to sea-level rise (e.g. Binschadler et al., 1998).

The two scenarios as described above were used as input to MAGICC. Three projections were calculated, based on low, mid and high temperature sensitivity as described in IPCC (1996) and on mid-value ice-melt parameters. The results are shown in Figures 2 and 3.

3.1 Scenario 1

Figure 2 shows projections of global mean temperature and sea level based on low, mid and high temperature sensitivity as described in IPCC (1996). Temperature continues to rise to 0.2–0.4°C above its 1990 level at about 2010, then decreases to present values by about 2030–2050. Latent sea-level rise continues for 15–25 years after 1990 reaching peak levels of 1–12 cm.

3.2 Scenario 2

When anthropogenic emissions continue to 2020 as in IS92a modified by the Kyoto Protocol, latent warming is much higher, ranging from 0.5–1.1°C by 2030, and remaining above the present value until >2100 (Figure 3). Sea-level rise is also much more pronounced, ranging from peak levels of about 4 cm to just over 30 cm between 2030–2060. In this case, latent sea level at a low climate sensitivity peaks after about 15 years but at high climate sensitivity, it continues to rise until after 2100.

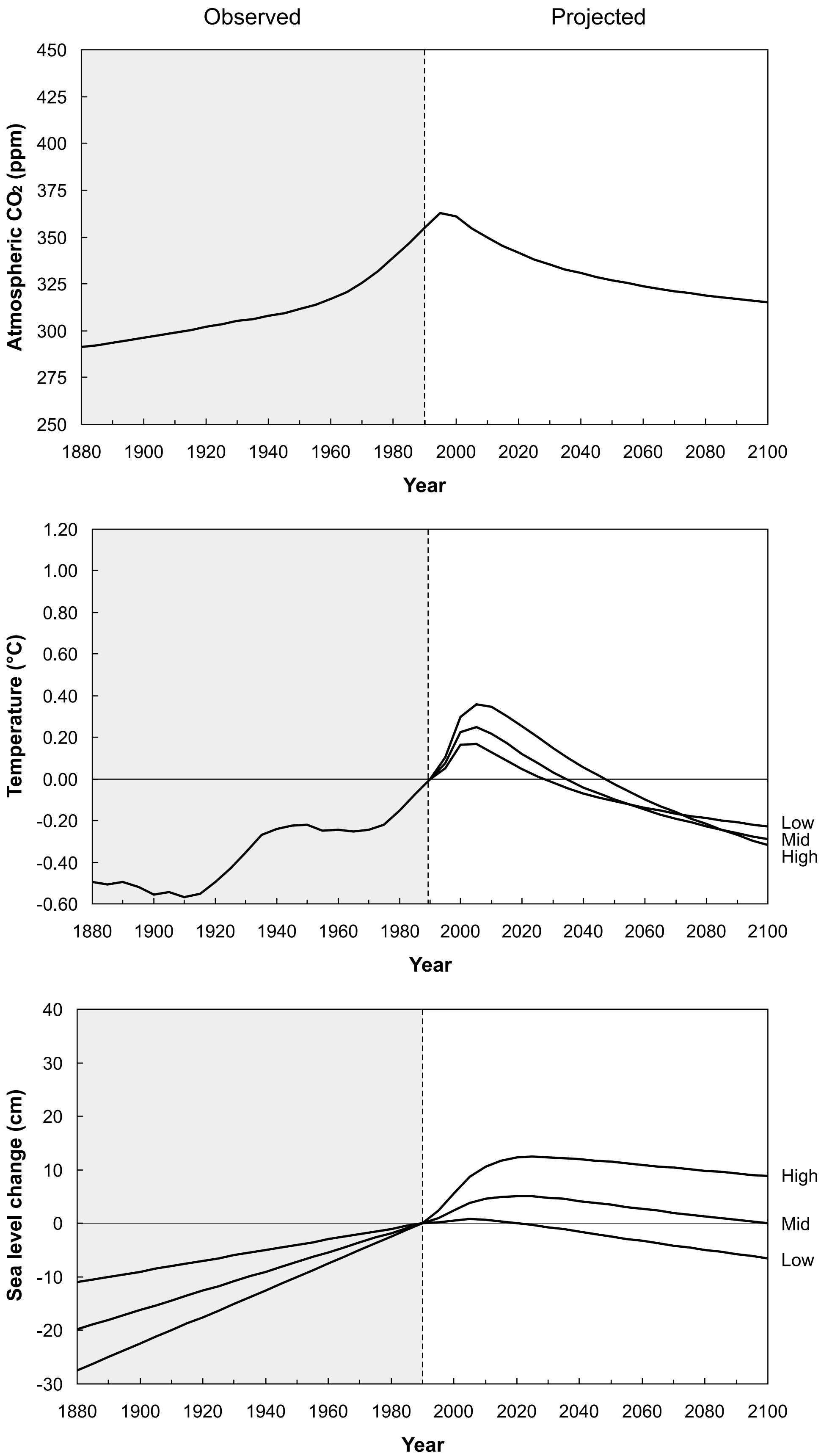

Figure 2. Observed global mean atmospheric CO_2 concentrations, temperature and sea-level changes to 1990 compared with projections to 2100 based on the IS92a emissions scenario to 1995, followed by a return of all greenhouse gas emissions to natural values from 2000 (Scenario 1). Projections of temperature and sea level are based on low, mid and high temperature sensitivity (1.5°C, 2.5°C & 4.5°C) and mid-value ice-melt parameters as in IPCC (1996). Observed data are as in IPCC (1996).

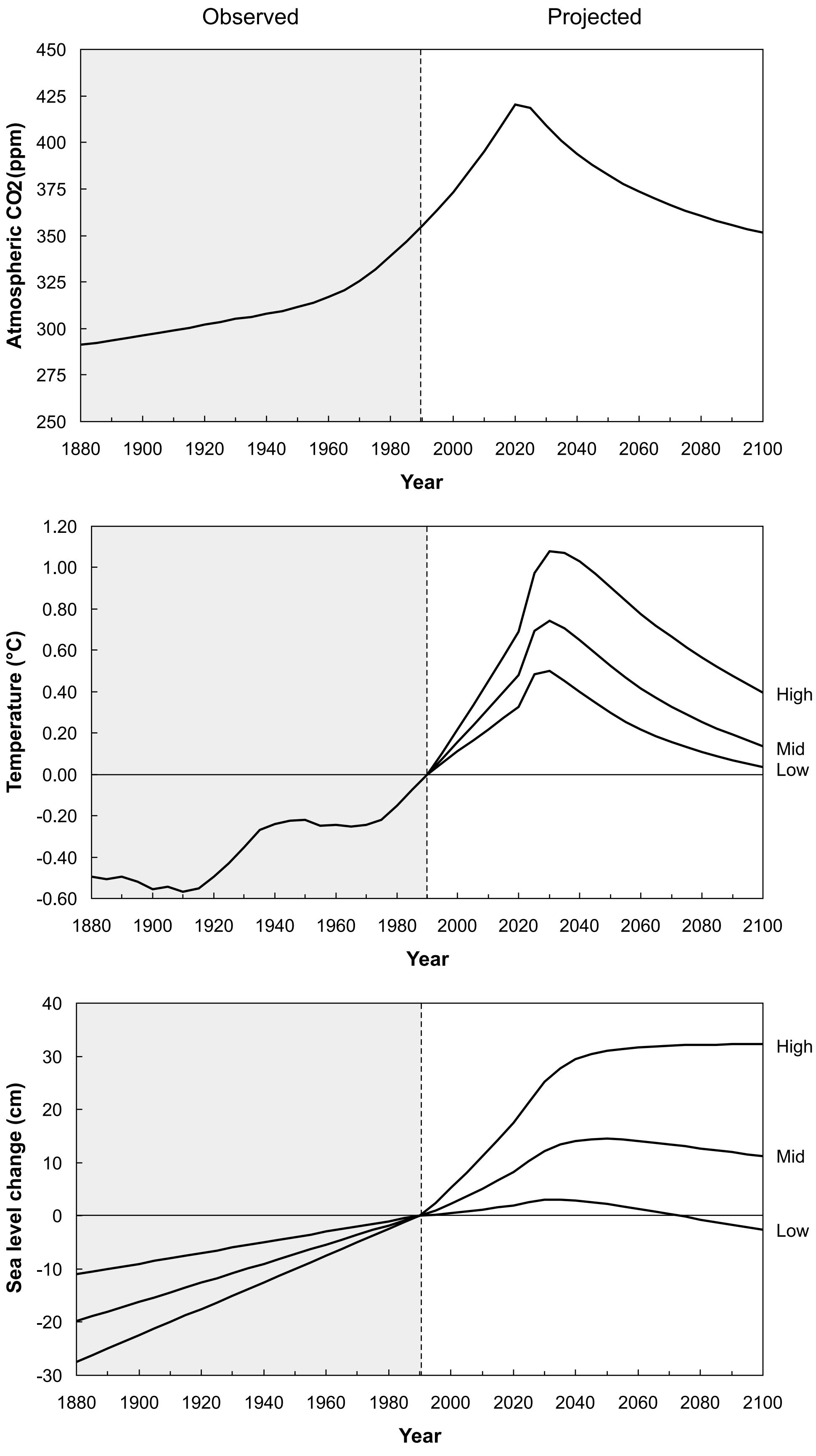

Figure 3. Observed global mean atmospheric CO_2 concentrations, temperature and sea-level changes to 1990 compared with projections to 2100 based on the IS92a emissions scenario to 2000, with a gradual application of the Kyoto Protocol to 2020 followed by a return of all greenhouse gas emissions to natural values from 2025 (Scenario 2). Projections of temperature and sea level are based on low, mid and high temperature sensitivity (1.5°C, 2.5°C & 4.5°C) and mid-value ice-melt parameters as in IPCC (1996). Observed data are as in IPCC (1996).

4. Discussion

It is reasonable to assume that rises in the near future will be a continuation of the historical rise, with the various components of sea-level rise changing relatively slowly due to the inertia of atmospheric changes and their related feedbacks. The West Antarctic ice sheet may be an exception as large-scale, rapid melting is possible, although the risk of this occurring in the next century is low (Oerlemans and van der Veen, 1998; Bindschadler, 1998). In this section, estimated historical sea-level rise from the tide gauge network, and from the various contributing factors, are used to narrow the range of estimated latent sea-level rise from Figures 2 and 3.

Historical global average sea-level rise surveyed in the IPCC Second Assessment Report by Warrick et al. (1996), has risen at about 1.8 mm/yr since 1880 with an uncertainty range of 1.0–2.5 mm/yr. This estimate was based on tide gauge records modified by estimates of vertical land movements (e.g. post-glacial isostatic rebound and tectonism). More recent satellite measurements estimate a rise of 1.8 mm since 1992 (Nerem et al., 1997) although this is a very limited period of time. The historical rise in sea level is due to a number of possible influences and is coincident with a rise in global temperature of 0.45 ±0.15°C over a similar period (Nicholls et al., 1996).

The estimated contributions from various components of sea-level rise are shown in Table 1, together with their individual ranges of uncertainty. Although the total range encompasses the range of estimates from the tide gauge network, there is a large negative component that indicates a fall in sea level, contrary to the evidence. The relative contributions of these components remains unknown, but a large negative contribution from both the Antarctic and Greenland ice sheets (leading to a fall in sea level) is highly unlikely, as it would have more than offset the other factors contributing to the observed sea-level rise (Warwick et al., 1996).

The probability of these large negative components contributing to sea level was tested. Jones (1998) shows that independent variables, when sampled individually and combined in ranges of uncertainty, lead to non-uniform distributions of probability. These distributions show that values close to the centre of such ranges are more likely than those at the extremes, rendering the values of -19 and +37 cm in Table 1 improbable, as long as the extremes of the component contributions are correctly represented.

This exercise was performed on the variables shown in Table 1, creating a composite range with a non-uniform distribution. Independence was assumed, where variables were randomly sampled within each range and then added together, the exercise being repeated 5,000 times. The resultant distribution, encompassing the full range of uncertainty in Table 1, is shown in Figure 4 where it is compared with the range of estimated sea-level rise from the tide gauge network.

Component contributions	*Low*	*Middle*	*High*
Thermal expansion	2	4	7
Glaciers and small ice caps	2	3.5	5
Greenland ice sheet	-4	0	4
Antarctic ice sheet	-14	0	14
Surface and groundwater storage	-5	0.5	7
Total	-19	8	37
Estimated from observations	10	18	25

Table 1: Estimated contributions to the past 100 years of historical sea-level rise in centimetres (Warwick et al., 1996).

The assumption of independence between these variables is largely based on a lack of scientific knowledge allowing them to be quantitatively linked in such an exercise. In reality, some combinations are much more likely than others. For instance, thermal expansion and glacial melting are probably co-dependent; water storage is largely related to land-use, so is independent of warming; and the combined contribution of the Antarctic and Greenland ice sheets to sea level is unlikely to have been negative as indicated by the tabulated 'low' estimates.

When the net negative contribution from the Antarctic and Greenland ice sheets is removed, the two are sampled together in the range 0–18 cm and the exercise is repeated as before, the resulting distribution resembles the historical record much more closely (Figure 5). Figures 4 and 5 support the conclusion of Warwick et al. (1996) that a large negative contribution to sea level from the combined responses of the Antarctic and Greenland ice sheets over the past century is unlikely.

Figure 6 shows the historical sea-level curves as simulated by MAGICC compared with the estimated historical record from the tide gauge network. The estimates for MAGICC indicate that for low climate sensitivity, a fall in historical sea level is projected, and at mid sensitivity, the rise is less than the minimum estimate for observed sea-level rise.

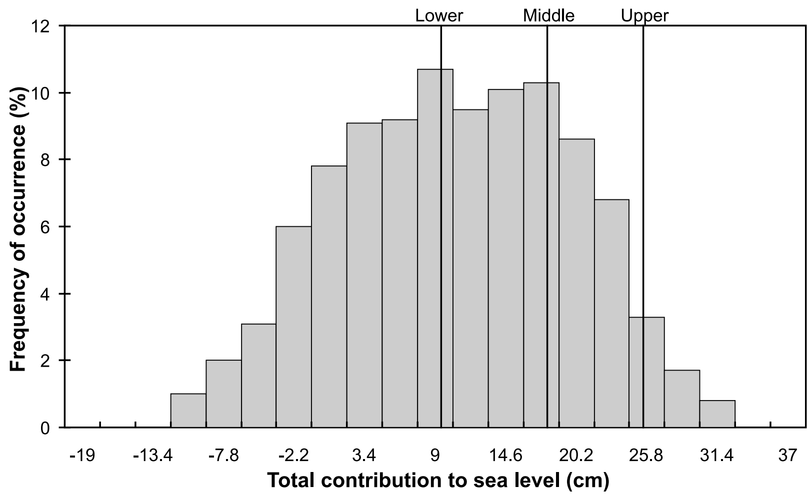

Figure 4: Probability distribution assuming all the uncertainties shown in Table 1 are independent. Note the position of estimated sea-level rise from the tide gauge network in the upper half of the distribution. This is based on 5,000 random samples taken within each of the ranges and added together.

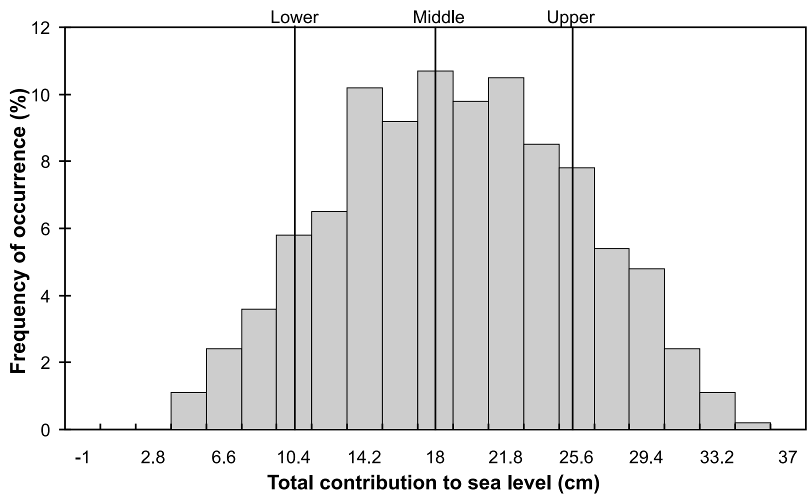

Figure 5: Probability distribution conducted as for Figure 4 but with Greenland and Antarctica sampled together within a range of 0–18 cm, therefore assuming no negative contribution.

The various components of historical sea-level simulated by MAGICC are shown in Table 2. At low climate sensitivity, the sea-level fall shown in Fig 6 is due to a joint negative contribution from Antarctica and Greenland. The mid case has a neutral contribution from the large ice-sheets and is consistent with the range of outcomes shown in Figure 5 despite being lower than the minimum historical estimate from the tide gauge network. The high case is consistent with both Figure 5 and historical estimates.

Given the inertia within ocean and ice systems, within a regime of rising greenhouse gas emissions latent sea-level rise would be expected to continue at, or above, the observed rate. In the low case projections from MAGICC shown in Figures 2 and 3, contributions from Antarctica remain negative after 1990, counteracting the other components that contribute to sea-level rise. As a result, the projected maximum rise of 0.8 cm in 2005 is less than the projected current rate of 1.8 mm/yr (Nerem et al., 1997) which would lead to a rise of 2.7 cm over the same period.

In conclusion, projections of sea-level over the next several decades, as shown in Figures 2 and 3, are likely to be under-estimated at low climate sensitivity. Accordingly, the range of projected latent sea-level rise from Figures 2 and 3 will be restricted to the mid to high case which offer a better representation of historical sea-level rise, and therefore, of continuing trends.

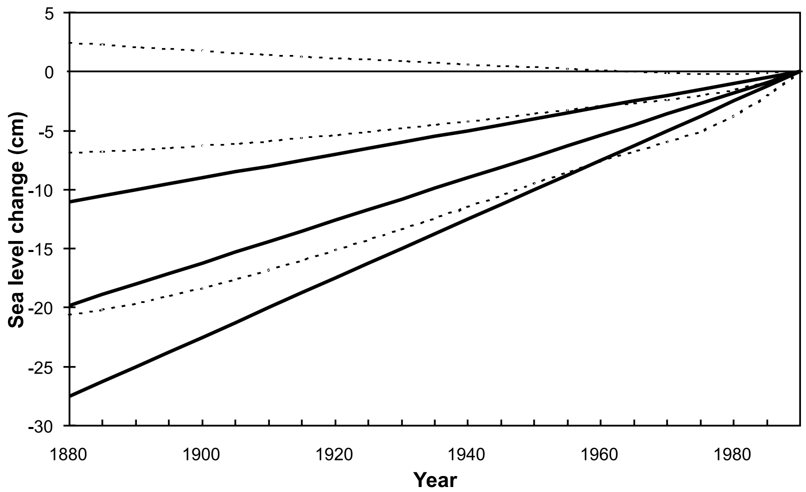

Figure 6: Comparison of the relationship between historical sea level estimates (solid lines) and sea level calculated by MAGICC (dashed lines) for the IPCC (1996) low, mid and upper estimates of climate sensitivity.

Scenario	Climate sensitivity (°C)	Thermal expansion (cm)	Glacial (cm)	Greenland (cm)	Antarctic (cm)	Total (cm)	Estimated sea-level rise (cm)
Low	1.5	2.6	0.7	0.1	-5.8	-2.4	11.0
Mid	2.5	3.8	2.7	0.6	-0.7	6.9	19.8
High	4.5	5.5	8.0	1.6	5.7	20.6	27.5

Table 2: Estimates of the various contributions to historical sea level change from 1880–1990 (110 years) as estimated by MAGICC compared with estimated historical sea-level rise over the same period.

5. Conclusion

This report aims to identify latent sea-level rise within the global climate system due to:

- anthropogenic emissions of greenhouse gases from pre-industrial times to the present; and
- projected anthropogenic emissions of greenhouse gases from pre-industrial times to 2020 given that the Kyoto Protocol emission targets are met.

Scenario 1 investigated latent sea-level rise resulting from emissions to 1995, based on a combination of estimated CO_2 emissions from fossil fuel and the IS92a scenario. Applying a low, mid and high climate sensitivity, a range of sea-level rise from 1–12 cm is projected, reaching a peak during the period 2005–2025, with the higher estimates taking longer to peak. However, taking historical trends into account, along with indications that MAGICC under-estimates these trends for the low case, a range of **5–12 cm** is estimated, peaking at about 2020–2025.

Scenario 2 applied estimated reductions of the Kyoto Protocol to Annex 1 countries to the IS92a through to 2020 and IS92a emissions for all other countries until 2020. The full range of projected latent sea-level rise from Scenario 2 was 3–32 cm but when modified with regard to historical trends as above, becomes **14–32 cm**. The peak for the mid case is reached by 2050 but at high climate sensitivity, sea-level projections are still rising at 2100.

As outlined earlier, the scenarios applied in this case are not intended to resemble or endorse any particular emissions strategy. They have merely been applied in order to determine the latent sea-level rise incurred by historical emissions in the first case, and a feasible scenario to 2020 incorporating the Kyoto Protocol, in the second case.

These projections also show that as the volume of greenhouse gases emitted into the atmosphere increases, the inertia of systems contributing to sea-level rise also increases, due to greater disequilibrium between atmospheric forcing and the state of the oceans and cryosphere. The quantity of latent sea-level rise in the global system will increase until emissions are reduced sufficiently to allow atmospheric concentrations of greenhouse gases to stabilise. When this occurs, latent sea-level rise plus any net forcing at stabilisation will contribute to further sea-level rises which may continue for centuries (e.g. Warwick et al., 1996: Figs 7.12 and 7.13).

Latent sea-level rise committed to by past emissions has the potential to threaten all regions where coastal impacts are currently marginal to severe. By 2020 the amount of latent sea-level rise may be sufficient to increase the vulnerability of regions where impacts are currently infrequent or not particularly severe, and to increase the severity of impacts in areas currently under threat.

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SPREP's Climate Change and Integrated Coastal Management Programme

Analysis of the Effects of the Kyoto Protocol on Pacific Island Countries

Part Two

Regional Climate Change Scenarios and Risk Assessment Methods

A research report prepared for the South Pacific Regional Environment Programme

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Executive summary

Introduction

The aim of this study is to survey the effect of the Kyoto Protocol and construct regional climate change scenarios for the South Pacific on behalf of the South Pacific Regional Environmental Programme (SPREP). This project has also been integrated into the Pacific Island Climate Change Assistance Programme (PICCAP), through the preparation of scenarios of temperature and rainfall for four regions covering Micronesia, Melanesia, and north and south Polynesia (Figure 1) for inclusion into the PACCLIM climate scenario generator.

This summary presents regional climate change scenarios for the South Pacific and the major conclusions from the body of the report. Quantitative regional scenarios are provided for changes to average temperature and half-yearly rainfall (May to October, November to April). These scenarios are based on projections from IPCC (1996) and output from a number of global climate models (GCMs). For some climate variables, limits to scientific knowledge allow only qualitative conclusions to be made. A section on climate change risk assessment and how risk may be affected by the Kyoto Protocol is also included.

The results presented here have varying degrees of confidence:

1. Low confidence indicates that modelling has produced an impact but the result is either preliminary, i.e. it has not been widely tested, or is only seen in a limited number of simulations. Much caution is warranted when making projections based on these results.

2. Moderate confidence indicates that the results have been reproduced a number of times in studies and/or are considered to be well-grounded in theory. Moderate confidence is equivalent to an 'each way' bet.

3. High confidence indicates that the results are considered to be robust. They have been found in a number of studies and occur under a broad range of conditions within model simulations. These outcomes are considered more probable than not.

No change is assumed where there is no information given due to a lack of evidence. Where no change is indicated with a degree of confidence, the status quo is supported by modelling evidence.

Implications of the Kyoto Protocol

The Kyoto Protocol (KP) was studied in detail to determine how it may affect the science of climate change. Several Articles in the Protocol allow 1990

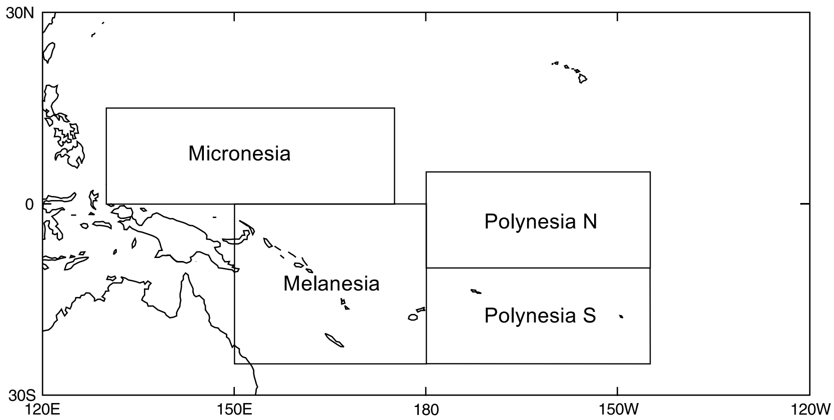


Figure 1: Location of the Micronesia, Melanesia and Polynesia (North and South) regions for which scenarios have been prepared

greenhouse gas baselines and post-1990 emissions to be altered under specified circumstances. Changes in the mixture of greenhouse gases will also lead to changes in the relative forcings per molecule of those gases, which will alter how they are represented in terms of the equivalent forcing of CO_2 as required by the Protocol. These factors will have a *very small* effect on how climate is represented in simple climate models compared to the effect that mitigation under the Kyoto Protocol has on global warming.

Three greenhouse gas emission scenarios IS92a, c and e were modified according to the Kyoto Protocol with the reductions of 5.2% for Annex 1 countries being applied until 2100 (Figure 2). Concentrations of the three major greenhouse gases in Annex 1 country emissions, CO_2, CH_4 and N_2O, were reduced by 2% in 2005, 5% in 2010 and 5.2% thereafter. Historical CO_2 emissions and IS92a values for CH_4 and N_2O were used for 1990 and 1995 in all three scenarios.

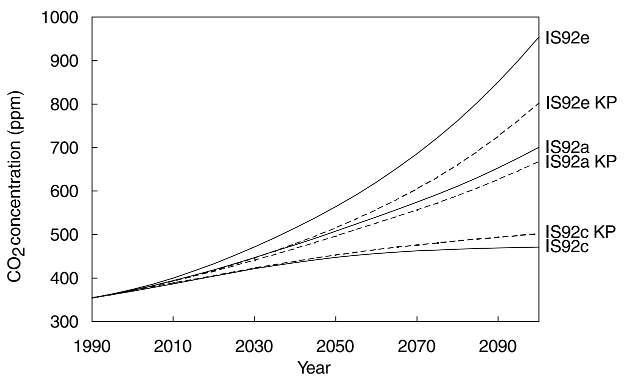


Figure 2: IPCC (1996) range of carbon dioxide (CO_2) concentration scenarios in parts per million (ppm) with Kyoto Protocol modified scenarios. IS92c is the low scenario, IS92a is the median scenario and IS92e is the high scenario

When applied to the upwelling-diffusion energy-balance model, MAGICC, global warming projections from IPCC (1996) increase slightly for the IS92c (low) scenario, and decrease for the IS92a (mid) and IS92e (high) scenarios (Table 1). The IS92c temperatures increase because Annex 1 emissions in these scenarios are actually lower than those specified by the Kyoto Protocol. An anomaly is also produced by the highest emission scenario IS92e which shows a lower range for 2050. This is due to the extremely high loading of sulphate aerosols which depresses the temperature in MAGICC, due to the unequal timing of cooling produced by sulphate aerosol (instantaneous) and global warming processes (delayed). However, a reduction in the use of fossil fuel would be expected to also reduce sulphate aerosol emissions, a feedback which has not been introduced at this time.

Global sea-level rise projections from IPCC (1996) increase slightly for the IS92c (low) scenario, decrease slightly for the IS92a (mid) and more substantially for the IS92e (high) scenario (Table 2). The IS92c sea levels increase for the same reason as for global temperature. A similar anomaly exists for the KP modified IS92e results for sea-level rise as described above for temperature.

For the purposes of forward planning or impact research we advise the use of the existing temperature and sea-level projections of the IPCC (1996) until projections based on new IPCC scenarios become available.

This is for the following reasons:

- The changes simulated by the KP modified IS92a–f emissions scenarios are small (Tables 13 and 14);

Scenario	2050		2100	
	KP modified range	Difference	KP modified range	Difference
IS92c	0.6 to 1.2	<+0.1	1.0 to 2.2	+0.1 to +0.2
IS92a	0.6 to 1.2	-0.1	1.3 to 2.7	-0.1 to -0.3
IS92e	0.5 to 1.0	-0.2 to -0.3	1.4 to 2.8	-0.4 to -0.7

Table 1: Projected changes in global temperature (°C) due to the Kyoto Protocol for the IS92a, IS92c and IS92e emission scenarios for 2050 and 2100

Scenario	2050		2100	
	KP modified range	Difference	KP modified range	Difference
IS92c	6 to 37	<+1	14 to 75	+1 to +3
IS92a	7 to 37	-1 to -2	18 to 82	-2 to -5
IS92e	7 to 35	-2 to -4	19 to 82	-5 to -12

Table 2: Projected changes in sea level (cm) for the IS92a, IS92c and IS92e emission scenarios for 2050 and 2100

- There are a number of uncertainties surrounding possible feedback effects as sulphate aerosol emissions would reduce with the reduced use of fossil fuels under the Kyoto Protocol; and

- The IPCC is currently developing new scenarios to replace the existing IS92a–f scenarios.

Therefore, recommended projections are 0.5–1.3°C in 2050 and 0.9–3.5°C in 2100 for temperature. Sea level projections are 6–40 cm in 2050 and 13–94 cm in 2100 (Table 3).

Regional climate change scenarios

Six coupled ocean-atmosphere climate simulations were included in the analysis of regional climate change scenarios: CSIRO Mark 2 GCM with and without sulphates, CSIRO DARLAM 125 km, DKRZ ECHAM4/OPYC3 GCM, Hadley Centre HADCM2 and the Canadian CGCM1 (Table 5: main report).

Regional scenarios for changes to average temperature and half-yearly rainfall have been created from scaled patterns from these models. The scaling technique is a recent innovation that extracts the relationship between each grid box value and global temperature over a transient simulation to obtain a local greenhouse signal free as possible of extraneous influences such as multi-decadal variability. These individual model patterns have been aggregated into a range of local change per degree of global warming.

These regional scenarios can be considered as projections that represent a sizeable range of possible future climates. However, the probability of these changes relative to other outcomes remains unknown, so they cannot be regarded as forecasts. Other possible changes that should also be considered but cannot be quantified include rapid climate change, and unexpected outcomes from processes not fully understood.

Temperature

When observed and simulated model temperatures under current climate are compared, GCMs reproduce spatial temperature patterns reasonably well when model resolution is taken into account. The large grid sizes compared to the size of Pacific Island countries restrict model output to the representation of marine air temperature, so issues relevant to detailed island climatology have not been solved.

Under climate change, regional warming is mostly below the level of average global warming, as would be expected over the ocean. Higher latitudes of the southern ocean show the least warming in the SPREP region, while the greatest warming tends to occur in the far west and the central and eastern equatorial Pacific.

Global warming	*Low*	*Mid*	*High*
2050	0.5°C	0.9°C	1.3°C
2100	0.9°C	2.0°C	3.5°C
Sea-level rise			
2050	6 cm*	20 cm	40 cm
2100	13 cm	49 cm	94 cm

Table 3: Global warming and sea-level rise scenarios for 2050 and 2100 (IPCC, 1996)

** Note that in Part One of this project (Jones, in press) the low sensitivity outcome for latent sea-level rise was rejected on the basis of evidence from historical sea-level rise. The whole range is included here, but a sea-level rise in the low to mid range is considered less likely than a rise in the mid to high range.*

Region	**Local warming per °C of global warming**	**Warming in 2050**			**Warming in 2100**		
		low	*median*	*high*	*low*	*median*	*high*
Micronesia	0.7 to 1.0	0.4	0.8	1.3	0.6	1.6	3.5
Melanesia	0.7 to 0.9	0.4	0.8	1.2	0.6	1.6	3.2
Polynesia N	0.8 to 1.0	0.4	0.8	1.3	0.7	1.6	3.5
Polynesia S	0.7	0.4	0.7	0.9	0.6	1.4	2.5

Table 4: Scenarios of temperature change (°C) for regions defined by the PICCAP Project

Regional scenarios based on the range of output from Table 6 in the main report, are shown in Table 4 here, with ranges defined by the lowest local warming per degree of global warming multiplied by the lowest global warming in Table 3 and the highest local warming multiplied by the highest global warming for both 2050 and 2100. A median point has been included, which is the median result multiplied by the mid global warming of Table 3. A high confidence is attached to this range.

Rainfall

Present rainfall patterns over the Pacific, mainly defined by the Intertropical Convergence Zone (ITCZ) and the South Pacific Convergence Zone (SPCZ), were broadly captured by four of the models. The other two produced less realistic results but all were deemed satisfactory for further analysis. All models show an increase in rainfall over the central and eastern Pacific over both half-years of May to October and November to April. Changes over other regions were smaller, tending towards increases. Movements of both the ITCZ and the SPCZ were noted but were not consistent between models. Rainfall invariably increases where warming over the ocean is greatest.

Regional scenarios for rainfall were difficult to construct, given the sometimes wide range of results. For these reasons, a median value has been included. When the range lies on either side of zero, the outliers are multiplied by the highest global warming for 2050 and 2100, while the median is multiplied by the mid warming from Table 5. Where both low and high extremes are on the same side of zero, the value closest to zero is multiplied by the lowest global warming temperature. The results are rounded to the closest 5%.

The May to October results are shown in Table 5, and results for the November to April period are shown in Table 6. Outliers can be noted in Polynesia N at the high end and Polynesia S at the low end. The November to April results contain no significant outliers. The median values show that most of the scenarios lean towards increases in rainfall, although the possibility of decreases cannot be ruled out in some regions, particularly during the November to April period. The most notable rainfall change is for Polynesia N during both seasons, with substantial changes in the median result. A moderate confidence is attached to the sign of rainfall change in Polynesia N, and a low confidence is attached to the sign of rainfall changes elsewhere.

Rainfall variability

The following types of rainfall variability were surveyed:

Interannual variability – an analysis of a single GCM showed no increase in rainfall variability between 1960 to 2100. Model variability is much lower than historical variability, because the model

Region	Response per °C of global warming			Change in 2050			Change in 2100		
	low	*median*	*high*	*low*	*median*	*high*	*low*	*median*	*high*
Micronesia	0	6	10	0	5	15	0	10	35
Melanesia	-4	2	6	-5	0	10	-15	5	20
Polynesia N	4	13	43	0	10	55	5	25	150
Polynesia S	-8	1	2	-10	0	5	-30	0	5

Table 5: Scenarios of rainfall change (%) for regions defined by the PICCAP Project for May to October, rounded to the nearest 5% for 2050 and 2100

Region	Response per °C of global warming			Change in 2050			Change in 2100		
	low	*median*	*high*	*low*	*median*	*high*	*low*	*median*	*high*
Micronesia	-2	1	4	-5	0	5	-5	0	15
Melanesia	-3	2	6	-5	0	10	-10	5	20
Polynesia N	6	14	20	10	15	26	20	30	70
Polynesia S	-6	1	3	-10	0	5	-20	0	10

Table 6: Scenarios of rainfall change (%) for regions defined by the PICCAP Project for November to April, rounded to the nearest 5% for 2050 and 2100

does reproduce the magnitude of ENSO fluctuations and the rain is simulated in a grid box rather than at a point. A low confidence is attached to this result.

Decadal variability – decadal and multi-decadal variations are produced by the CSIRO and other GCMs. Analysis of their behaviour under climate change has yet to be conducted. However, methods need to be developed so that they can be explicitly incorporated into climate scenarios.

Rainfall intensity – increases in daily rainfall intensity are expected in many regions where rainfall increases, remains the same or decreases slightly. Even with an appreciable decrease in average rainfall, reductions in daily rainfall intensity can be negligible (Figure 19). The results concerning rainfall intensity come from a number of models and studies, so have a high confidence attached.

Sea-level rise

Historical sea-level rise over the Pacific from tide gauge records adjusted for post-glacial rebound is consistent with global estimates. Due to the preliminary nature of studies regionally varying sea-level change, **it is recommended that projections of global sea-level rise from IPCC (1996) and shown in Table 3 continue to be used in impact and adaptation assessments for the time being.**

ENSO

ENSO is the dominant influence on climate variability in the Pacific, producing large oscillations in temperature, winds, rainfall, sea level, surface pressure and a number of other variables. Therefore, knowledge of how ENSO changes under climate change is a key to understanding climate change impacts in the Pacific region. Analyses of model output show that the ENSO phenomenon is likely to continue out to 2100. Patterns and analyses of temperature and rainfall produced for this study show that the GCMs produce a more El-Niño-like mean state over the Pacific under climate change. Rainfall increases are also distributed in a El-Niño-like pattern but they generally increase over most of the Pacific, except where warming is least. These analyses need to extend from the NINO3 and NINO4 regions which are ±5° from the equator, to the whole of the Pacific to obtain a broader picture of how ENSO may change.

Tropical cyclones

Numbers – there is no evidence that tropical cyclone numbers may change.

Intensities – a general increase in tropical cyclone intensity, expressed as possible increases in wind speed and central pressure of 10–20% at the time of CO_2 doubling, now appears likely. How this affects the risk posed by severe storms needs to be determined on a regional basis. A moderate confidence is attached to this result.

Regions of formation – no significant change in regions of formation were noted in the DARLAM 125 km resolution simulation, although it is possible they may change in response to long-term changes to ENSO.

Regions of occurrence – There appear to be no major changes in regions of occurrence except to note that tropical cyclones may track further polewards. A low confidence is attached to this result.

Impacts and risk

Impacts

This summary is restricted to impacts that may be directly affected by the outcomes of the previous section. A more extensive review is in Section 4.1 of the main report.

Coastal areas will continue to be affected by ENSO variability, tropical cyclones and wave climates. ENSO appears likely to continue, although how it may manifest remains uncertain. Tropical cyclones may become 10–20% more intense at $2\times CO_2$ occurring roughly between 2030 and 2060 which would be likely to increase storm surge heights. There is no information about how wave climatologies may or may not change on a seasonal basis. Sea-level rise is additive to ENSO sea-level variability, surge and swells. Therefore current risks are likely to persist and probably increase at least at the rate determined by sea-level rise.

Over the past twelve months, scientific understanding of the status of coral reefs under climate change has altered. There is evidence of a likely reduction in the calcification rates of benthic calcareous organisms (coral and algae) due to the increased acidity of sea water under higher CO_2 levels in the atmosphere. This has changed the attitude of marine researchers from one of assuming reefs were generally fairly resilient under climate change to one of increased concern (Done, pers. comm.). Calcification may reduce the rate of reef growth, strength, or both, making reefs less resilient to sea-level rise and reducing the sediment supply to beaches and islands. Together with threats from turbid and nutrient runoff, coral bleaching and over-use, coral reefs are now considered to be more vulnerable than previously believed.

Intense rainfall on a daily, monthly and seasonal basis may increase in the northern Polynesian region and further east, in conjunction with large increases in average rainfall. Similar but smaller changes may also occur elsewhere. A positive outcome would be an increase in water supply due to higher recharge in the relevant regions, particularly where groundwater lenses are important. Increases in extreme rainfall frequency would lead to more flooding. This may affect:

- planning of drainage works;
- zoning of industrial and residential areas;
- dam design;
- agricultural risk management;
- sewage disposal;
- terrestrial sediment and nutrient pollution of reefs;
- land degradation; and
- transport and telecommunication infrastructure.

Risk

A risk assessment framework has been introduced and an example of risk analysis carried out. The framework involves a focus on impact outcomes via the construction of impact thresholds. This requires input from stakeholders, ie. from those who manage and use the resources impacted upon. Stakeholders are also central to the introduction and analysis of adaptation options. The framework proposes a scientifically-based risk assessment process for climate change with the potential to be integrated with existing adaptation programs in the Pacific, such as integrated coastal zone management.

A risk analysis, comparing risk before and after the implementation of the Kyoto Protocol, was carried out on two thresholds for a hypothetical coastline: a 50 cm sea-level rise, and an atmospheric concentration of CO_2 of 560 ppm (associated with a possible reduction of calcification rates in reef communities). The risk of both these thresholds being exceeded was 16% in 2075 and 44% in 2100. When analysed for combined risk under the IS92a–f scenarios and KP-modified scenarios in 2075 and 2100, the risk under the Kyoto Protocol was reduced by 9% and 6% respectively. This shows that for these two impacts, implementing the Kyoto Protocol reduces the risk by <10%, or alternatively, delays it by less than a decade. The results are shown in Table 7.

Year	IS92a-f	KP modified IS92a-f	Difference
2075	16	7	-9
2100	44	44	-6

Table 7: Risk of threshold exceedence (%) for the combined risk of a sea-level rise of 50cm and an atmospheric CO_2 content of 560 ppm 2075 and 2100 according to the IS92a–f and IS92a–f Kyoto Protocol modified scenarios

1. Introduction

1.1 Context

This report has been prepared by CSIRO Atmospheric Research for the South Pacific Regional Environment Programme (SPREP; Figure 1). The main objective of the report is to improve the regional understanding of the science of climate change and to analyse how this could be affected by the United Nations Framework Convention on Climate Change (UNFCCC) process and the Kyoto Protocol in the South Pacific region. The Kyoto Protocol was agreed to by the Third Session of the Conference of the Parties (COP3) on 11 December 1997 and is now open for signing and ratification by the Parties, including Pacific island countries. So far seven Pacific island countries have signed the protocol and two, Fiji and Tuvalu, have ratified it (as of 15 January 1999).

To assist the Pacific island governments, the SPREP Secretariat and the Alliance of Small Island States (AOSIS) to develop sound policies to implement the UNFCCC and the Kyoto Protocol, CSIRO Atmospheric Research has reviewed and analysed the Kyoto Protocol and its implications for the science of climate change. This analysis modifies the IS92a–f emission scenarios according to the Kyoto Protocol, then estimates the resultant effect on global warming and sea-level rise using the MAGICC upwelling-diffusion energy-balance model. The analysis and results are presented in Section 2 of the report.

Section 3 takes output from a selection of Global Climate Models (GCMs), performs analyses, and compares these to existing climatologies for the Pacific region. The information gained from these analyses is used to develop regional climate change scenarios. As part of the Pacific Island Climate Change Assistance Programme (PICCAP), scenarios of temperature and rainfall are also prepared for four regions covering Micronesia, Melanesia, and north and south Polynesia. Possible changes to climate variability, and extreme events covering such aspects as rainfall variability, tropical cyclones and El Niño–Southern Oscillation (ENSO) are also examined.

Section 4 summarises impacts on South Pacific islands with reference to the scenarios presented in Section 3. A framework for the risk assessment of climate change impacts is also outlined. An example, analysing the effect of the Kyoto Protocol on the risk of climate change on a hypothetical coastline, is presented to demonstrate how the framework operates. Adaptation within the context of the risk assessment is discussed.

Finally, Section 5 provides a synthesis of the results presented in the report and makes recommendations for further research arising out of the study.

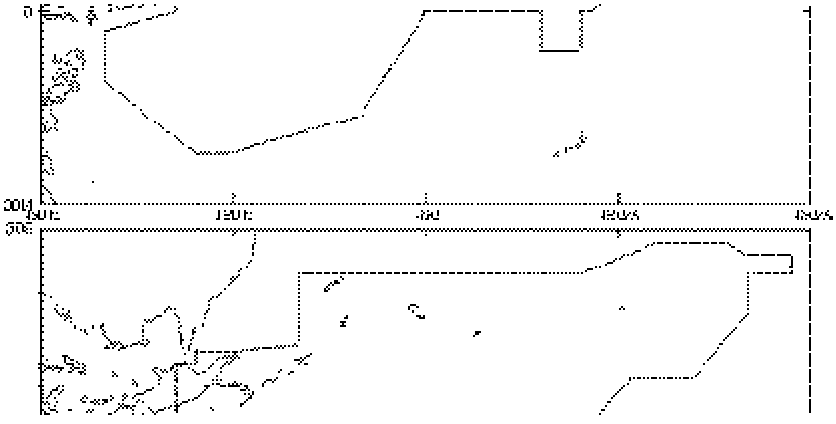


Figure 1: Location of the South Pacific Regional Environment Programme (SPREP) region

1.2 Limitations of this study

The science of climate change is complex and can be very difficult to understand. Climate change is affected by a number of complex systems including the atmosphere, the biosphere and the oceans. Additionally, the activities of over five billion humans affect these systems on a global scale. Due the degree of effort being put into understanding the science of the enhanced greenhouse effect and its relative youth, the scientific understanding can change very quickly. For these reasons, although we have represented the science as best as we can, a number of caveats apply to the use of the information contained in this report.

1.2.1 Use of scenarios

Great care must be attached to the use of climate change scenarios. The *IPCC Technical Guidelines for Assessing Climate Change Impacts and Adaptations* defines a scenario as 'a coherent, internally consistent and plausible description of a possible future state of the world' (Carter et al., 1994). Although the major aim of constructing scenarios is broadly predictive, great care has been taken in the past to ensure that climate change scenarios are not used as forecasts (e.g. IPCC, 1996; CSIRO, 1996). This is because they are limited representations of the future and do not adequately represent all possibilities.

Scenarios are proposed with no information about the probability of their occurrence, so are not intended for use as forecasts but for the diagnosis of certain conditions. They allow various questions to be answered that meet the following format – 'What happens if a certain chain of events is followed?'. Their major use in impact studies is to determine sensitivity and vulnerability.

Projections are possible future outcomes based on modelling results that utilise a number of scenarios. Rather than representing the most likely evolution of climate in the future, they are aimed at estimating and understanding the response of the climate system to a range of possible forcing scenarios (IPCC, 1996). IPCC (1996) includes a number of projections based on the full set of IS92a–f emission scenarios, including global temperature and sea-level rise (see Section 2). Despite representing a substantial range of possible future emission pathways, the probability of such outcomes is still largely unaddressed, although Mahlman (1997) makes an attempt to attach probabilities to various outcomes of climate change modelling.

As such, projections are normative scenarios that try to describe what the world may be like in the future based on current knowledge. It may be argued that this type of scenario is a forecast. However, the projections in IPCC (1996) are constructed using only a limited number of the uncertainties known to affect climate change. For instance, projections of temperature and sea level contain ranges of uncertainty due to climate sensitivity and emissions scenarios but do not contain the uncertainties associated with the radiative forcing of greenhouse gases and sulphate aerosols, or gas mixing models that produce concentrations of specific gases within the atmosphere (Visser et al., in press). Furthermore, the possibility of rapid, non-linear climate change is not incorporated in such scenarios, and is difficult to incorporate with our current state of knowledge, although we do know that such events are a relatively frequent aspect of past climate changes (e.g. Schaaf and Thurow, 1997; Jones et al., 1998).

The boundary between a scenario and projection is unclear, as the two are a continuum. The difference between the two relies largely on our confidence in the science that shaped them.

We ask users of the information contained within this report to bear these caveats in mind. Some guidance will be given on the basis of expert judgement but users are reminded that many limitations exist. For instance, a number of models have been used in projecting regional changes in temperature and rainfall but they do not include all models, nor do models incorporate all of the effects which we know may influence climate. It is not possible to assess the likelihood of occurrence of the ranges of local warming or rainfall change produced in Section 3.

Where results are based on the analysis of a single model, this certainly should not be assumed to be a projection, unless it is supported by other modelling and a statement expressing a specified degree of confidence in the outcomes is made.

1.2.2 Uncertainty

The issue of uncertainty also introduces limitations. There are two characteristics of uncertainty relevant to climate change studies: source and type. A source of uncertainty is defined by a single process, or set of related processes, often represented by a computer model. For example, in calculating global temperature, there are at least four sources of uncertainty: greenhouse gas emission scenarios, atmospheric mixing of gases, radiative forcing of greenhouse gases and aerosols, and climate sensitivity (Visser et al., in press). If we wish to continue the analysis to impacts, further sources of uncertainty include regional changes to climatic variables, biophysical uncertainties and socio-economic uncertainties.

Types of uncertainty cut across these sources. Knowing the type of uncertainty often indicates whether it can be limited through dynamic modelling, statistical analysis, or may be fundamental to a system (e.g. chaos). Types of uncertainty include incomplete scientific knowledge; behavioural uncertainty, which includes social, political and economic behaviour; and chaotic effects. The latter occur in complex systems where small differences in initial conditions, or minor changes, can produce large, non-linear outcomes (sometimes called the butterfly effect).

In Section 4.2 we outline a risk assessment framework that aims to manage these uncertainties. Instead of a linear framework, where impact scientists and policymakers act upon scenarios or projections of climate change produced by climate scientists, we propose a framework where users survey the broad scenarios for climate change, determine how they affect their own area of interest, then suggest outcomes which would be desirable, or undesirable. Such outcomes, or thresholds, may represent a level of profitability in various activities, conservation of species and habitat, continuance of cultural values, or on the negative side, conditions where a desired activity can no longer operate.

Where these thresholds can be linked to climatic variables, and scenarios can be created with more or less a degree of confidence, risk analysis can be performed. This process requires a similar level of involvement by climate modellers as is occurring now, but the results are more attuned to the needs of impact assessments. Further assessment can then examine adaptations or the need for mitigation in accordance with the Articles of the UNFCCC (Section 4.2).

2. The effect of the Kyoto Protocol on greenhouse gas emission and climate scenarios

The Kyoto Protocol, agreed to by the third Conference of Parties of the UNFCCC on 11 December 1997, requires Annex I parties of the UNFCCC to reduce their emissions of six greenhouse gases by at least 5 per cent by 2008 to 2012. Those greenhouse gases are CO_2, CH_4, N_2O, hydrofluorocarbons, perfluorocarbons and SF_6.

These six gases have different global warming potentials on a molecule-by-molecule basis, i.e. some lead to greater warming than others. Under the protocol, these differences are to be measured by converting the individual global warming potentials of each gas into carbon dioxide equivalents. The equivalent forcings of these six gases in total are to be reduced by at least 5 per cent by 2008 to 2012. This allows a certain amount of flexibility, as individual nations will find some gases easier to reduce than others.

A planned reduction in the emissions of greenhouse gases will lead to lowered projections of climate change when compared to 'business as usual' emission scenarios. However, quantifying these reductions is not straightforward. For instance, an improved understanding of the science surrounding greenhouse gas inventories during the period of the protocol may require the revision of past emission profiles, altering the 1990 baseline estimates for Annex I countries from which the protocol requirements are set.

The relative balance between different greenhouse gases in the atmosphere will also affect their global warming potentials (the measure which the protocol will use to calculate the CO_2 equivalent of different combinations of the target greenhouse gases) over the period the protocol is in effect. The relative forcings of greenhouse gases may also be changed by improvements in scientific knowledge.

Article	Description	Effect
3.1	The Parties shall ensure that their aggregate anthropogenic carbon dioxide equivalents of the greenhouse gases in Annex I are reduced by at least 5% by the commitment period 2008– 2012.	Carbon dioxide equivalents will change depending on the composition of the atmosphere (e.g. Wigley, 1998)
3.3	The net changes in greenhouse gases from sources and sinks relating to afforestation, reforestation and deforestation shall be used to meet the commitments of the Kyoto Protocol.	Calculating these changes will alter the current estimated 1990 baseline emissions of Annex I countries and results are subject to large uncertainties.
5.2	Methodologies for estimating anthropogenic emissions from sources and sinks of greenhouse gases not controlled by the Montreal Protocol (e.g. CO_2, CH_4, N_2O) shall be those agreed to by the Intergovernmental Panel on Climate Change (IPCC) and the COP. These methodologies can be reviewed and revised but only for the purposes of ascertaining compliance.	Revisions in methodologies will change both estimates of 1990 carbon stocks and subsequent changes to all greenhouse gas emissions.
5.3	As for 5.2, but for the calculations of greenhouse gas equivalence.	Revisions of methodologies will change post-1990 emission histories.

Table 1: Summary of Articles within the Kyoto Protocol that affect the simulation of projected climate change

The major uncertainties within the protocol that relate to the projections of future climate change are listed in Table 1. As indicated by the simulations presented below, these uncertainties are likely to be small compared to the overall effect of the Kyoto Protocol.

2.1 Greenhouse gas emission scenarios

The MAGICC upwelling-diffusion energy-balance model developed by Wigley and Raper (1992, 1993, 1997) is used to estimate the effects of the Kyoto Protocol on global temperature and sea level. This simple model, tuned to the outputs of more complex GCMs, takes scenarios in the form of emissions and produces estimates of atmospheric greenhouse gas concentrations, global warming and sea-level rise. The version is that used in the IPCC second assessment report (Houghton et al., 1996).

This version of MAGICC is limited in that only CO_2, CH_4, N_2O and sulphate aerosols can be changed explicitly and the halogen-related compounds subject to the Montreal Protocol are fixed. Therefore the effects of the protocol can only be simulated for three of the six greenhouse gases subject to the Kyoto Protocol. However, as CO_2 comprises about 85 per cent of the total forcing, CH_4 about 10 per cent, N_2O about 7 per cent and total halogens only 6–8 per cent (comprising those covered under the Montreal Protocol and those under the Kyoto Protocol), it was felt that the explicit modelling of CO_2, CH_4 and N_2O would suffice. (Note: the total is over 100 per cent as it allows for the negative forcing of sulphate aerosols and other gases).

Three greenhouse gas emission scenarios, IS92a, c and e (Figure 2) were modified according to the Kyoto Protocol. The versions used were those used in IPCC (1996) where the CO_2 data had been updated from the earlier IS92a–f scenarios with CH_4 and N_2O remaining unchanged. 1990 and 1995 Annex I country and global emissions for CO_2 for all three scenarios were based on historical estimates from Wigley (1998). The 1990 and 1995 estimates of Annex I CH_4 and N_2O emissions were based on the IS92a scenario, on the assumption that historical emissions during the 1990s have resembled the IS92a scenario most closely of all the IS92a–f scenarios.

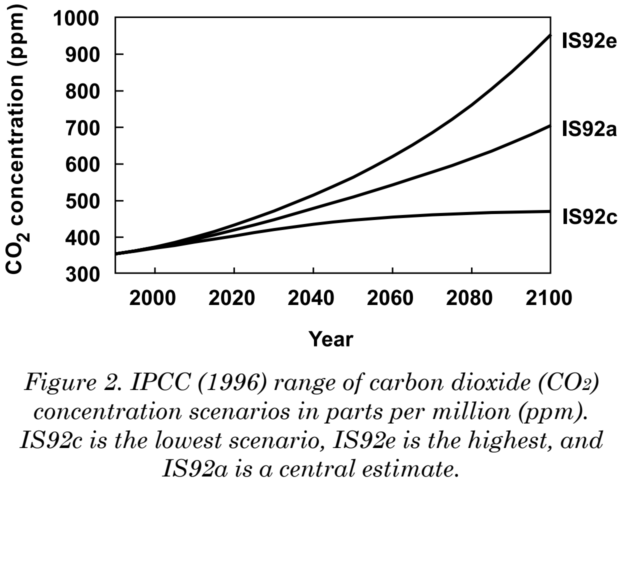


Figure 2. IPCC (1996) range of carbon dioxide (CO_2) concentration scenarios in parts per million (ppm). IS92c is the lowest scenario, IS92e is the highest, and IS92a is a central estimate.

All emissions were partitioned into Annex I and developing countries, with the Annex I total emissions being reduced by 2 per cent in 2005, 5 per cent in 2010 and 5.2 per cent from 2015 to 2100. (Note: While developed countries are listed in Annex I of the UNFCCC, in the Kyoto Protocol, they are included under Annex B.) These were added to the rest of the world emissions from 2005 to estimate global emissions. For the IS92a scenario, estimates of Annex I CO_2 and rest of the world emissions were obtained from Wigley (1998). Annex I emissions for the updated IS92c and e scenarios were not available, so they were taken as a percentage of the total emissions from the 1992 versions and recalculated for the updated CO_2 emissions in the 1995 versions. Estimates of Annex I CH_4 and N_2O emissions were modified from the 1992 data (Table 2). As for the analyses carried out for the IPCC Second Assessment Report (Kattenberg et al., 1996), the effects of CO, NOx and VOCs are omitted and CFCs are restricted to obey the Montreal Protocol.

These scenarios assume the stabilisation of Annex I emissions at 5.2 per cent below 1990 levels for the period following that covered by the Kyoto Protocol, i.e. from 2015 to 2100. This is in contrast to three post-2012 pathways investigated by Wigley (1998): no more emission reductions after 2012 (i.e. emissions of Annex I countries recommence, parallel to the IS92a scenario), stabilised Annex I emissions as above and stabilised Annex I emissions with a further annual 1 per cent (compound) reduction. The choice of applying a stabilised 5.2 per cent reduction for Annex I countries from 2015 is to investigate the effect of the Kyoto Protocol if it is implemented and maintained over the long term.

As fossil fuel burning is reduced as a part of emission reductions, sulphate aerosol emissions will also be reduced. This will reduce the negative forcing imposed by sulphates, partly compensating for any reduction in temperature coming from greenhouse gas mitigation. While it would be valuable to estimate how large this effect may be, the uncertainties surrounding radiative forcing make this a difficult task, requiring more resources than available for this project. For example, the direct effects of radiative forcing are -0.3 W m^{-2} with an uncertainty range of -2 to 0.8 W m^{-2}, and indirect effects are -0.8 W m^{-2} with an uncertainty range of 0.0 to 1.5 W m^{-2} (Schimel et al., 1996).

IS92a

Year	CO_2	CO_2 mod.	CH_4	CH_4 mod.	N_2O	N_2O mod.
1990	6.11	6.11	506	506	12.9	12.9
1995	6.51	6.51	525	525	13.3	13.3
2000	7.26	7.09	545	545	13.8	13.5
2005	8.10	7.67	568	563	14.1	13.7
2010	8.84	8.11	588	580	14.5	14.0
2015	9.58	8.66	611	602	14.9	14.3
2020	10.42	9.31	636	625	15.4	14.7
2025	11.26	9.97	659	651	15.8	15.0
2050	13.86	12.57	785	762	16.6	15.9
2075	16.31	14.58	845	801	16.7	16.1
2100	20.56	18.30	917	841	17.0	16.4

IS92c

Year	CO_2	CO_2 mod.	CH_4	CH_4 mod.	N_2O	N_2O mod.
1990	6.1	6.11	506	506	12.9	12.9
1995	5.9	6.51	510	525	13.1	13.3
2000	6.2	6.78	526	535	13.6	13.4
2005	6.5	6.66	540	548	13.8	13.6
2010	6.9	6.83	551	557	14.1	13.7
2015	7.1	7.05	563	572	14.4	13.9
2020	7.4	7.35	574	585	14.7	14.2
2025	7.7	7.71	589	601	15.0	14.5
2050	6.8	7.72	613	636	15.0	14.7
2075	5.3	7.09	584	621	14.2	14.2
2100	4.8	6.93	546	593	13.7	13.9

IS92e

Year	CO_2	CO_2 mod.	CH_4	CH_4 mod.	N_2O	N_2O mod.
1990	6.1	6.11	506	506	12.9	12.9
1995	6.9	6.51	527	525	13.4	13.3
2000	7.8	7.17	556	549	13.8	13.5
2005	8.9	7.58	581	571	14.3	13.8
2010	10.2	8.14	606	592	14.8	14.1
2015	11.4	8.92	634	619	15.2	14.5
2020	12.6	9.82	661	646	15.8	15.0
2025	14.0	11.02	692	677	16.3	15.4
2050	19.3	15.11	845	794	17.6	16.7
2075	26.6	20.75	958	851	18.1	17.3
2100	35.9	28.52	1072	921	19.1	18.3

Table 2: IS92a, c and e scenarios for CO_2, CH_4 and N_2O modified according to the Kyoto Protocol, with stabilisation of Annex I countries at 5.2 per cent below 1990 levels 2015–2100. The 1990 and 1995 values for CO_2 are from Wigley (1998) and for CH_4 and N_2O are from IS92a.

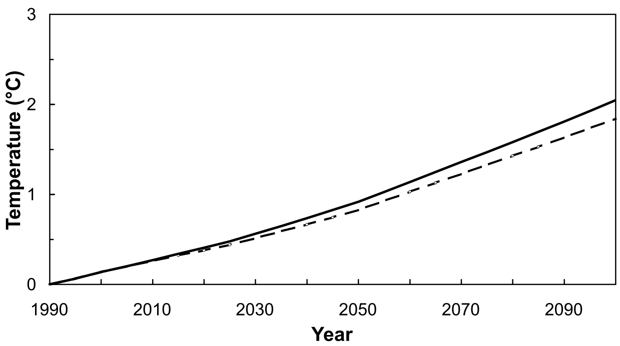


Figure 3: Projected global warming from the IS92a emission scenario and the KP modified IS92a scenario with a climate sensitivity of 2.5°C.

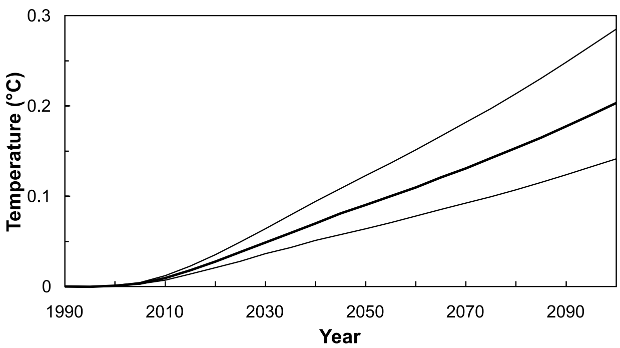


Figure 4: Differences in projected global warming between the IS92a emission scenario and the KP modified IS92a scenario at a climate sensitivity of 1.5°C (lower), 2.5°C (mid) and 4.5°C (upper).

2.2 Global temperature scenarios

Projected warming from MAGICC that utilises the IS92a–f scenarios with increasing sulphate aerosols ranges between 0.0 and 3.5°C for 2100 (IPCC, 1996). Projected global warming from the IS92a and IS92a Kyoto Protocol (KP) modified scenarios was calculated using MAGICC and is shown in Figures 3 and 4.

Figure 3 shows that the effect of the Kyoto Protocol with a mid-range global warming of 2°C by 2100 (based on the IS92a scenario), is 0.2°C. Simulations using the same emission scenario and varying climate sensitivity produce results of a similar magnitude. Figure 4 shows the difference between global warming projections from IS92a and the KP modified IS92a scenario for the climate sensitivities of 1.5°C, 2.5°C and 4.5°C producing differences ranging from about 0.1°C to 0.3°C by 2100. These are similar to the reductions produced by Wigley (1998) using the same model but with a slightly different KP modified scenario.

Possible changes across the range of IS92a–f scenarios were estimated by carrying out similar comparisons for IS92c and IS92e, the lowest and highest emission scenarios respectively. For IS92c, the temperature actually increased by 0.1 to 0.2°C by 2100 when the Kyoto Protocol was applied. The increase occurs because the IS92c scenario contains a greater degree of mitigation than the Kyoto Protocol requires for Annex I countries. IS92e, the highest scenario, showed the largest reductions, ranging from 0.4 to 0.7°C in 2100 (Table 3).

In percentage terms, temperature projections based on the IS92a scenario are reduced by about 10 per cent independent of climate sensitivity in both 2050 and 2100. The increased warming projected from the IS92c scenario is small at 2050 but rises to over 10 per cent by 2100. The decrease in the IS92e scenario is over 20 per cent for both 2050 and 2100. Modifications of halogens as per the Kyoto Protocol (hydrofluorocarbons, perfluorocarbons and SF_6) could not be made, although this will have little impact on the results as their net forcing is small.

An anomaly is produced by the highest emission scenario IS92e which shows a lower range than for IS92a and c for 2050. This is due to the extremely high loading of sulphate aerosols in IS92e which depresses the temperature in MAGICC due to the unequal timing of cooling produced by sulphate aerosols (instantaneous) and global warming processes (delayed). Therefore, this anomaly disappears by 2100 as warming outstrips the cooling effect.

A reduction in the use of fossil fuels will reduce sulphate aerosol emissions, with a positive impact on temperature. However, it is difficult to quantify this effect in context of the current project as outlined in the previous section. Doubts can also be raised about the level of sulphate aerosols assumed in the IS92a–f scenarios, especially those rates contained within the higher emission scenarios. As technology to prevent sulphate emissions is rapidly evolving, greenhouse gas and sulphate aerosol emissions are becoming increasingly decoupled. It may be unrealistic to expect such high sulphate aerosol emissions in the future.

The application of the Kyoto Protocol to Annex I countries in IS92e while emissions of the rest of the world continue to rise sharply appears unrealistic. It is also counter to the aim of stabilising greenhouse gases as required by the UNFCCC. Enting (1998) shows that applying mitigation solely to Annex I countries with the rest of the world following the IS92a emission scenario is not sufficient to lead to stabilisation.

Year	*IS92 low (°C)*	*IS92 mid (°C)*	*IS92 high (°C)*	*KP low (°C)*	*KP mid (°C)*	*KP high (°C)*	*Diff. low (°C)*	*Diff. mid (°C)*	*Diff. high (°C)*	*Diff. low (%)*	*Diff. mid (%)*	*Diff. high (%)*
IS92c												
2050	0.54	0.80	1.16	0.57	0.83	1.20	0.02	0.03	0.04	4	4	4
2100	0.87	1.32	1.97	0.99	1.48	2.20	0.12	0.17	0.23	14	13	12
IS92a												
2050	0.64	0.92	1.30	0.57	0.83	1.18	–0.06	–0.09	–0.12	–10	–10	–9
2100	1.40	2.05	2.95	1.26	1.84	2.67	–0.14	–0.20	–0.29	–10	–10	–10
IS92e												
2050	0.67	0.96	1.34	0.51	0.73	1.04	–0.16	–0.22	–0.31	–24	–23	–23
2100	1.72	2.48	3.52	1.36	1.97	2.81	–0.36	–0.51	–0.72	–21	– 21	–20

Table 3: Summary of global warming obtained from the IS92a, c and e scenarios and those modified according to the Kyoto Protocol (Table 2)

In 1996, the IPCC decided to develop a new set of reference emission scenarios for the analysis of potential future climatic changes and associated impacts, as well as a reference for the socio-economic analysis of long-term mitigation options. These are currently being prepared for release, and will provide the basis for climate modelling studies from 1999. However, it may take some time before confidence can be expressed in the outputs from these scenarios. Until such results become available, it is advisable to use IPCC (1996) projections for global warming based on output from the IS92a–f scenarios, as projected changes coming from the Kyoto Protocol are small compared to this range.

2.3 Global sea-level rise scenarios

The impact of the Kyoto Protocol on sea-level rise was examined in the same way as for temperature (Figures 5 and 6). The modelled decrease in sea-level rise for the Kyoto modified IS92a emission scenario at climate sensitivities of 1.5 to 4.5°C is 2–5 cm by 2100.

The pattern of change for the IS92c and e scenarios is similar to that for temperature. IS92c modified for the Kyoto Protocol leads to a slight increase in sea-level rise of 1–3 cm, while IS92e shows a decrease of 5–12 cm (Table 4). However, in percentage terms the differences are slightly lower than for temperature. This is due to thermal inertia in both ocean and ice systems that delays the rise of sea level as compared to the more rapid atmospheric warming.

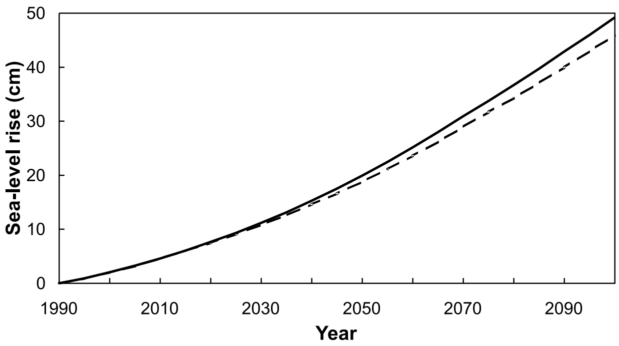


Figure 5: Projected sea-level rise from the IS92a emission scenario and the KP modified IS92a scenario with a climate sensitivity of 2.5°C

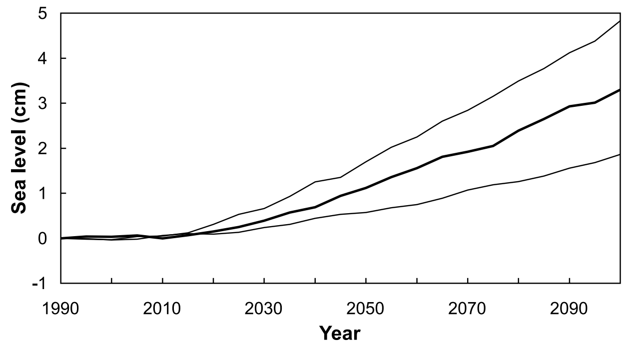


Figure 6: Differences in projected sea-level rise from the IS92a emission scenario and the KP modified IS92a scenario at a climate sensitivity of 1.5°C (lower), 2.5°C (mid) and 4.5°C (upper).

Year	*IS92 low (cm)*	*IS92 mid (cm)*	*IS92 high (cm)*	*KP low (cm)*	*KP mid (cm)*	*KP high (cm)*	*Diff. low (cm)*	*Diff. mid (cm)*	*Diff. high (cm)*	*Diff. low (%)*	*Diff. mid (%)*	*Diff. high (%)*
IS92c												
2050	6.3	18.3	36.7	6.5	18.7	37.4	0.2	0.4	0.7	3	2	2
2100	13.1	38.5	72.0	14.4	40.9	75.3	1.3	2.4	3.3	10	6	5
IS92a												
2050	7.5	19.9	38.8	6.9	18.8	37.1	–0.6	– 11	–1.7	–8	–6	–4
2100	20.0	49.2	86.4	18.1	45.9	81.6	–1.9	–3.3	–4.8	–9	–7	–6
IS92e												
2050	8.1	20.7	39.7	6.6	17.9	35.4	–1.5	–2.8	–4.3	–19	–14	–11
2100	24.0	55.2	94.1	19.2	46.8	81.8	–4.8	–8.4	–12.3	–20	–15	–13

Table 4: Summary of sea-level rise obtained from the IS92a, c and e scenarios and those modified according to the Kyoto Protocol (Table 2)

3. Regional climate change scenarios

3.1 Present climate

The South Pacific island nations considered in this study roughly cover the area between 30°S and 20°N, and from 120°E to 120°W. The large-scale atmospheric circulation features in this region include the subtropical high pressure belt, the easterly trade winds, the Intertropical Convergence Zone (ITCZ), the South Pacific Convergence Zone (SPCZ), the monsoon trough, and the Hadley and Walker circulations (Bjerknes, 1966, 1969; Mullan, 1992). The year-to-year variability in the climate of this region is largely modulated by the El Niño–Southern Oscillation (ENSO) phenomenon (Fleer, 1981; Rasmusson and Carpenter, 1982; Behrend, 1984; Ropelewski and Halpert, 1987; 1996; Hay et al., 1993; He et al., 1998). Variations in sea surface temperatures (SSTs) across the equatorial Pacific have been linked to changes in the east–west or Walker Circulation (Bjerknes, 1969). Major circulation features described above are shown in Figures 7a and b for January and July. Rainfall in the South Pacific region indicates a complex pattern of seasonal variation that includes single and double peaks and also uniform distribution.

ENSO plays a major role in determining the interannual variability of rainfall in South Pacific islands. Previous studies demonstrate a close connection between SSTs in the Pacific, the Walker Circulation and rainfall in this region (Bjerknes, 1969; Rasmusson and Carpenter, 1982; He et al., 1998). During normal conditions, the ascending branch of the Walker Circulation is located over the western Pacific where warmer SSTs exist. The descending branch of the Walker Circulation is located over the subtropical central/eastern Pacific where cooler SSTs are found. This pattern of the east–west circulation is strengthened, in terms of anomalies, during La Niña events when greater warming is found over the western Pacific and cooling in the eastern Pacific. More rainfall is received due to strong convective activity over the western Pacific. In contrast, during El Niño years, the Walker Circulation is weakened and the active convective region moves over the central Pacific, near the dateline. During El Niño events, more rainfall is received in the central and eastern Pacific regions.

The ENSO phenomenon also affects the location of the SPCZ (Vincent, 1994). The SPCZ shifts eastward during El Niño events and moves westward during La Niña events. This is a band of cloud and rain that generally extends from Papua New Guinea south-eastwards towards South America. It is a preferred location for the genesis of tropical cyclones (Hastings, 1990, Basher and Zheng, 1995). Greater numbers of tropical cyclones form over the western Pacific during La Niña events, whereas during El Niño events, more tropical cyclones form over the central Pacific.

Apart from these phenomena, an intraseasonal or within-season oscillation with a time scale of 30 to 50 days also strongly influences the rainfall variability of this region. The 30 to 50 day oscillation is phased-locked to the seasonal cycle and has an eastward propagating characteristic and modulates the intensity of convective systems in the Australian and equatorial Pacific regions (Mullan, 1992; Suppiah, 1993).

Trends in temperature and rainfall in the South-West Pacific have been described in recent studies (Jones et al., 1986; Salinger et al., 1995; Folland and Salinger, 1995). These studies demonstrate relatively coherent climatic variations within the South-West Pacific. An increase in temperature has been recorded during the last century. However, there are some regional differences. A temperature increase trend began about 1950 south-west of the SPCZ and in New Zealand, and a systematic change in temperature and rainfall trends throughout the entire South-West Pacific occurred around the mid-1970s (Salinger et al., 1995) with an eastward displacement of the SPCZ by about 150 km. Salinger et al. (1995) also confirm that the strong interannual variability in rainfall in this region is controlled by ENSO.

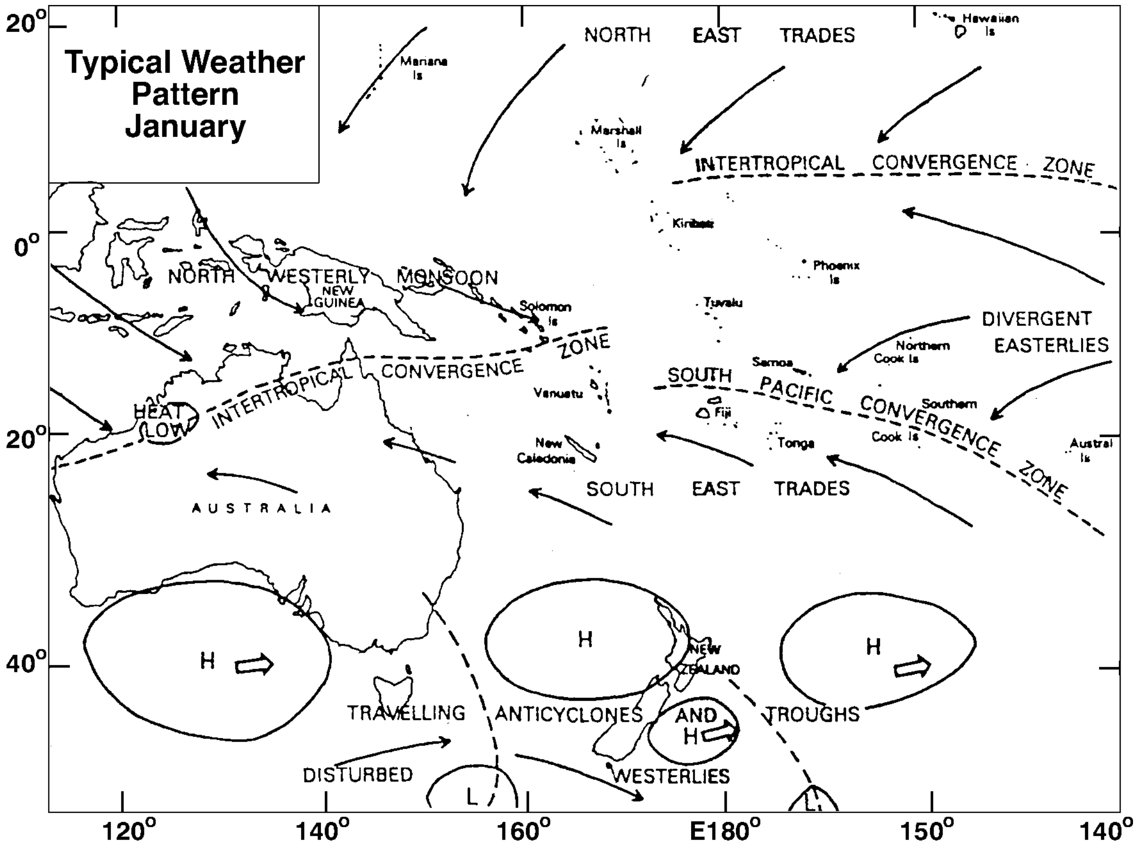


Figure 7a: Typical circulation features in the South-West Pacific during the summer

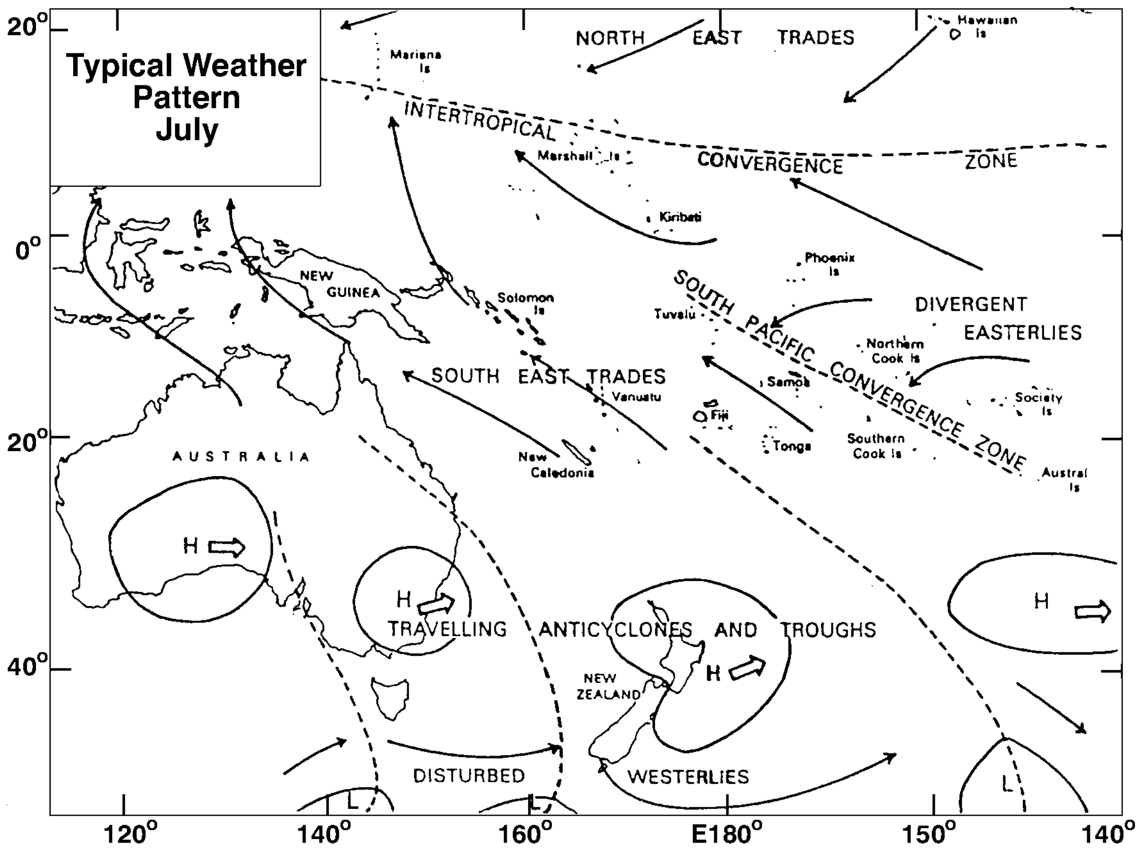


Figure 7b: Typical circulation features in the South-West Pacific during the winter (after Steiner, 1980)

3.2 Climate change modelling

Estimates of the future climate come from the representation of the Earth's climatic system through general circulation, or global climate models (GCMs). These models are as physically representative as possible within the limitations of scientific knowledge, the ability to represent physically phenomena on an appropriate scale, and computer capacity. They link the atmosphere, ocean and biosphere both vertically and horizontally. The models are divided into three-dimensional grid boxes, the scale and number of those boxes being limited by the computer power available to carry out the necessary computations.

The current generation of GCM is the *coupled*, or ocean-atmosphere GCM which contains three-dimensional representations of both the ocean and atmosphere. Since the ocean has a much larger thermal inertia and slower circulation relative to the atmosphere, climate equilibrium between the ocean and atmosphere (where the net energy flux between them is zero) is not reached for many hundreds of years. In these experiments, the enhanced greenhouse effect is simulated by gradually increasing the atmospheric loading of greenhouse gases in a so-called transient run. Where higher resolution over a region is desired, a regional climate model (RCM) can be nested within a GCM, being forced by boundary conditions taken from that GCM. Older 'slab' or mixed-layer ocean GCMs containing a grossly simplified ocean are not capable of simulating gradually changing conditions, so have largely been superseded.

CSIRO Atmospheric Research currently simulates climate change using the CSIRO Mark 2 coupled GCM and the DARLAM RCM. Three experiments are analysed here. In the two GCM simulations, the CSIRO coupled-ocean-atmosphere model has been integrated for 140 years from 1961 to 2100 with a gradually increasing CO_2 concentration equivalent to the forcing produced by all greenhouse gases in the IS92a emission scenario. One simulation has the direct affects of atmospheric sulphate aerosol added (i.e. the indirect effects, or atmospheric feedbacks, are omitted) which provides a negative forcing relative to the greenhouse-gas-only run. The regional climate model, DARLAM, with a higher resolution of 125 km, was nested in the greenhouse-gas-only simulation in order to provide higher resolution data. The model domain extends well east of Australia into the Pacific Ocean (Figure 8), covering most of the SPREP region.

Results from three GCMs from other modelling groups have also been analysed to provide data for this project. All models used in scenario development are summarised in Table 5. In the analyses presented in Sections 3.3 and 3.4.1, the

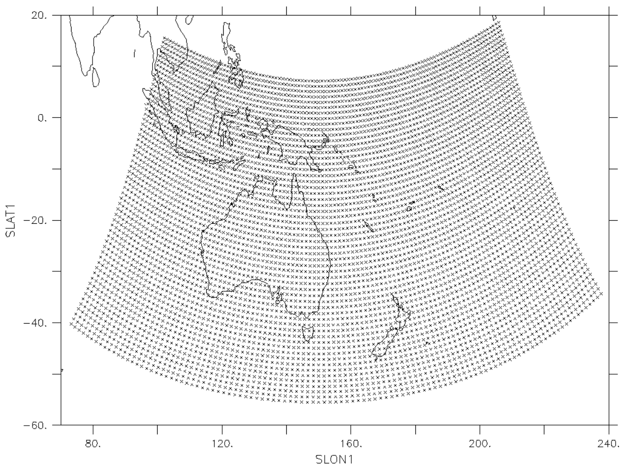


Figure 8: Domain of DARLAM simulation nested with transient GCM simulation—crosses denote grid points

Centre	Model	Emission Scenario	Features	Years
CSIRO, Australia[1]	Mk2	IS92a equivalent CO_2	No sulphates, GM ocean	1881– 2100
CSIRO, Australia	Mk2 with sulphates	IS92a equivalent CO_2	Sulphates, GM ocean	1881– 2100
CSIRO, Australia[2]	DARLAM 125 km	IS92a equivalent CO_2	Nested in CSIRO Mk2	1961– 2100
DKRZ, Germany[3]	ECHAM4/OPYC3	IS92a	No sulphates	1860– 2099
Hadley Centre, UK[4]	HADCM2	1% CO_2 pa	No sulphates	1861– 2100
Canadian CCMA[5]	CGCM1	1% CO_2 pa	No sulphates	1900– 2100

1 Gordon and O'Farrell (1996)
2 McGregor and Katzfey (1998)
3 DKRZ-Model User Support Group (1992), Oberhuber (1992)
4 Cullen (1993)
5 Flato et al. (submitted)

Table 5: Model runs used to produce the regional scenarios presented in this report

control climate is considered as the first 30 years from 1961 to 1990 with transient climate change being analysed over the whole simulation (the method is described in the next section).

A number of analyses of climate were carried out with the CSIRO Mark 2 GCM (no sulphates) and DARLAM. In Sections 3.4.2 and 3.4.3, the first and last 40 years of the 140 year run were analysed. Other important milestones in simulations of IS92a forced runs is that $2\times CO_2$ occurs in about 2035 and $3\times CO_2$ in about 2075.

3.2.1 Scenario generation methods

GCMs are the best tools available to simulate present and future climate change, but they do not give firm answers to some important questions. For example, although many large-, synoptic- and local-scale circulation features dominate the climates of the South Pacific islands at various time scales, some of these are inadequately simulated by present GCMs. This makes it difficult to assess possible future climate changes. However, a number of conclusions can be made with varying degrees of confidence, ranging from projections made with reasonable confidence, to scenarios that are plausible but which have little confidence attached.

The scenarios presented here have also been prepared for the Pacific Island Climate Change Assistance Programme (PICCAP) for incorporation into the scenario generator and impacts package PACCLIM, developed by the International Global Change Institute, University of Waikato. PACCLIM contains three regions: Micronesia, Melanesia and Polynesia (Figure 9). In the analyses presented in this report, Polynesia has been divided into northern and southern regions due to differing responses under climate change. These scenarios build on two earlier regional scenarios developed for the South Pacific by Pittock (1992) and Pittock et al. (1995). In this project, several new scenario construction techniques are introduced.

The scenarios prepared for the PICCAP Project are scaled scenarios which convert GCM output into local change per degree of global warming. For temperature, this local change is measured in degrees Celsius per degree of global warming. Rainfall is presented as a percentage change per degree of global warming. The technique most widely used to calculate these changes has been to average a period of simulation (usually 30 years at $2\times CO_2$) to create a pattern of global warming for each month of the year. These averages are subtracted from those of a similar length period of control climate for each grid box. This figure is divided by the average global warming for the model during the simulation period under climate change. This procedure eliminates the individual climate sensitivities of different GCMs, allowing the comparison of regional patterns from those models.

This method dates from the period when slab-ocean GCMs only permitted climate change to be simulated through an instantaneous change in greenhouse gases. The coupled ocean–atmosphere GCM allows the much more realistic application of historical and projected greenhouse gases over time to force the model in a so-called 'transient' simulation. This produces a gradually increasing global temperature over time, the magnitude of which is influenced largely by the climate sensitivity of the model and the emission scenario used (usually IS92a or 1 per cent CO_2 per annum; Table 5). Patterns of change under these simulations can be calculated by regressing local climate against global warming on a monthly basis for each grid box (Hennessy et al., 1998).

The major advantage gained by using this method is the elimination of transitory and non-linear influences such as multi-decadal variability. This was particularly problematic for short runs, as it was difficult to ascertain to what degree a pattern was influenced by global warming or by multi-decadal variability. The pattern that is produced is similar to that of earlier methods—local change per degree of global warming—but it represents the

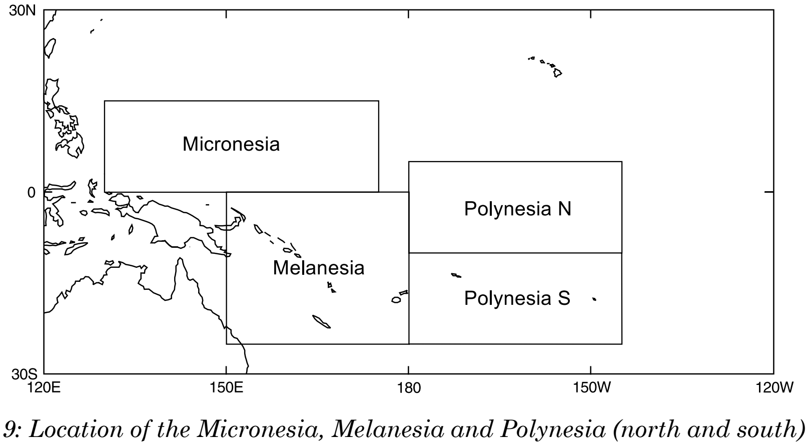


Figure 9: Location of the Micronesia, Melanesia and Polynesia (north and south) regions for which scenarios have been prepared.

global warming signal over the entire period of simulation. Techniques are being developed to assess the influence of multi-decadal variability separately to the global warming signal, but time and resources do not allow those techniques to be presented here.

3.3 Regional mean surface and ocean temperature

Resolution is a major problem when estimating changes in temperature from climate models, particularly for small landmasses in oceanic regions. Papua New Guinea is the only SPREP nation to be represented as a landmass in the GCMs analysed. New Zealand, south of the SPREP region, is represented by one to four grid boxes, depending on the resolution of the particular GCM. Even DARLAM, at a resolution of 125 km, in addition to Papua New Guinea only has one grid point representing the Solomon Islands. Therefore, temperature from the climate models analysed in this study represents marine air temperature across the SPREP region.

In addition to large-scale climatological factors, the relationship between an island and its climatology depends on the island's size, its topography and vegetation. While most SPREP nations have a maritime climate, with parts of Papua New Guinea being the exception (Campbell, 1996), on larger islands temperatures will demonstrate a greater variability than marine air temperature. These locations generally have a larger diurnal temperature cycle, and may heat faster than the surrounding ocean under climate change, though this additional rate of warming compared to that of the surrounding ocean will be small (<0.1°C per degree of global warming).

In general, all climate models reproduce observed temperatures reasonably well on an annual basis, with the DKRZ model perhaps showing a slight warm bias (Figure 10). The CSIRO GCM with sulphates is not shown in Figures 10, 11 and 13–16 due to its similarity to the non-sulphate simulation. Temperature changes in the SPREP region range from 0.6°C per degree of global warming to >1.0°C, although they are generally below the global average (Figure 11). Areas of greatest warming tend to be associated with the Australasian land mass (including Papua New Guinea) and the north-east of the region, with the least warming to the south-east. For the PICCAP regions, the models produce a variable response in Micronesia (0.7–1.0°C per degree of global warming), Melanesia is slightly cooler (0.7–0.9°C per degree of global warming), north Polynesia is the region of greatest warming (0.8–1.0°C per degree of global warming) and south Polynesia the region of least warming (0.7°C per degree of global warming).

Most coupled GCMs show slower warming in the high latitudes of the southern ocean than elsewhere (Whetton et al., 1996). Originally, it was thought this may have been due to limitations within the ocean-mixing schemes in coupled GCMs but this result has remained consistent as the dynamics of ocean modelling have improved (Hirst et al., 1996). The rate of warming over the ocean has implications for both rainfall change (next section) and sea-level rise (Section 3.5).

3.4 Regional precipitation patterns and intensity

Firstly, this section offers a brief description of GCM simulations of rainfall for the periods November to April and May to October concentrating on the CSIRO GCM. The section then describes changes under enhanced greenhouse conditions based on patterns of response to the greenhouse effect. Changes in SST and rainfall in the standard western Pacific, NINO4 and NINO3 regions are discussed in relation to ENSO under enhanced greenhouse conditions. These areas include western Pacific (5°N–5°S, 120°E–160°E), NINO4 (5°N–5°S, 160°E–150°W) and NINO3 (5°N–5°S, 150°W–90°W) regions (Figure 12) and are used to understand changes in ENSO behaviour. Changes in the frequency of monthly rainfall are also discussed, comparing the first and last 40 years of data from the CSIRO GCM to get a broad picture of changes in rainfall variability and intensity.

3.4.1 Mean seasonal rainfall changes

Figures 13 and 14 show observed and coupled GCM simulated rainfall for the November to April and May to October periods. Here the simulated mean rainfall results of 1961–1990 are compared with the long-term average observed rainfall climatology from Legates and Willmott (1990). The CSIRO, Hadley Centre and DKRZ GCMs broadly reproduce the large-scale patterns of rainfall for both periods. Most importantly, the rainfall maxima of the SPCZ and ITCZ over tropical Pacific are captured, with the CSIRO, Hadley Centre and DKRZ GCMs also producing the rainfall minima between the SPCZ and ITCZ. The CSIRO DARLAM and CCCMC models produce a large maximum in the broad area of the SPCZ, however, the ITCZ is poorly developed. Within CSIRO DARLAM, this is due to two reasons:

- edge effects due to the boundary of the model with the host GCM truncating important features, including the ITCZ; and

- internal variations within the large oceanic domain that propagate independently of those same features within the host GCM. This is not a feature of simulations over the Australian continent and may be due to the ocean's surface providing inadequate buffering over such a large domain.

For these reasons, patterns of rainfall change from the CSIRO GCM are preferred, although DARLAM has been retained for some further analyses.

Although the spatial patterns are well simulated, the CSIRO model underestimates the amount of observed rainfall in these zones. The Hadley Centre and DKRZ GCMs produced maxima of the right magnitude, however, they are too broad in extent. In addition to being limited by poor rainfall distributions, CSIRO DARLAM and the CCCMC model are also too moist.

Figures 15 and 16 show the simulated percentage change in the May to October and November to April half-yearly rainfall under transient conditions. This transient period includes 2× and $3\times CO_2$ concentrations in the atmosphere. All models simulate rainfall increases in the eastern Pacific but they are not spatially consistent. The DKRZ GCM shows an increase in the intensity of the ITCZ across the central and eastern Pacific of up to 50 per cent per degree of global warming, with a reduction to the south of up to 25 per cent, showing similar changes in both half years. The SPCZ moves south-west with some intensification in the western Pacific of up to 25 per cent per degree of global warming.

The Hadley Centre GCM shows a north-east movement and intensification of rainfall in both the SPCZ and ITCZ in the May to October half year, with drying in the far western Pacific. The November to April half year is less clear, but appears to show a similar movement. Increases in rainfall in the central and eastern Pacific are 45 to 50 per cent for both seasons in a band surrounding the equator.

The CSIRO GCM produces less intense changes over the equatorial Pacific. There is a southward movement in the ITCZ, whereas the SPCZ remains fairly stationary. Slight increases occur over the north-western Pacific, especially in May to October, whereas there are slight decreases in the southern part of the western Pacific. CSIRO DARLAM is broadly similar. The CCCMC shows a north-easterly shift in the SPCZ, with decreases in the western Pacific.

Although the models are not consistent in the way that they simulate changes to the tropical convergence zones, they do consistently produce rainfall increases where surface temperature increases are highest, and vice versa. On the basis of this evidence, the areas of greatest warming across the Pacific are likely to be associated with the largest rainfall increases.

In the PICCAP regions, the models produce small positive to negative changes on average in Melanesia and south Polynesia in the May to October half year. For Micronesia these changes range from 0 to 10 per cent per degree of global warming, and for north Polynesia, range from 4 to 43 per cent per degree of global warming. This is the region of largest variation of possible rainfall change.

In the November to April half year, small positive to negative changes are again simulated in Micronesia, Melanesia and south Polynesia. For north Polynesia, simulated rainfall increases range from 6 to 20 per cent per degree of global warming, so this region experiences rainfall increases in both half years, some of which are substantial.

3.4.2 Year-to-year variations and trends in simulated rainfall

Year-to-year variations, or interannual variability, dominate many impacts where rainfall is a key climate variable, particularly agriculture and water supply. Interannual variability is limited in climate models due to their limited ability to simulate phenomena such as ENSO (which is associated with large variations in annual rainfall). The ability of GCMs to simulate ENSO is discussed further in Section 3.6.

The CSIRO GCM was analysed over the western Pacific, NINO4 and NINO3 regions (Figure 12). Figures 17 (a), (b) and (c) show year-to-year variations and trends in simulated rainfall in those regions. Both total rainfall and year-to-year variations in seasonal rainfall in mm pa decrease from west to east, being large in the western Pacific, moderate in the NINO4 region, and much smaller in the NINO3 region. Rainfall simulated by the CSIRO GCM shows an increasing trend during the 140-year period right across the tropical Pacific. The increasing trend is large in NINO4 and NINO3 while it is small in the western Pacific.

The coefficient of variability in Figure 17 (the standard deviation as a percentage of the average) ranges from about 10 to 15 per cent in all three regions, with no significant change over the simulation. Observed values are much higher in the equatorial Pacific, ranging from 20 to 60 per cent (Musk, 1976). This may be partly due to the large grid size of the model but is also due to the lack of fully developed ENSO variations.

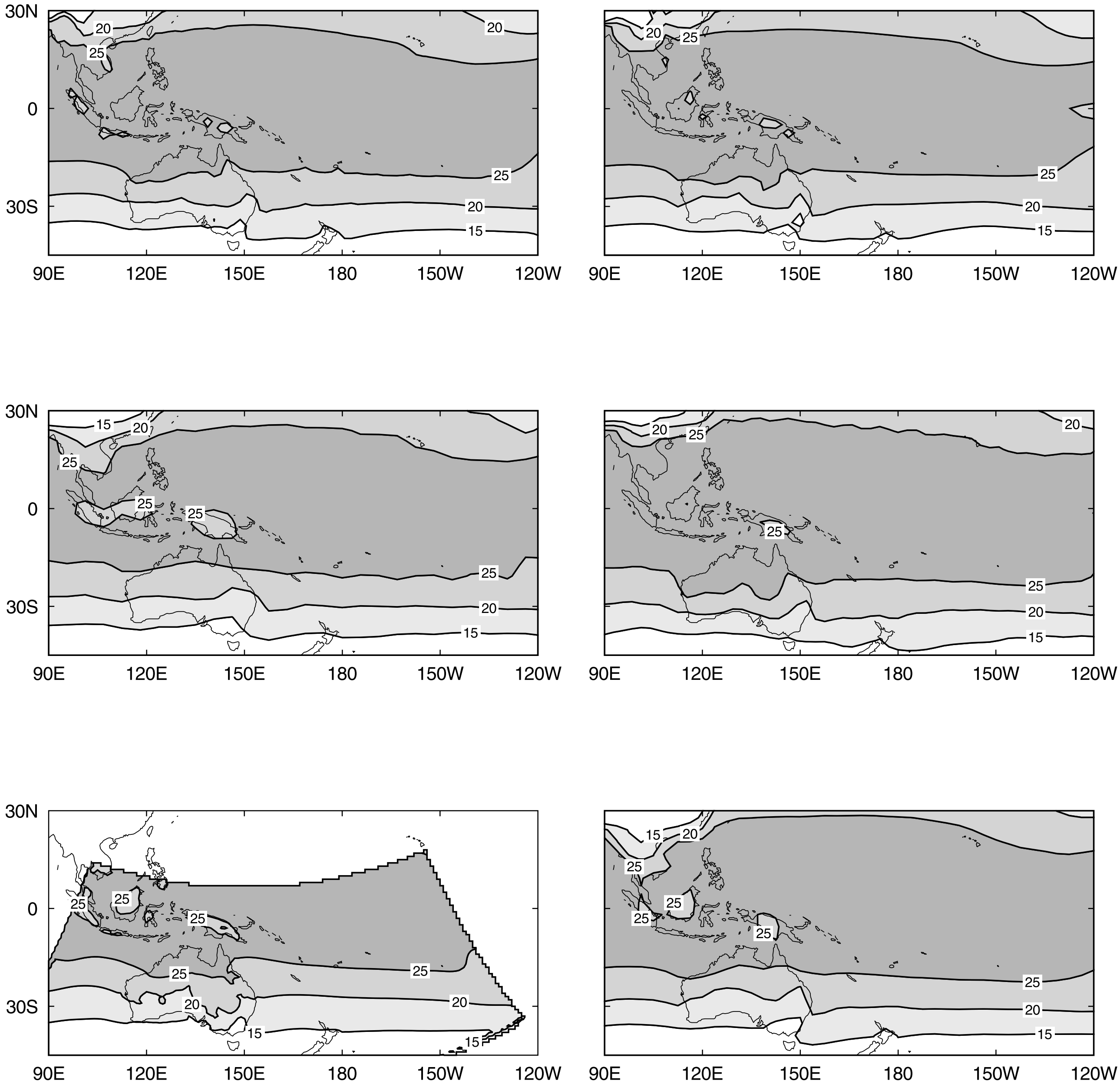


Figure 10: Observed temperature and simulated annual average temperature from five climate models centred over the western Pacific for the period 1961–1990

Model	Micronesia	Melanesia	Polynesia N	Polynesia S
CSIRO	0.8	0.8	0.8	0.7
CSIRO sulphates	0.8	0.8	0.9	0.7
CSIRO DARLAM	0.7	0.7	0.8	0.7
DKRZ	0.8	0.8	0.8	0.7
Hadley Centre	1.0	0.9	1.0	0.7
CCCMA	0.8	0.8	1.0	0.7

Table 6: Temperature change per degree of global warming from the models shown in Table 5 (°C)

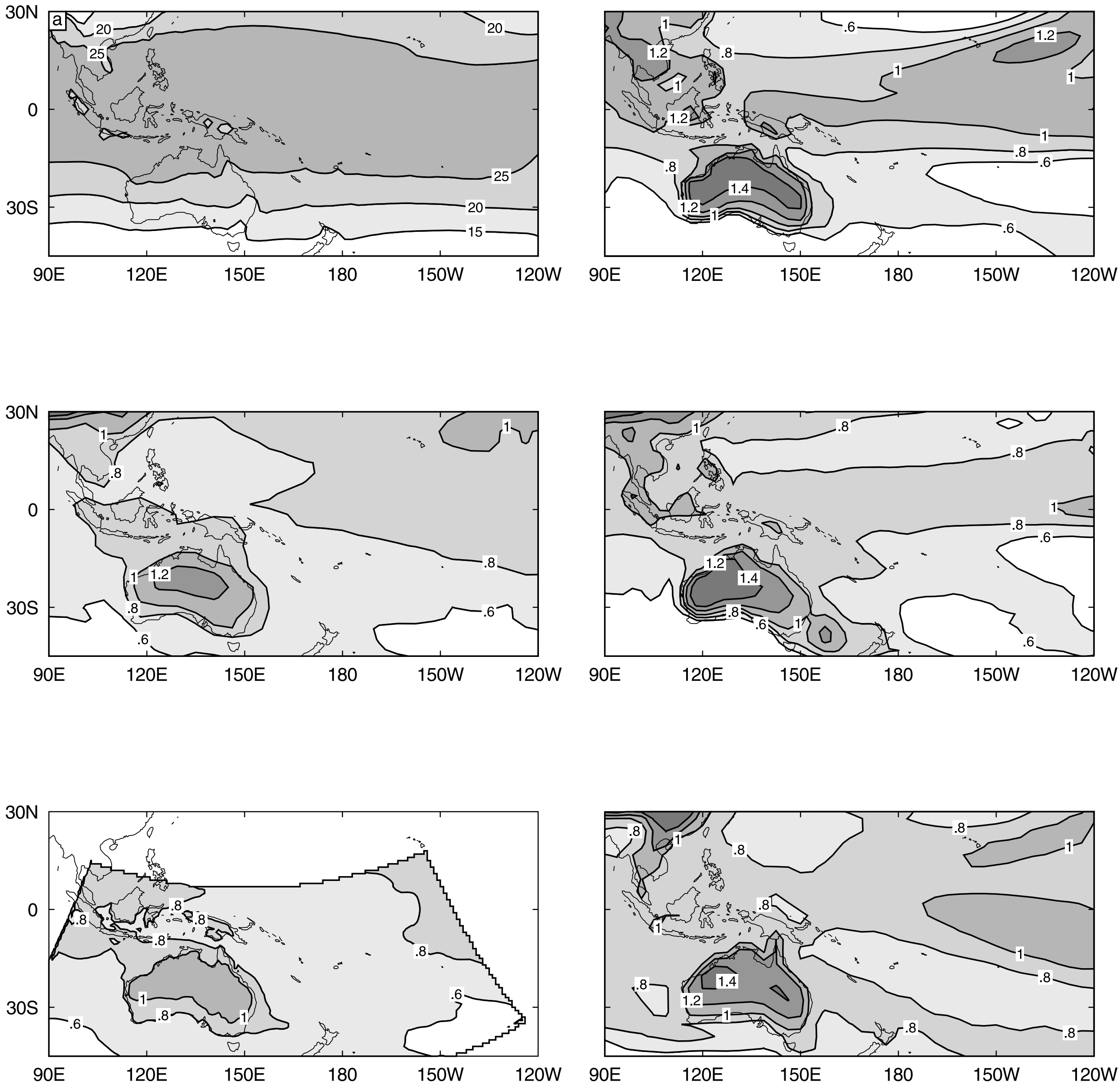


Figure 11: Observed temperature and simulated change of annual average temperature obtained by regression of local change on global average change, shown in °C per degree of global warming, from five climate models centred over the western Pacific, from the commencement of the model run to 2100 (see Table 5)

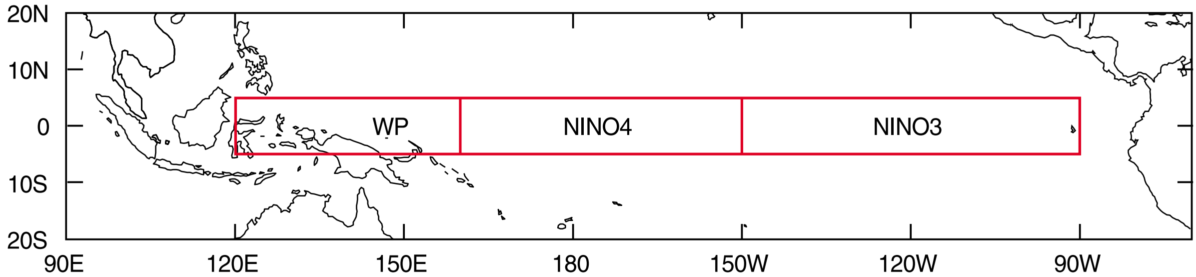


Figure 12: Map of regions used for averaging rainfall.
Regions from west to east are western Pacific, NINO4 and NINO3.

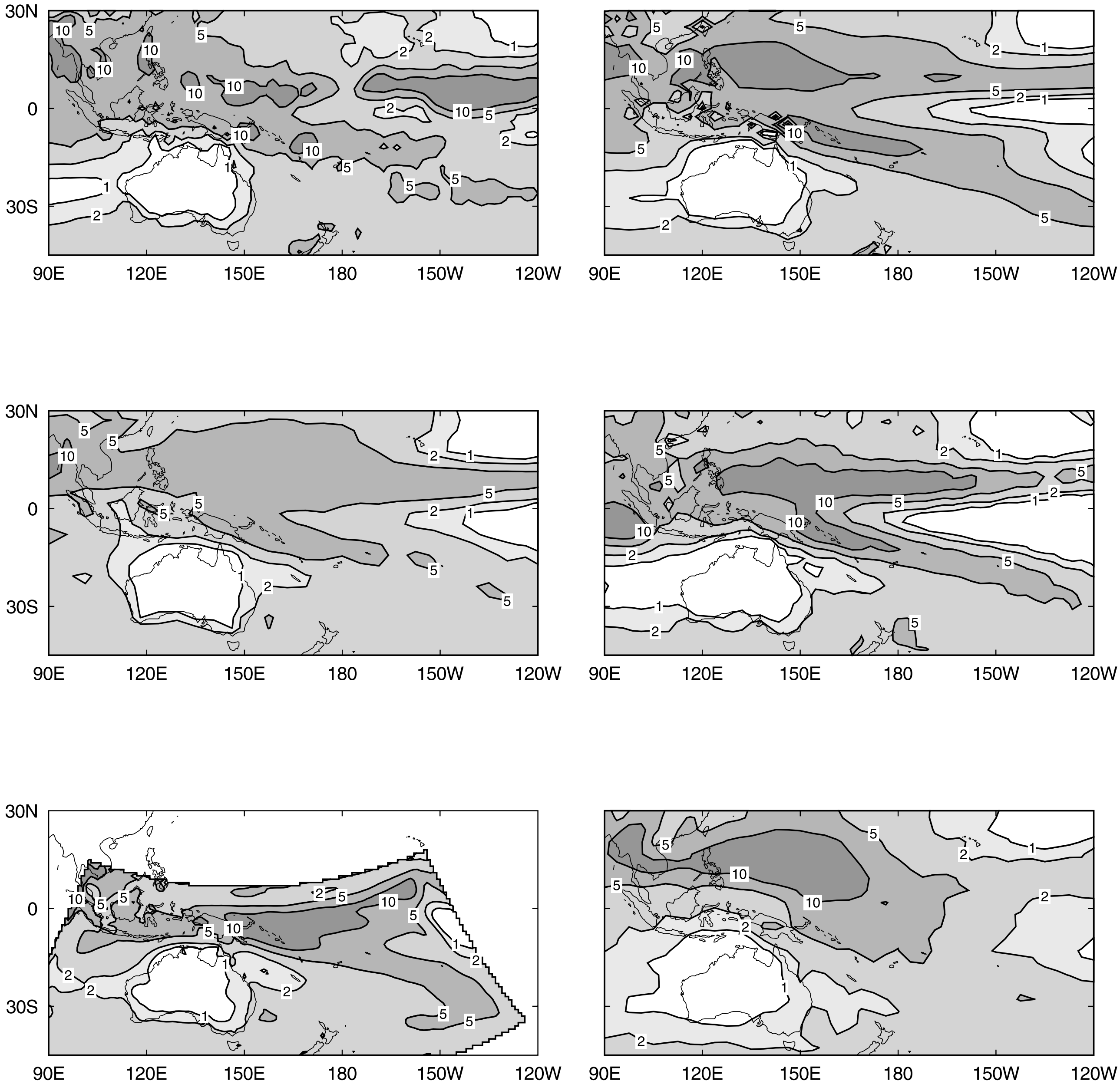


Figure 13: Observed rainfall (Legates and Willmott, 1990) and simulated total rainfall for the period May to October from five climate models centred over the western Pacific for the period 1961–1990

Model	Micronesia	Melanesia	Polynesia N	Polynesia S
CSIRO	6	2	4	1
CSIRO sulphates	6	1	5	2
CSIRO DARLAM	10	2	12	2
DKRZ	6	6	27	–8
Hadley Centre	5	–1	43	2
CCCMA	0	–4	13	0

Table 7: May to October rainfall change per degree of global warming from the models shown in Table 5 (%)

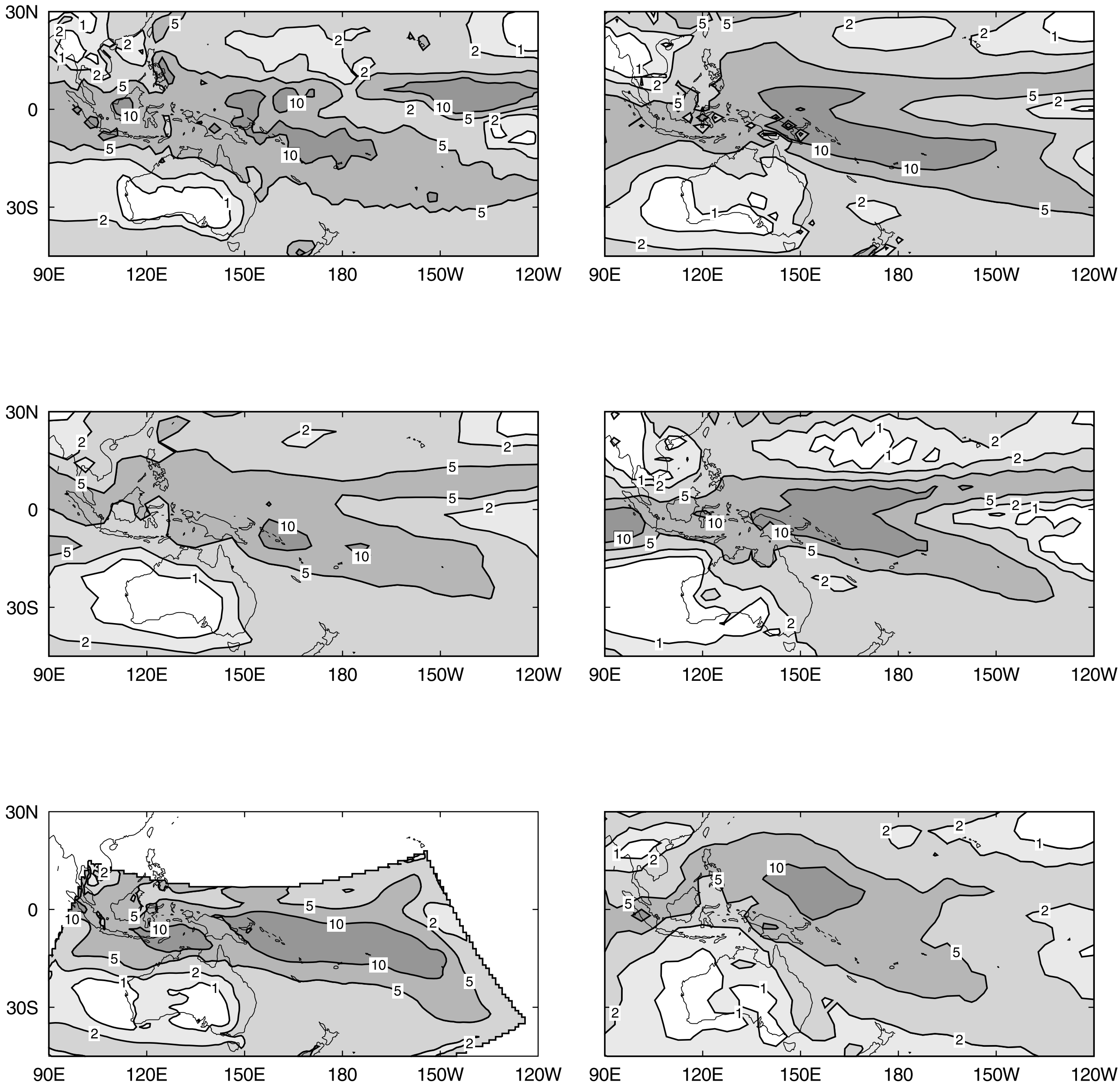


Figure 14: Observed rainfall (Legates and Willmott, 1990) and simulated total rainfall for the period November to April from five climate models centred over the western Pacific for the period 1961–1990

Model	Micronesia	Melanesia	Polynesia N	Polynesia S
CSIRO	1	1	6	1
CSIRO sulphates	1	1	7	1
CSIRO DARLAM	2	3	14	3
DKRZ	0	6	18	–6
Hadley Centre	4	2	20	–1
CCCMA	–2	–3	14	3

Table 8: November to April rainfall change per degree of global warming from the models shown in Table 5 (%)

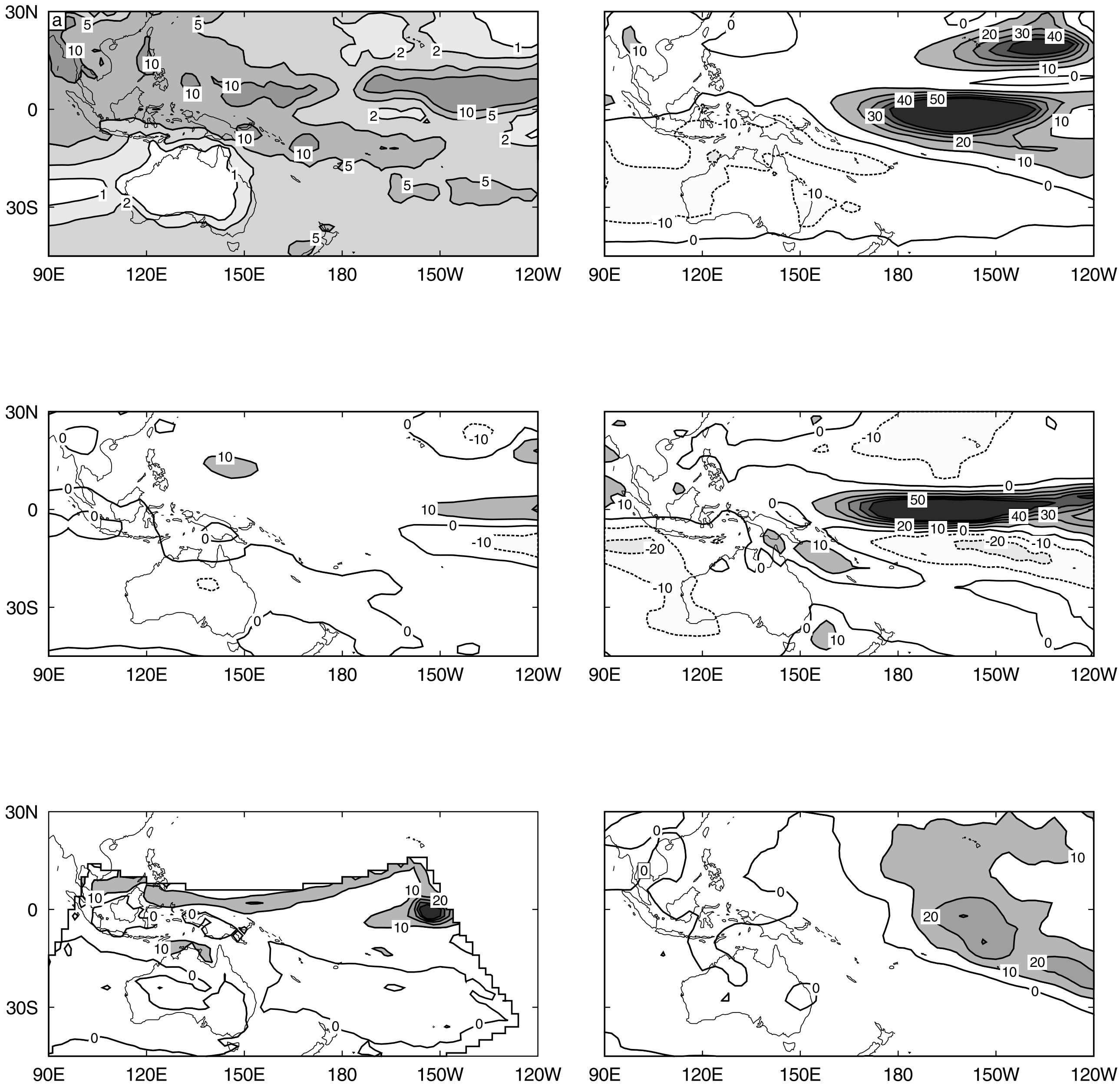


Figure 15: Observed rainfall (Legates and Willmott, 1996) and simulated change of total rainfall for the period May to October in per cent change per degree of global warming from five climate models centred over the western Pacific, from the commencement of the model run to 2100 (see Table 5)

	Micronesia	Melanesia	Polynesia N	Polynesia S
Dec. – Feb.	– 31	– 16	– 129	**–3.1**
June – Aug.	**+7.5**	– 99	– 28	**–3.5**

Table 9: December–February and June – August rainfall change per degree of global warming for the Hadley Centre UKHI model used by Hennessy et al. (1997). All figures are percentage of total rainfall. Numbers in bold represent regions shown in Figure 19.

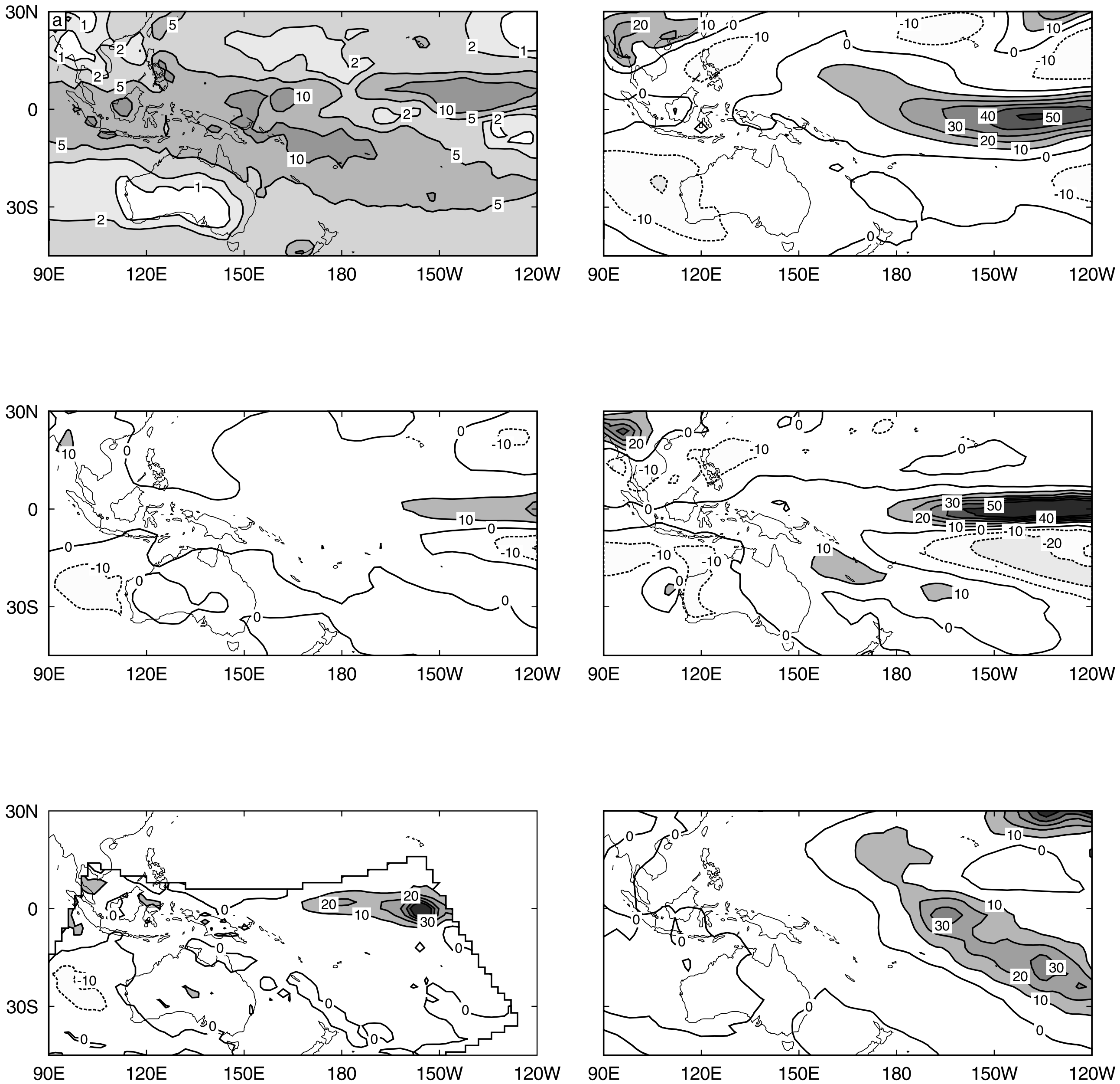


Figure 16: Observed rainfall (Legates and Willmott, 1996) and simulated change of total rainfall for the period November to April in per cent change per degree of global warming from five climate models centred over the western Pacific, from the commencement of the model run to 2100 (see Table 5)

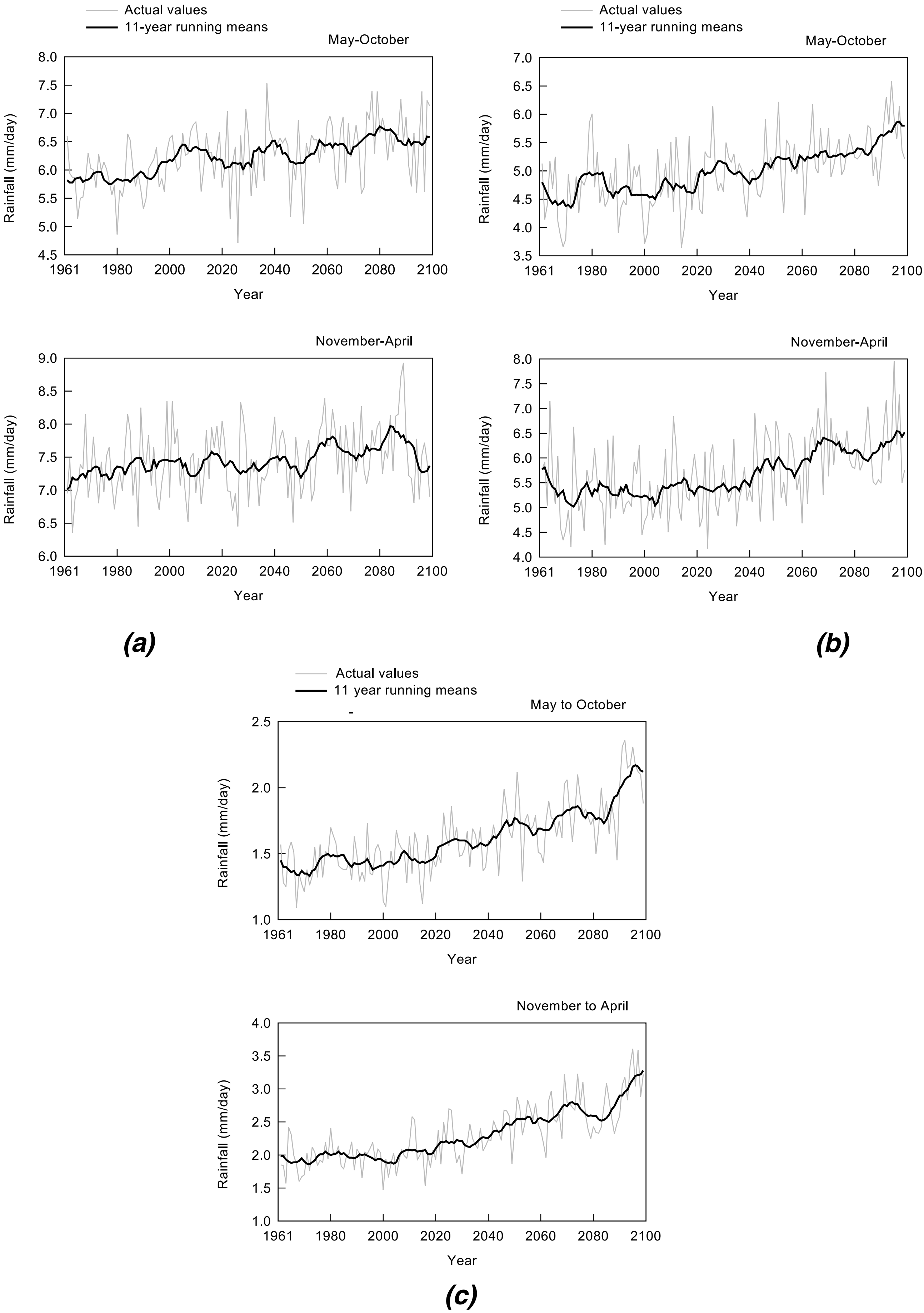


Figure 17: Simulated year-to-year variability and trends in summer and winter half-year rainfall in the CSIRO GCM over (a) western Pacific, (b) NINO4 and (c) NINO3 regions. Actual values are shown by thin lines and smoothed values by a 11-year running means are shown by thick lines.

3.4.3 Decadal variations and climate change

Decadal variations, or multi-decadal variability are slow variations of climate on time scales of ten years or more, recently reviewed by Latif (1998). Decadal variations are important for several reasons. They are slow variations and can therefore mimic or obscure the slow global warming signal. For the next few decades, they will possibly be larger in amplitude than the global warming signal (Hulme et al., in press). They can also interfere with the ENSO variations, changing relationships between such measures of ENSO as the SOI and variables like rainfall (e.g. Nicholls et al., 1996), and possibly with relationships between ENSO and tropical cyclones as well.

Relationships between decadal variations and a number of different phenomena have been demonstrated, ranging from Australian rainfall to Atlantic tropical cyclones (e.g. Walsh and Kleeman, 1997). It is quite possible that relationships between decadal variations and tropical cyclones in the SPREP region await discovery. One limitation is the lack of a genuinely reliable long-term tropical cyclone record. In the era before meteorological satellites, which began in the late 1960s, many cyclones in the SPREP region were missed because they occurred over the open ocean far from land. This relatively short record hinders the study of longer term variations such as decadal variability. These variations need to be studied further using both observations and climate model simulations to determine how they change with climate change.

Climate models do produce well-defined decadal variations in rainfall as shown in Figure 17 which can only be captured in long GCM runs. However, how they may affect climate change in the future has not been studied, save for an analysis from Hulme et al. (in press). Given that the technique presented in Section 3.2.1 of producing the global warming signal on a local basis removes transitory influences such as decadal variations, it is important to develop scenario-building techniques that can explicitly represent them.

3.4.4 Rainfall intensity

Frequency distributions of monthly rainfall amounts comparing the first and last 40-year periods for the three regions in Figure 12 are shown in Figures 18 (a), (b) and (c). These figures indicate a shift towards heavier rainfall classes during both half-year periods. In particular, the shift is large in November to April in the NINO3 region which receives less total rainfall compared with the NINO4 and western Pacific regions. The NINO4 region is coincident with the north Polynesia region in Figure 9, so Figure 18 (b) reflects the large increases in rainfall noted in Tables 7 and 8 (The local change per degree figures of 4 and 6 per cent for the CSIRO GCM would lead to total changes exceeding 12 per cent with the warming encountered during this 40-year period).

A number of modelling studies show that greenhouse warming may lead to increased extreme daily rainfall intensity and frequency (Gordon et al., 1992; Houghton et al., 1996; Hennessy et al., 1997). For a doubling of present CO_2 levels in earlier slab-model versions of the CSIRO and Hadley Centre GCMs, extreme rainfall events (which recur on average once every one to ten years) in Europe, USA, Australia and India increase in magnitude by 10 to 25 per cent (Hennessy et al., 1997). The return period of extreme rainfall increases by a factor of 2 to 5.

Since daily rainfall data were not available for the models listed in Table 5, an analysis of daily data from the earlier Hadley Centre model (UKHI) was performed over Micronesia, Melanesia, north Polynesia and south Polynesia. Results for mean rainfall change (per cent) per degree of global warming are shown in Table 9 for December–February and June–August. The best correspondence between these results and those in Tables 7 and 8 are for south Polynesia in December–February and June–August, and Micronesia in June–August. Extreme daily rainfall intensities and return periods for these regions are shown in Figure 19.

Figure 19 also shows selected rainfall return periods for the Micronesia and south Polynesia regions. Over Micronesia in June–August, the model simulates a 1-in-10-year rainfall event of about 100 mm/day for the present climate. Under $2\times CO_2$ conditions, equivalent to a warming of 3.5°C in this model, the 1-in-10-year event becomes 115 mm/day. Similarly, the 1-in-5-year event increases from about 80 mm/day to 95 mm/day. This can also be interpreted as doubling of the frequency of 100 mm/day events (i.e. the 1-in-10-year event becomes a 1-in-5-year event). The 15–18 per cent increase in intensity is smaller than the 26 per cent increase in mean rainfall (7.5 per cent per degree of global warming). A smaller percentage increase in extreme rainfall intensity relative to the increase in mean rainfall is unusual for GCM simulations—usually percentage increases in extreme rainfall tend to exceed increases in the mean.

For south Polynesia in June–August and December–February, the model simulates little change (0–5 per cent) in the intensity and frequency of extreme rainfall despite 11–12 per cent decreases in mean rainfall (3 per cent per degree of global warming). Other model simulations have also shown this tendency for small decreases (or slight

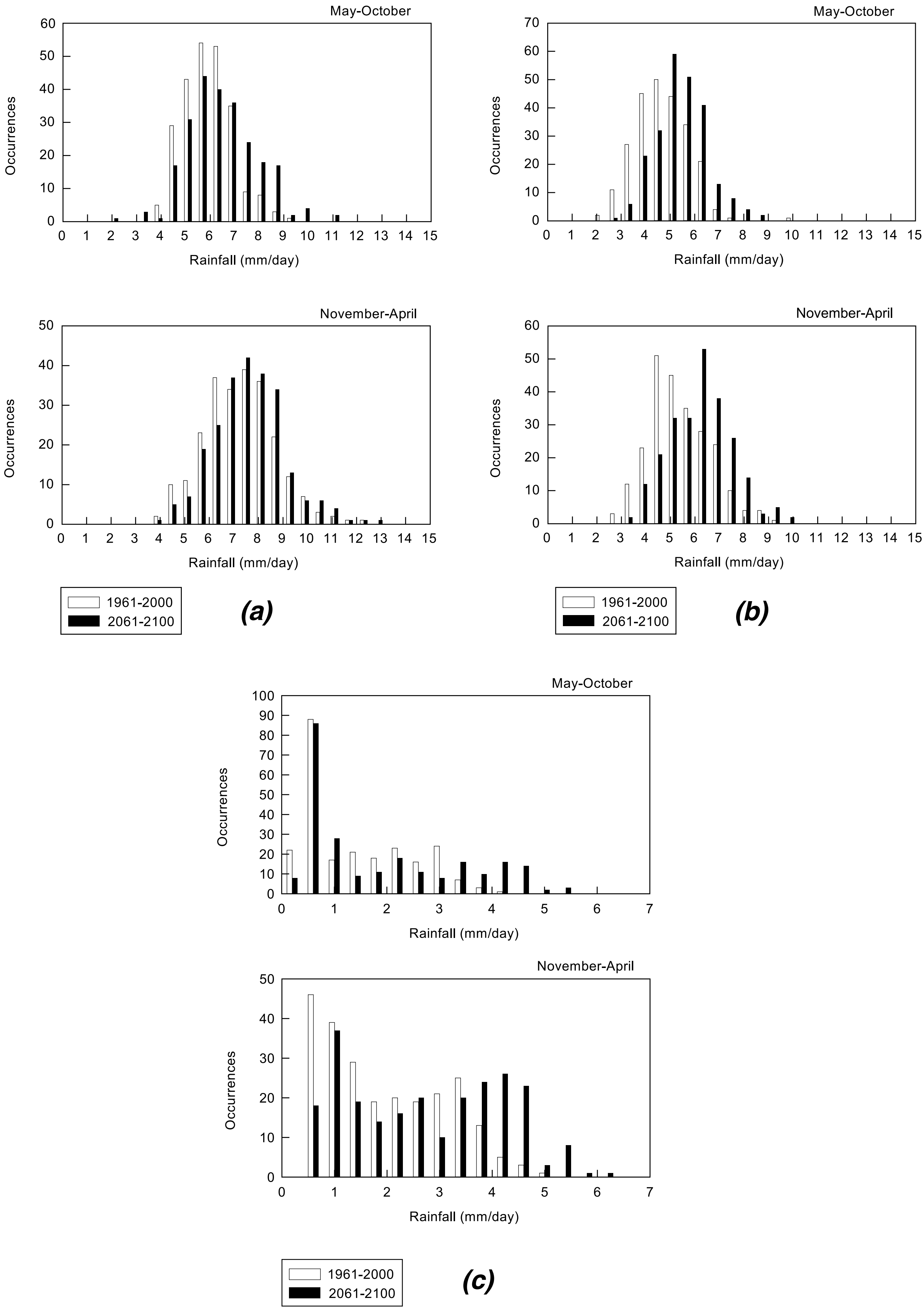


Figure 18: Simulated frequency distribution of monthly rainfall for summer and winter half years for (a) western Pacific, (b) NINO4 and (c) NINO3 regions.
Blank bars represent the first 40 years, while solid bars represent the last 40 years.

increases) in rainfall intensity associated with moderate decreases in the mean (Hennessy et al., 1998). These results demonstrate how rainfall intensity may change with climate change. Similar sets of analyses should be carried out with more recent model simulations, and also to determine the nature of the relationship between changes in mean rainfall and rainfall intensity.

3.5 Sea-level rise trends

Global average sea-level rise was discussed in detail in Part One of this publication. Historical global average sea-level rise is estimated at between 10 and 25 cm over the past 100 years, while the possible rise due to the IS92a–f emissions scenarios, with allowances for uncertainties in climate sensitivities and ice-melt parameters, is 13–94 cm by 2100 (Warrick et al., 1996).

When impacts are an important part of sea-level rise considerations, several issues need to be raised. Firstly, relative sea level is most important for impacts; i.e. sea-level change at a site relative to the surrounding land surface. This incorporates changes in the elevation of the land surface due to post glacial rebound and tectonic effects. However, to measure sea-level change as a factor of climate change, land movements need to be subtracted from land-based measurements of sea level. At present, such movements are estimated through the modelling of post-glacial isostatic adjustments (e.g. Tushingham and Peltier, 1991; Peltier, 1994). Several other methods are being implemented: geodetic surveys linking tide gauges to deep benchmarks (Turner, 1997) and global positioning satellite methods (Woodworth, 1997) but results from these methods may not be forthcoming for up to two decades.

This section looks at historical sea-level rise and scenarios of sea-level rise under the enhanced greenhouse effect.

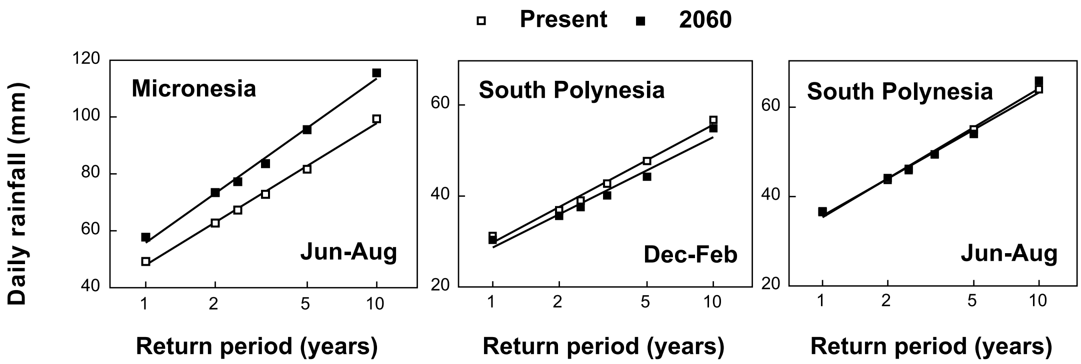


Figure 19: Extreme rainfall return periods averaged over Micronesia and south Polynesia from the superseded Hadley Centre model. Open squares are for present conditions and solid squares are for 2×CO_2 conditions. Results for other regions were not computed (see text).

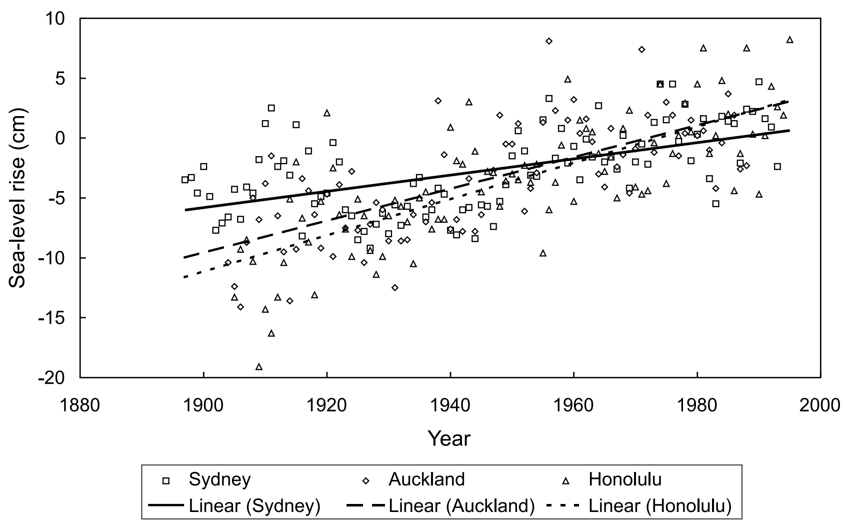


Figure 20: Sea-level rise trends for raw data from the Sydney, Auckland and Honolulu tide gauges relative to the 1961–1990 average at each of the stations

3.5.1 Historical sea-level rise in the Pacific

The estimated historical global sea-level rise from 1880 to 1990 is estimated be 1.8 mm pa with an uncertainty range of 1.0–2.5 mm pa (Warrick et al., 1996) or 10–25 cm over a century. This estimate was based on tide gauge records modified by estimates of vertical land movements (e.g. post-glacial isostatic rebound and tectonism). More recent satellite measurements estimate a rise of 1.8 mm pa since 1992 (Nerem et al., 1997) although this is a very limited period of measurement.

Three long-term tide gauges in the eastern and central Pacific have been in operation from early this century: Sydney, Auckland and Honolulu. The trends from the raw data are 0.7, 1.3 and 1.5 mm pa respectively (Figure 20). When post-glacial rebound calculated by Tushingham and Peltier (1991) is factored in, then these figures become 1.3 mm pa for Sydney and 1.9 mm pa for Honolulu. Douglas (1991) gives a figure of 1.6 mm pa for the Pacific Ocean, corrected to 2.0 mm pa when post-glacial rebound is added. All these estimates are well within the range of historical global sea-level rise from Warrick et al. (1996).

There is an extensive network of tide gauges across the Pacific but other than the three above, which are west, south and north of the SPREP area, only three within the SPREP region have recently reached 50 years' duration, the minimum period given as sufficient for estimating long-term trends (Douglas, 1991). The South Pacific Sea Level and Climate Monitoring Project maintains a network of high resolution monitoring stations in 11 island countries of the South Pacific with the aim of measuring the relative motions of land and sea at each station; carrying out a supplementary survey and geodetic programme to measure movements of the crust; helping identify changes to sea levels with reference to a similar network of stations in Australia and elsewhere in the world; and collaborating with ongoing international geodetic programmes. This project has been under way since 1992, too brief a period to determine long-term sea-level trends (National Tidal Facility, 1998).

Sea-level trends and relative movement of the land surface are shown in Table 10. Due to the brevity of the record, the sea-level trends are dominated by climate variability, such as the recent El Niños of 1994–1995 and 1997–1998, and other atmospheric, oceanographic and geological processes (National Tidal Facility, 1998).

3.5.2 Future sea-level rise in the Pacific

The IPCC Second Assessment Report anticipates a sea-level rise over the next 100 years of about 5 mm pa with an uncertainty range of 2–9 mm pa (Warrick et al., 1996) equivalent to a rise of 13–94 cm by the year 2100. This is an increase of 1.5–3.5 over the historical rate of rise.

The major factors affecting both historical and projected future sea-level rise are (Warrick et al., 1996):

- thermal expansion of the ocean;
- melting of glaciers and small ice caps;
- contributions from the Antarctic ice sheet;
- contributions from the Greenland ice sheet;
- possible changes to surface water and groundwater.

Location	**Length of data** *(months)*	**Trend** *(mm pa, 1992–1998)*	**Uplift/subsidence** *(mm pa, 1992 –1996)*
Cook Islands	69	+6.4	+0.6
Tonga	70	+24.9	+0.6
Fiji	74	+4.2	+0.5
Vanuatu	60	+7.8	-0.3
Samoa	69	-20.2	-1.8
Tuvalu	67	-37.5	+0.2
Kiribati	67	-26.8	+0.5
Nauru	64	-30.0	+0.1
Solomon	50 [with gaps]	-40.1	+0.1
PNG	37 [with gaps]	-29.5	-0.2
Marshall	63	-0.6	-0.1

Table 10: Recent sea-level trends from the South Pacific Sea Level and Climate Monitoring Project (1992 to November 1998; National Tidal Facility; 1998) and uplift/subsidence as inferred from relative movements between the monitoring station and nearest deep benchmark (1992–1996; Turner, 1997)

The individual influences of these factors vary over time. For instance, surface warming of the oceans and changes to small glaciers occur relatively quickly, whereas deep warming of the ocean and large changes in the mass balance of the Antarctic ice sheet occur more slowly and may have a continuing impact on sea level, lasting for hundreds to thousands of years.

Although temperature is the most significant driver of sea-level rise via thermal expansion and glacial melting, changing precipitation can affect the mass balance of both small and large ice masses. For example, warmer temperatures are expected to lead to an increase in precipitation over Antarctica, leading to an increase in snow accumulation which remains frozen under prevailing temperatures, providing a negative influence on sea level over the next century (Smith et al., 1998). On the other hand, if warming occurs in areas close to 0°C, ice may melt, contributing to sea-level rise. This possibility has been raised for the Greenland ice cap (Thompson and Pollard, 1997).

Increased breakup or outflow of ice from grounded ice shelves and glaciers may also contribute to sea-level rise (e.g. Bindschadler et al., 1998). The possibility of the West Antarctic ice sheet contributing to large-scale, rapid melting has been raised, although the risk of this occurring in the next century is low (Oerlemans and van der Veen, 1998; Bindschadler, 1998). Recent observations of basal melting on grounded ice sheets in the Antarctic have led Oppenheimer (1998) to conclude that the overall future contribution of the Antarctic may be positive, with melting exceeding accumulation by 0–19 cm per century, exceeding the maximum limit for Antarctica of 8 cm per century produced by MAGICC. However, this is only one scenario of three he proposes.

Scenarios for global sea-level rise have been discussed extensively in Part One of this project (Jones, in press). These scenarios allow for changes to the first four influences on sea level listed above but not for changes to terrestrial water balance.

Regional sea-level rise scenarios require dynamic, spatially explicit models that adequately represent ocean processes, ice-ocean and land-ice interactions in addition to conserving the mass balance of water in the ocean and atmosphere, and on land. At present, models are much more limited than this, as shown by previous studies.

Gregory (1993) simulated water-balance changes from ice sheets and thermal expansion based on GCM to produce spatial maps of sea-level changes. The outcomes were limited by the initial conditions of the GCM (Gregory, 1993), and the assumption of constant salinity (Jackett et al., submitted). Cubasch et al. (1994) also produced maps of global sea-level rise averaged from an ensemble of four short runs utilising different initial conditions to 2035. This showed that the results were not subject to different initial conditions, as was the case for rainfall and soil moisture, due to the thermal inertia and slow response of the oceans.

Jackett et al. (submitted) have used output from three simulations of the CSIRO GCM (separate to those analysed in this report) to apply an inversion technique that converts changes in the heat profile of the ocean and the speed of ocean mixing into equivalent sea-level changes. This technique accounts for thermal expansion only.

An important result from coupled GCMs is the reduction of thermohaline circulation (Kattenberg et al., 1996)—the rate at which surface waters are drawn into the deep ocean, ventilating it over long periods of time. In the MAGICC upwelling-diffusion energy-balance model used in Part One of this report, this effect tends to increase the simulated rate of sea-level rise. However, according to Jackett et al. (submitted), their model produces the opposite effect.

Two simulations utilising output of the CSIRO Mark 2 GCM demonstrate similar patterns of sea-level rise. In the SPREP region, values are close to the global average of 18 cm in 2100 produced from an IS92a CO_2-only run and 20 cm produced from an IS92a CO_2-equivalent run scaled downwards to allow for the radiative effects of aerosols. These global averages can be compared to the value of the thermal expansion component of sea-level rise from MAGICC for the IS92a scenario with a sensitivity of 2.5°C which is 27 cm in 2100 (Warrick et al., 1996). The sensitivity of the CSIRO GCM is 2.1°C at doubling of equivalent CO_2 (Hirst, pers. comm.) so as it is slightly less sensitive to climate. Therefore, the figure of 20 cm in 2100 is comparable to the MAGICC estimate. Jackett et al. (submitted) expect a reduction in the thermal component of sea level of between 12 and 30 per cent from their model compared to those produced by MAGICC and given in Warrick et al. (1996).

Regarding spatial patterns of sea level, Jackett et al. (submitted) state that 'we have not yet accumulated enough knowledge on how regional variations in sea-level rise depend on the topography of the ocean model and on the details of the changes in wind stress as the atmosphere and ocean circulations change'. For this reason, maps of sea-level change in the Pacific have not been reproduced, as the work needs to be corroborated by results from other research groups. As the results produced by Jackett et al. (submitted) show changes in thermal expansion for the SPREP region distributed around average global level, global scenarios of sea-level rise produced by models such as MAGICC are recommended for continuing impact assessments utilising sea-level rise.

3.6 ENSO and links to climate change

3.6.1 ENSO under the current climate

ENSO plays a major role in determining the interannual variability of rainfall in South Pacific islands. Previous studies describe a close connection between SSTs in the Pacific, the Walker Circulation and rainfall in this region (e.g. reviews by Philander, 1989; Trenberth, 1996). During La Niña events, more rainfall is received due to strong convective activity over the western Pacific. During El Niño events, more rainfall is received in the central and eastern Pacific regions where most of the South Pacific islands are situated. Trends in temperature and rainfall in the South-West Pacific show increases after 1950 and have systematic changes after the mid-1970s (Salinger et al., 1995). The present rainfall climatology also indicates strong interannual variability that is linked to ENSO.

Delcroix (1998) provides an overview of oceanic and atmospheric variability in the equatorial Pacific on ENSO time scales with the analysis covering the years 1961–1995. On ENSO time scales, sea surface temperature amplitudes were found to be greatest in the central eastern Pacific basin, precipitation anomalies were greatest in the west and sea-level changes were large in both the east and west. El Niño/La Niña changes summarised by Delcroix (1998) include in the central and eastern Pacific, warmer/colder sea surface temperatures and higher/lower sea level. In the western Pacific, a rainfall increase/decrease occurs east of about 150°E (with a decrease towards Australia) with a sea level increase/decrease. Sea-surface temperature variations with ENSO can be ±3°C in the Pacific, compared to normal variations of less than half that (Evans et al., 1998). Sea-level variations on the ENSO time scale can be ±25 cm (Delcroix, 1998).

Large-scale ENSO-related sea-level changes are represented by two types which be separated into east-west and north-south movements (Delcroix, 1998). Low/high levels in the western Pacific lag about six months behind high/low levels in the eastern Pacific for El Niño/La Niña. The changes in the west extend away from the equator while the changes in the east remain in a narrow band (Delcroix, 1998). A north-south movement lags the Southern Oscillation Index by about one year, where it oscillates about 6°N in the western central and 10°N in the far western Pacific basin.

Understanding these phenomena is critical for developing scenarios that can be tested using climate models under current climate (Delcroix, 1998) as they involve larger-than-normal interannual variations that lead to significant impacts in the Pacific island states. This knowledge is obligatory for exploring how these phenomena may behave under climate change.

3.6.2 Simulated ENSO-like behaviour under climate change

As ENSO plays an important role in modulating the climatic variability of this region, we have looked at the results of the CSIRO GCM. This model simulates a greater warming under enhanced greenhouse conditions over the eastern Pacific as do other coupled-ocean atmosphere models (Meehl and Washington, 1996; Timmermann et al, 1998; Figure 11). Possible changes in ENSO under enhanced greenhouse conditions were examined by calculating the frequency distribution of SST anomalies after removing the long-term trend in SST over the western Pacific, NINO4 and NINO3 regions.

Differences in SSTs between the western Pacific and NINO3 regions were also calculated to examine the changes in the east-west, or Walker Circulation (Bjerknes, 1969). First, long-term trends in SSTs in the western Pacific and NINO3 regions were removed. Second, the difference in SST anomalies (deviations from long-term trends) between the western Pacific and NINO3 regions have been calculated. Third, average temperature differences between the western Pacific and NINO3 regions for the first and last 40 years have been calculated and then these differences have been added to SST anomalies that were derived as differences between two regions after removing the long-term trends. Last, frequency distributions have been calculated using the SST anomalies added to regional temperature differences.

Figure 21 shows the frequency distributions of SST anomalies for the NINO3 region that are derived as anomalies from long-term trends. This figure also shows the degree of interannual variability simulated between the first and last 40-year periods. The distribution of SST anomalies in the NINO3 region does not change significantly, contrary to results obtained from the higher resolution DKRZ GCM by Timmermann et al. (1998).

Figure 22 shows the frequency distribution of SST anomalies as differences between the western Pacific and NINO3 regions added to mean temperature change between these regions. This figure indicates a tendency for a weakening of the

Walker Circulation as represented by SST anomalies. On the basis of existing evidence about the relationship between ENSO, the Walker Circulation and Pacific islands (Bjerknes, 1969; Rasmusson and Carpenter, 1982 and others), a weakening of the Walker Circulation and SST anomalies toward the El Niño phase should lead to increased rainfall in the central and eastern Pacific. This is consistent with the simulated widespread increase in rainfall in the eastern Pacific in all GCMs analysed (Figures 15 and 16).

One plausible explanation for increased rainfall under enhanced greenhouse conditions is the increased water-holding capacity in the warmer atmosphere that leads to more rainfall. The increase in rainfall across the eastern and central Pacific is clearly related to greater warming in these regions. This is interpreted as a more El Niño-like mean state under climate change. The consistency of this result in other GCMs, notably in the north Polynesian region (Tables 7 and 8), suggests this is a robust result. However, the effect of this outcome on the many aspects of climate variability in the Pacific that are part of ENSO under current climate, remains uncertain, as spatial patterns with ENSO outside the equatorial zone may change.

3.7 Tropical cyclones and links to climate change

Tropical cyclones are formally defined as tropical weather systems with distinct cyclonic circulations that have maximum sustained winds stronger than 17 ms^{-1}, or about 61 km hr^{-1}. The SPREP region is regularly affected by tropical cyclones, often resulting in significant damage. In the North Pacific, major typhoons regularly affect parts of the SPREP region. In December 1997, Supertyphoon Paka crossed the Marianas islands, passing only about 30 km north of the main business centre on Guam. Damage in these islands was in the hundreds of millions of US dollars. In the South Pacific, in January 1998, Tropical Cyclone Ron passed over the Kingdom of Tonga, damaging about 67 per cent of the buildings on the island of Niuafo'ou and destroying most of the island's indigenous agricultural production. Other serious storms have occurred frequently in the past in many parts of the SPREP region. Therefore the effect of climate change on the behaviour of tropical cyclones needs to be considered.

Figure 23 shows the occurrence of tropical cyclones in the SPREP region.

Occurrence is defined as the number of days that the centre of a tropical cyclone was detected in a region of about 250 km by 250 km over a period of 20 years. Thus Figure 23 shows approximately how frequently the various parts of the SPREP region are affected by tropical cyclones.

The likely effect of climate change on tropical cyclones is controversial because of the difficulties involved in simulating these relatively small, intense weather systems in climate models with coarse horizontal resolution.

Low-pressure systems that have several of the observed characteristics of tropical cyclones can be generated by climate models, but their structure is greatly simplified. These tropical cyclone-like vortices (TCLVs) can be detected in climate model simulations, and are seen to form in similar regions to observed cyclones, have similar tracks, but in many cases have weaker intensities (Bengtsson et al., 1995; Walsh and Watterson, 1997).

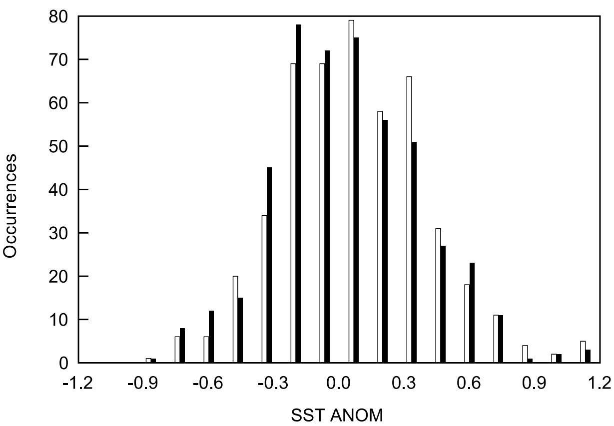


Figure 21: Frequency distribution of NINO3 SST anomalies during the first 40 years (blank bars) and last 40 years (solid bars) of the transient experiment. The frequency distributions were calculated after removing the linear trends.

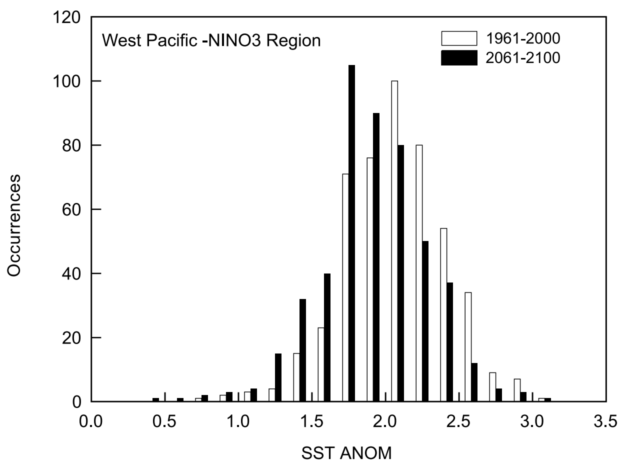


Figure 22: Frequency distribution of SST anomalies based on the difference between western and eastern Pacific. In this figure, differences between temperature increases between the first and last 40 years are also added.

Additionally, the simulated TCLVs lack several of the important observed features of real tropical cyclones, such the eye. Nevertheless, in the last few years a number of inferences have been made about the effect of climate change on tropical cyclones, and are reported here.

When assessing the affect of climate change on tropical cyclones, changes in the following phenomena are important:

- numbers;
- intensities (i.e. maximum wind speeds or central pressure);
- regions of formation;
- regions of occurrence i.e. what happens to tropical cyclones after they form; and
- the combined effect of any changes in tropical cyclone characteristics and sea-level rise.

3.7.1 Tropical cyclone numbers

Currently, little can be said about the likely impact of climate change on tropical cyclone numbers. Although climate models can give a reasonable representation of the current distribution of tropical cyclone formation (e.g. Bengtsson et al., 1995; Walsh and Katzfey, 1998), the real response of the models to climate change, in terms of their ability to predict changes in the frequency of tropical cyclone formation, is unknown. An example of this problem is provided by the study of Bengtsson et al. (1995), in which tropical cyclone formation in the Southern Hemisphere was more than halved under enhanced greenhouse conditions. In contrast, Yoshimura et al. (1999) simulate only a small decrease in tropical cyclone numbers under enhanced greenhouse conditions, while Walsh and Katzfey (1998) predict little change.

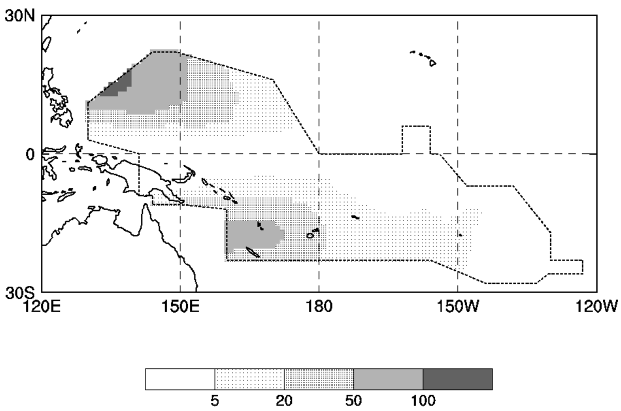


Figure 23: Tropical cyclone occurrence in the SPREP region. Units are numbers of cyclone-days per 5-degree grid square over a 30-year period. Data from the Joint Typhoon Warning Center, Guam.

Thus, consensus on this issue has yet to be achieved.

Tropical cyclone numbers in the Southern Hemisphere section of the SPREP domain are closely associated with the phase of ENSO. Revell and Goulter (1986) and Hastings (1990) showed that in negative SOI seasons (El Niños) tropical cyclones tended to form further north-east, and in positive SOI seasons (La Niñas) further south-west in this region. Therefore, any change in ENSO characteristics in a warmer world would possibly have an effect on the formation of tropical cyclones in this region. As discussed above, the effect of climate change on ENSO is currently uncertain but there is a possibility of a more El Niño-like mean state.

In the Northern Hemisphere section of the SPREP region, the relationships to ENSO in the current climate are more equivocal (Chan, 1985; Lander 1994). In general, the simultaneous relationship between the SOI and tropical cyclone formation in these locations is weak. Nevertheless, Chan (1985) states that the number of cyclones east of 150°E (in the Marshall Islands) increases during an El Niño event. Additionally, Lander (1994) found that a large number of formations occur late in the season of an El Niño event east of 160°E and south of 20°S (again, in the Marshall Islands). During a La Niña event, no such formations occur. Thus any change in ENSO under climate change will alter tropical cyclone formation north of the equator in the SPREP region, but less so than south of the equator.

In summary, no definite statements can be made about the effect of climate change on tropical cyclone numbers in the SPREP region.

3.7.2 Tropical cyclone intensities

There is an emerging consensus that climate change may lead to some increases in tropical cyclone mean and maximum intensities. Most relevant in this context is the recent paper of Knutson et al. (1998). In this work, cyclones were inserted into a high-resolution numerical model over the western North Pacific region and run under current climate and enhanced greenhouse conditions, and the evolution of their intensities was studied over several days. In general, both mean and maximum intensities increased somewhat in a warmer world, although the statistical significance of the result was marginal.

South of the equator, Walsh and Ryan (1998) used a similar, but independently developed technique, to examine cyclone intensity change near Australia. These results would be most relevant to the section of the SPREP region west of the dateline. They found increases in mean cyclone intensity that were not statistically significant, except for increases in maximum storm intensities that were marginally significant (at the 90 per cent level). Similar results have been found in the very recent study of Knutson and Tuleya (1999). There is no work which would be directly applicable to the Southern Hemisphere tropical regions east of the dateline, although the emerging consensus in favour of a general increase in cyclone intensities may apply there also.

Plausible theoretical reasoning also supports some increase in cyclone intensities. This approach involves formulating a parameter that can utilise the output of a GCM to derive typical maximum cyclone intensities at various locations, depending upon the average atmospheric and oceanic conditions that are encountered. This technique assumes that the maximum potential intensity (MPI) of a tropical cyclone is related to the total amount of energy available in the atmosphere. Such parameters (Emanuel, 1987, 1991; Holland, 1997) suggest modest to moderate increases (10–20 per cent) in maximum tropical cyclone intensities in a warmer world. Unfortunately, as with most climate modelling estimates, the uncertainties in this technique are fairly large, as the errors in a GCM simulation of the MPI in the current climate are typically larger than the forecast change in intensity.

In summary, a general increase in tropical cyclone intensities, particularly of maximum intensities, in a warmer world is now considered likely. Whether this would lead to significant real-world impacts in terms of substantially reduced intervals between severe storms at specific locations still needs to be established.

3.7.3 Regions of formation

Based on our current understanding of the formation of tropical cyclones, there is little reason to expect substantial changes in their current regions of formation except perhaps due to changes in ENSO and the Walker Circulation. Tropical cyclones form almost entirely in the tropical regions, and this is not expected to change. Under the current climate, tropical cyclones only form where sea surface temperatures are at least 26°C. While it is true that the area likely to experience such temperatures will increase as a result of global warming, there are good theoretical reasons to believe that the sea surface temperature does not directly determine whether a tropical cyclone will form. Rather, tropical cyclone formation is related to atmospheric conditions, and the sea surface temperature threshold is determined by these conditions. Under an enhanced greenhouse climate, it is therefore considered likely that a new, higher sea surface temperature threshold would be established for tropical cyclone formation, since the atmospheric conditions would also change (Holland, 1997). The net result would be little change in the regions of cyclone formation. The results of Walsh and Katzfey (1998) tend to support this conclusion.

In summary, no change in regions of formation is expected, except possibly in association with changes in ENSO.

3.7.4 Regions of occurrence

Occurrence is defined above. Like cyclone numbers, occurrence in the Southern Hemisphere part of the SPREP domain is tied closely to the phase of ENSO. In the South Pacific, Basher and Zheng (1995) found significantly different relationships between the phase of ENSO and tropical cyclone occurrence west of 170°E from those east of this meridian. They found that the primary influence on tropical cyclone occurrence west of this line was the local sea surface temperature, while east of it the SOI played a large role. The relationship between the SOI and cyclone occurrence becomes stronger as one travels east in the SPREP region south of the equator: relationships between cyclone occurrence and the SOI are moderate west of 170°E and in the region 170°E–170°W, and become very strong in French Polynesia. Thus any effect of climate change on ENSO could potentially have large impacts on cyclone occurrence, particularly in the eastern regions of the SPREP domain.

Another effect that may be important in some regions of the SPREP domain is the possibility that increased sea surface temperatures in a warmer world may give cyclones more energy to persist longer than they would in the current climate, thus tracking further poleward. This mechanism could increase cyclone occurrence in the poleward margins of the SPREP domain (i.e. the Northern Marianas and southern French Polynesia). Some supporting evidence for this effect has been obtained by Walsh and Katzfey (1998), although it remains tentative. More work needs to be done to investigate the reality and magnitude of this effect.

In summary, there is some evidence to suggest that tropical cyclones might track further polewards under enhanced greenhouse conditions. At present, this is not a firm conclusion, but merely indicates that the issue should be pursued further.

3.7.5 The combined effect of changes in tropical cyclone characteristics and sea-level rise

This issue is relevant to a most damaging aspect of tropical cyclone impacts, the storm surge. This is a dome of water forced ahead of the storm by strong winds. The vulnerability of a particular location to storm surge varies greatly, depending upon height above sea level, orientation of topography, and the near-shore slope of the ocean floor. This means that vulnerability to storm surge must be evaluated specifically for each location.

As mentioned above, it is now considered likely that climate change will lead to some increase in maximum tropical cyclone wind speeds and lower central pressure. Maximum cyclone wind speed is more relevant to the issue of storm surge damage than average cyclone wind speed, as generally only the more intense storms cause storm surges large enough to be a concern, although this can also depend strongly upon location and elevation.

Sea-level rise and storm surge effects are linearly additive; in other words, whatever storm surge occurs at a particular location under the current climate can simply be added to the sea-level rise at that location (McInnes, personal communication, 1998). The possibility that the two effects could combine to change the return periods of storm surges (the typical length of time between events) is currently being investigated for the Cairns region in Australia (McInnes et al., 1998). This methodology could be applied to locations in the SPREP region. Future sea-level rise is discussed in Section 3.5.

In summary, the combined effect of increases in cyclone intensities and sea-level rise could potentially have substantial effects on storm surges in the SPREP region. However, these effects are very location-dependent, and would need to be examined separately for each location.

3.7.6 Analysis of recent climate model results

As mentioned earlier, a major limitation of constructing climate change scenarios for tropical cyclones is the lack of confidence in predicting the effect of climate change on ENSO. Nevertheless, current coupled climate models do simulate aspects of the behaviour of ENSO in the present climate (Wilson and Hunt, 1998), and it is instructive to evaluate the ability of such models to simulate the observed behaviour of tropical cyclones.

Many coupled models suggest that ENSO will continue to oscillate in a warmer world in some form (IPCC, 1996). As mentioned above, under El Niño conditions, cyclones currently can occur far east of the dateline, and much further east than is typical under La Niña conditions. It is therefore of interest to examine the question of whether this known geographical relationship between cyclone formation and ENSO can be simulated in a coupled model, and whether these relationships continue in some form in a warmer world. This analysis has initially been performed only for the Southern Hemisphere regions of the SPREP domain using the regional climate model DARLAM (Table 5).

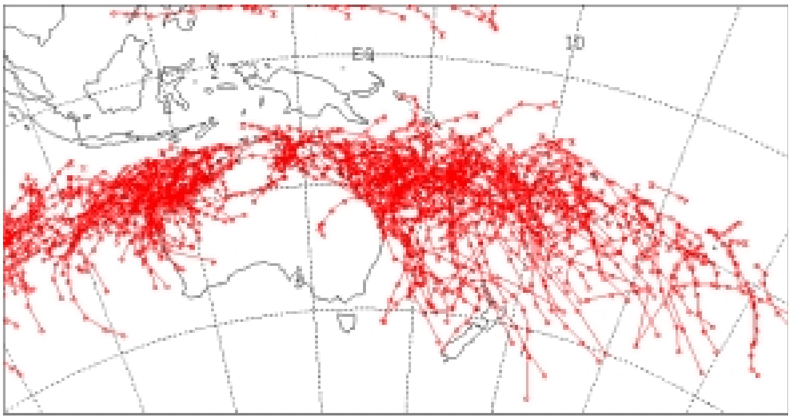
Observed tropical cyclone tracks and simulated tracks are shown in Figure 24, for the observed January–March period from 1967 to 1986 and a simulated 20-year January–March period. In general, simulated tracks occur in regions similar to observed tracks, although the density of simulated tracks in some regions is larger than observed. This is more clearly shown in Figure 25, which gives the details of occurrence east of 145°E over the South Pacific region. This clearly shows too much TCLV occurrence in the model compared to observed tropical cyclone occurrence in reality.

Under 3×CO_2 conditions, occurrence tends to decrease somewhat north of 30°S, largely due to a decrease in TCLV formation. Whether this is a reliable response remains uncertain, since as mentioned earlier, there is little consensus regarding changes in cyclone numbers under climate change. South of 30°S, there is a slight increase in occurrence despite the decrease in formation. This is consistent (at least qualitatively) with the results of Walsh and Katzfey (1998), who found an increase in occurrence at higher latitudes. The results of this model in this context must be regarded as preliminary. Additionally, the horizontal resolution of this model simulation is insufficient to address the question of intensity change.

3.7.7 Changes in the relationship between cyclone formation and ENSO

The ENSO variations studied here are represented by the Southern Oscillation Index (SOI) diagnosed from the model. SOI is defined here as the normalised anomaly of the mean sea-level pressure difference between the model grid points nearest to Tahiti and to Darwin. Proxy El Niño and La Niña years were selected from the DARLAM simulation by examining the model-generated SOI.

Both El Niño and La Niña years were selected for current climate and 3×CO_2 conditions. The amplitude of ENSO SST variations generated by the forcing GCM is considerably weaker than observed, but the SOI variations generated by DARLAM are similar to observed.



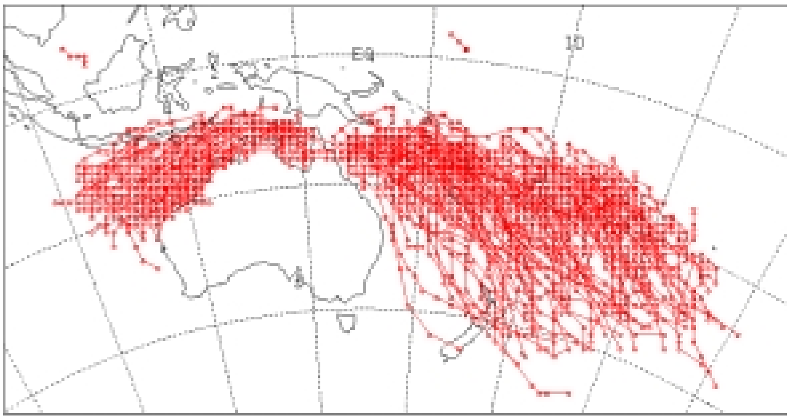


Figure 24: (top) Observed tropical cyclone tracks for the period 1967–1986, for January–March; (bottom) simulated regional climate model TCLV tracks for a 20-year period, for January–March conditions

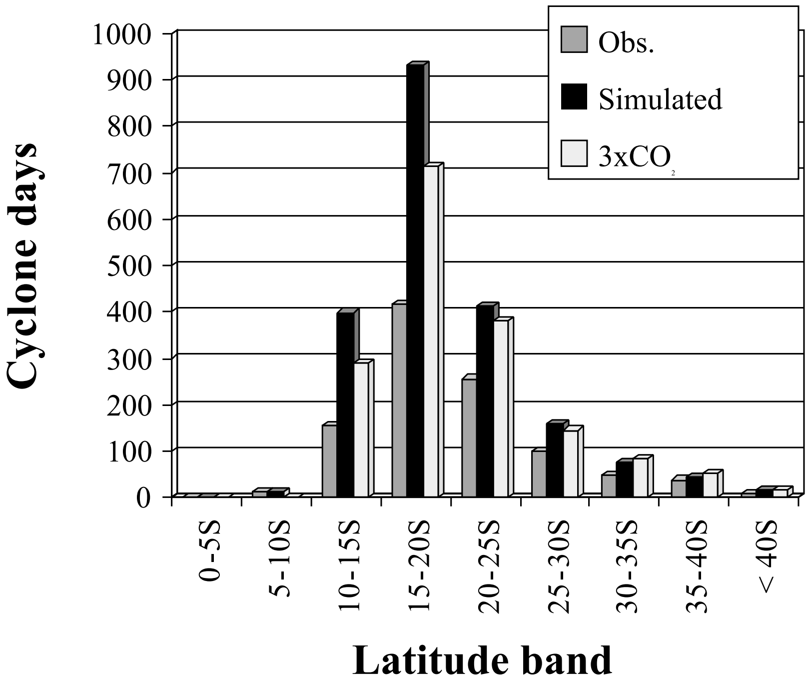


Figure 25: Occurrence by latitude band for observed tropical cyclones and TCLVs east of 145°E simulated in the current climate and under 3×CO_2 conditions

There are distinct differences in the TCLV simulations under El Niño and La Niña conditions. Figure 26 compares regions of occurrence for El Niño and La Niña conditions in the current climate. Occurrence is defined here as the number of storms passing through a box 250×250 km, and thus describes the formation and subsequent movement of the TCLVs. For clarity, we calculate totals along the latitudinal coordinate of the DARLAM model grid shown in Figure 8, sum up the cyclone occurrence for each longitudinal value, and plot the results for El Niño and La Niña conditions. Two longitudinal values at a time are summed together. This procedure gives 54 data points instead of the 108 in the model.

Under El Niño conditions, for both the current climate and 3×CO_2 conditions, the TCLVs in the South-West Pacific tend to travel considerably further east than under La Niña conditions. In addition, the occurrence of TCLVs is much larger close to the coast of Australia under La Niña conditions. This is in agreement with the observed behaviour (Basher and Zheng, 1995). Given that coupled model predictions suggest that ENSO is likely to continue under enhanced greenhouse conditions (Timmermann 1998; Wilson and Hunt, 1998; Section 3.6), the results shown here suggest that this geographical relationship between ENSO and cyclone occurrence may continue.

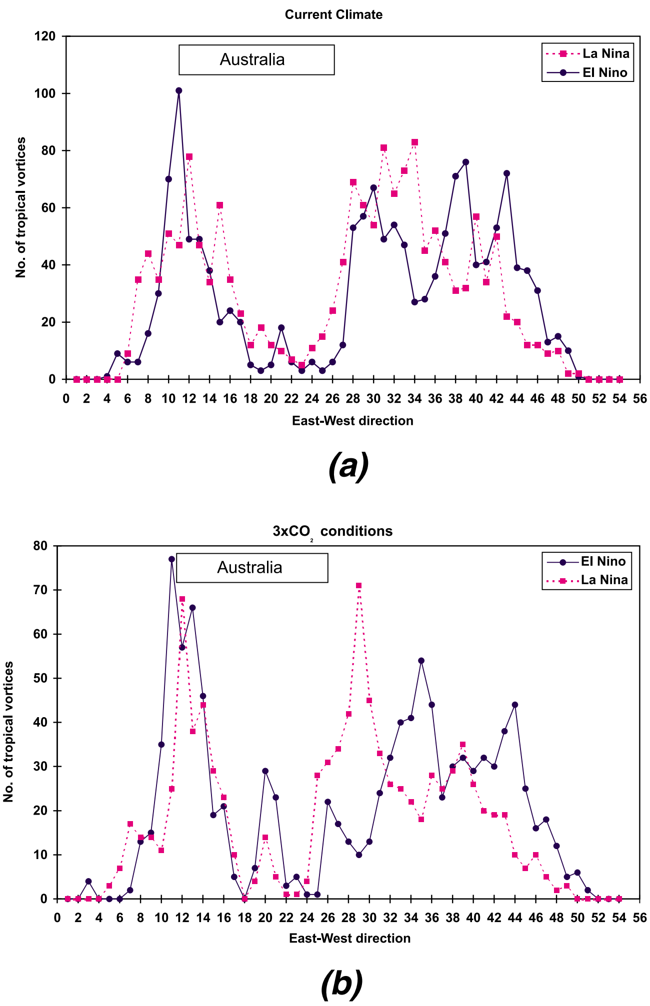


Figure 26: Number of simulated TCLVs in each longitudinal coordinate column, summed two at a time, for (top) current climate conditions, and (bottom) 3×CO_2 conditions. The approximate longitudinal extent of Australia is indicated by the box.

4. Impacts and risk

4.1 Summary of potential impacts

Impacts of climate change and variability, including sea-level rise, have been extensively reviewed in several reports and conference proceedings, including Nurse et al. (1998) in the *IPCC Special Report on the Regional Impacts of Climate Change: An Assessment of Vulnerability*. Other reports and proceedings include Hay (1998), Pittock (1999), a special issue of the *Journal of Coastal Research* (Leatherman, 1997), a special issue of *Climatic Change* on the impacts of climate change on tropical forests (Markham, 1998), and earlier reports by Pernetta and Hughes (1990), Hay (1991), and Hay and Kaluwin (1993).

4.1.1 General considerations

A large array of possible climatic change and sea-level stress factors may contribute to impacts in Pacific island countries. These include increases in average and extreme temperatures of both air and water; changes in the amount and seasonality of rainfall and in extreme rainfall rates; changes in the number, location and intensity of tropical cyclones; the absolute amount of sea-level rise and its rate, and of extreme sea-level excursions; increases in carbon dioxide concentrations in the atmosphere and related changes in ocean acidity; and possible increases in ultraviolet radiation.

Changes in the east-west temperature gradient across the tropical Pacific, which is closely related to ENSO, would also affect the position of the South Pacific Convergence Zone, and relates to changes in cloudiness, regional rainfall, temperature and sea level.

As noted above in Section 3, many of these factors are poorly known, even as to the sign of change at the local scale. Even when ranges of uncertainty are known, it is difficult to quantify them with any confidence. Some of the stress factors, such as increases in average temperature, sea level, and CO_2 concentrations, can be expressed quantitatively in terms of a range of possibilities (and with some measure of probability) at some future date (Pittock and Jones, 1999). In such cases, the risk of particular impacts or outcomes can be estimated if the impacted systems are sufficiently well understood. (However, some of the impacted systems are extremely complex and not yet well enough understood to identify how they will respond to certain stresses.)

Other climatic stress factors, such as changes in extreme events and in local rainfall remain very uncertain, so the magnitude and probability of impacts cannot yet be determined. Non-climatic stress factors, such as land-use change, chemical pollutants, turbid freshwater outflows and extraction of building materials may also impact on island systems, interacting with climatically driven stresses. In the following sections we will discuss what is known about some impacted sectors, and go on to discuss a risk assessment approach to estimating impacts.

4.1.2 Water resources

Water resources in small island states come from various sources: direct rainfall; stream flow and surface water storages; fresh water lenses; or desalinisation and imports. The last (the dominant source e.g. in the Maldives) is of no concern here, except to note that any increase in the price of fossil fuels as a result of mitigation programmes may make desalinisation less economic. Changes in direct rainfall, as we have seen above, are uncertain in particular locations. However, the effectiveness of rainfall will decrease due to increased evaporation, which will dry out soil or vegetation faster and increase water demand. Stream flow will be even more affected by the cumulative effects of increased evaporation from watersheds, except in the case of heavy rainfalls which may quickly saturate catchments.

The frequency and intensity of heavy rainfalls will affect streamflows, soil erosion, and loadings of sediment, nutrients and pollutants. In general, where average rainfall does not decrease, we may expect increased extreme rainfall events and thus increased heavy flow events and loadings (Fowler and Hennessy, 1995). This may adversely affect downstream sectors, including reefs and reef lagoons via lower salinity, increased turbidity, excess nutrients and damaging pollutants

(Larcombe et al., 1996). Resulting impacts in many parts of the South Pacific will depend on the pattern of rainfall changes associated with possible changes in ENSO. Impacts will also be exacerbated by non-climatic stresses such as land-clearing, overgrazing, excess application of fertilisers and chemicals, and sewage and other pollutants. Many of these non-climatic stresses will be related to socio-economic factors such as population growth.

Fresh water lenses are a major source of water in many low-lying atoll states (Buddemeier and Oberdorfer, 1990). They will be adversely affected by any decrease in island area due to sea-level rise, and by overtopping by waves or storm surges associated with tropical cyclones. Recharge events will also be affected by rainfall intensity with, in general, light rains being more subject to losses by surface evaporation, and heavy rains to losses by run-off into the sea. Depending on the vegetation, topography and porosity of the overlying soil, effects of rainfall intensity will differ from one location to another.

An important factor in regard to water resources is the seasonality of rainfall with, in many small island states, excess rainfall in the wet season(s) and water shortages in the dry season(s). Changes in the low rainfall seasons are generally of greatest importance, but often least well known, and certain to be most affected by increased temperatures and evaporation. In general one might expect that global warming will lead to greater water stresses in the dry seasons, so that island states with extended dry seasons will be most vulnerable. Further analyses of the results in Tables 7 and 8 are needed to determine whether this risk is real.

4.1.3 Coastal areas

Coastal areas will be affected by mean sea-level rise and extreme sea-level excursions, both of which will lead to erosion, inundation and changes in alignments (Leatherman, 1997; Nurse et al., 1998). The last will occur especially as a result of changed wave climates. Such effects will all be strongly modified by the presence of coral reefs and reef flats, which in general provide some protection against waves and extreme events.

Extreme sea-level events will occur, as now, due to tropical cyclone storm surges (localised, short-lived, but up to several metres high, with high winds and wave energy) and ENSO (fluctuations of ±25 cm, lasting months, depending on the phase of ENSO occur in various locations: Delcroix, 1998).

Changes to ENSO are still quite uncertain, although El Niños and La Niñas are expected to continue about some changed average condition at least through 2100.

As seen above, tropical cyclones are expected in general to be somewhat more intense (0 to 20 per cent higher wind speeds and lower central pressures by the time of doubled CO_2). This will appreciably increase the magnitude of storm surges, although changes in the return period for a given surge height will also depend on possible changes in storm numbers and location. Total numbers are not expected to change substantially, but tropical cyclones (TCs) may tend to travel further polewards, thus increasing the frequency of occurrence at some higher latitude island locations. Changes in ENSO and the position of the SPCZ will also modify the frequency of occurrence at particular longitudes, since more TCs occur in the Coral Sea region in El Niño years, and more near the dateline in La Niña years. In combination with increased storm surge height and mean sea-level rise, higher wave energy will tend to overtop fringing reefs, mobilise more debris from the reef flats, and increase wave setup.

The growth rate and viability of fringing reefs may also be adversely affected by a number of stresses from more intense run-off (lower salinity extremes and higher pollution levels and turbidity), coral bleaching due to a combination of factors including high water and air temperatures, and reduced calcification rates due to increased acidity of sea water from higher CO_2 levels in the atmosphere (Buddemeier et al., 1998; Gattuso et al., 1999). In some cases, reduced growth rates of coral may mean that the reefs fail to keep up with rising sea level, threatening reef viability and reducing the effectiveness of the reefs as barriers to wave action. This is most likely where reefs are already stressed due to other factors, or where local tectonic movements mean that there will be a greater than average rate of sea-level rise. Regional variations in rates of sea-level rise will also be a significant factor.

The response of reefs to all these factors is very complex and not well understood. However, a rough estimate of a generalised threshold for the drowning of coral reefs (Hopley and Kinsey, 1988), at about 8 mm y^{-1} sea-level rise, suggests that, with some reduction in growth rates due to acidification of the water, such a threshold may be reached during the 21st century in some locations (Pittock, 1999). In many cases, this may not threaten reef survival over the next century but could lead to changes in species composition with more rapid growing branching species favoured, thus reducing the strength and stability of reef extension over the longer term.

4.1.4 Agriculture

Agricultural production in the small island states includes subsistence agriculture, particularly in many of the low-lying atolls and inland villages of the larger islands. In addition there are cash crops, particularly sugar cane, but also copra and in some cases coffee and tea at higher elevations. Climatic change impacts will come from increased temperatures, evaporation, water demand, and changes in rainfall. Increased stresses from these causes will be greatest in the dry season, so that changes in dry season rainfall will be critical (Singh et al., 1990; Singh, 1994). Lowland crops, particularly various types of taro, may be affected by changes in the water table and salinity, which will be most affected by sea-level rise as well as by changes in water demand and pollution from growing populations (Nurse et al., 1998). Higher ambient temperatures will also lead to more rapid post-harvest deterioration of crops, which is a major problem in many locations (Aarlbersberg, 1992).

Cash crops will perhaps be more affected by changes in world commodity prices, some of which will be due to changes in supply and demand brought about by climate change elsewhere (Fischer et al., 1994; Reilly et al., 1994). Scenarios for such price changes are very uncertain, depending on what climate change scenarios are used, the degree of adaptation assumed, and how much beneficial effect of increasing carbon dioxide concentrations on crop production is assumed to occur in the field. In general C3 type crops, which include most tropical crops except maize, sugar cane and pineapple, will benefit more from the CO_2 fertilisation effect than C4 crops. However, at least in the first half of next century when warming is less advanced, mid-latitude crop production is likely to benefit more from climate change, and low latitude crops least (Pittock, 1995).

Any increase in tropical cyclone damage will have major impacts on agriculture in the small island states. As average intensity is likely to increase, numbers are not expected to change dramatically, and tropical cyclones may travel further polewards. Many small island states may experience increased coastal inundation and erosion, wind damage and riverine flooding, adversely affecting crop production and food supplies, and disrupting transport infrastructure.

4.1.5 Forests

Forests in small island states can be broadly grouped into mangrove forests in coastal zones, lowland forests, and montane forests in some higher elevation island states such as Samoa and Papua New Guinea. A comprehensive review of the impacts of climate change on such forests (excepting mangroves) can be found in the proceedings of a workshop held in Puerto Rico in 1995 (Markham, 1998).

All three forest types will be affected by a combination of climate change and other stresses, notably fragmentation due to clearing which will adversely affect regeneration and increase the exposure of the remnants to higher temperatures, soil degradation and other effects (Malcolm, 1998; Whitmore, 1998).

Mangroves are likely to be adversely affected by sea-level rise, especially where sedimentation rates are low. Indeed, mangrove communities are more likely to survive in macrotidal, sediment-rich environments such as northern Australia (Semeniuk, 1994; Woodruffe, 1995) than in microtidal, sediment-starved environments such as those in most small island states (Parkinson et al., 1994). Responses will vary between species (Ellison and Stoddart, 1991).

Lowland forests will be most affected by warming, coastal erosion, tropical cyclone damage, seasonal drought, and hydrological changes including salinisation of groundwater. Increasing seasonal drought would lead to the local extinction of some less drought-tolerant species, while invasion by more drought-tolerant species would be slow and hindered by fragmentation (Condit, 1998; Borchert, 1998). Increased fire frequency and severity may be induced by greater seasonal drought and by increased lightning frequency (Goldammer and Price, 1998), with dry conditions favouring fire spread from human ignition as has occurred recently in South-East Asia.

According to Loope and Giambelluca (1998), tropical montane forests are among the most sensitive ecosystems to climate change. They have particularly steep microclimatic gradients, and relatively small changes in macroclimate may trigger major local changes in rainfall, cloud cover and humidity. Changes in interannual variability associated with ENSO or the position of circulation features such as the ITCZ and SPCZ could lead to major drought occurrences and fire damage, while changes in tropical cyclone incidence may also have major impacts on the structure of the forests. They are particularly vulnerable to a combination of climate change and fragmentation (Malcolm, 1998).

4.1.6 Health

Human health will be affected by direct disturbances to physical and biological systems, notably water supply, agriculture and food supply, disease vector breeding grounds (from sea-level rise, changing water tables, forest disturbance and warmer temperatures).

Changes in the incidence of extreme sea-level and weather events in particular will lead to injury, loss of life and outbreaks of disease (McMichael and Martens, 1995; McMichael et al., 1996; Nurse et al., 1998).

Outbreaks of dengue fever in South Pacific islands have been particularly well documented in association with the ENSO phenomenon (Hales et al., 1996).

Vulnerability to climate-change induced health problems is greatest for populations which are already marginal (Woodward, 1997). This includes many small island states where population growth, isolation and limited resources are already problems. Additional stresses due to climate change and sea-level rise may indirectly affect human health due to their effects in increasing the problems of economic development and retardation of improvements on health services. Increasing pressures to relocate due to sea-level rise will create special problems of urbanisation and translocation to unfamiliar environments, with associated disruption of traditional social customs and relationships.

4.2 Risk assessment

Impact assessments generally follow a linear methodology, applying the results of climate change models to a range of emission scenarios, then applying the resulting range (which is bound by a lower and upper limit) to one or a series of impact models. The uncertainties surrounding the inputs and processes involved result in ranges of possible outcomes that are often very large (Section 1.2.2), a phenomena known as the uncertainty explosion (Henderson-Sellers, 1993). A simple example is projected global sea-level rise which ranges between 13 and 94 cm from 1990 to 2100 (Warrick et al., 1996).

These uncertainties manifest in the following manner:

- The wide ranges of uncertainty surrounding the various factors associated with the enhanced greenhouse effect result in very broad estimates for factors such as global warming (Section 2.2), sea-level rise (Section 3.4) and for regional climate changes (Sections 3.2, 3.5–3.7).

- These uncertainties are compounded by further uncertainties associated with our understanding of socio-economic and biophysical systems, meaning that predictions of impacts are usually too broad for setting policy or management guidelines relating to mitigation and adaptation.

- The delay between emissions and impacts may range from decades (e.g. atmospheric) to centuries (e.g. deep ocean circulation, large ice sheets).

For these reasons, although the straightforward prediction of impacts may indicate that systems are vulnerable to climate change, the method is limited in its ability to identify levels of dangerous climate change, or adaptation options as required by the UNFCC.

CSIRO Atmospheric Research is developing a risk assessment framework to try and overcome these limitations. Risk assessment is an activity that aims to maximise benefit and minimise loss in the face of uncertainty by assessing the likelihood of possible future outcomes and by using this information to influence changes in behaviour through adaptation and mitigation. Although the methodology is broadly based on the earlier technical guidelines compiled under the auspices of the IPCC (Carter et al., 1994), there are several important additions:

- The method of assessment is explicitly iterative rather than linear. Instead of prediction, the aim of the assessment becomes how to achieve or avoid certain thresholds.

- Stakeholders participate in the construction of these thresholds which are location dependent, activity dependent and value dependent. This activity becomes the starting point for adaptation.

4.2.1 Impact thresholds

While the conventional approach to impact assessment is linear and prescriptive, where climate modelling provides scenarios which are then applied to impact models to predict outcomes, the construction of impact thresholds involves a diagnostic component which addresses impacts in the initial stages of the process. Impacts, in the form of possible outcomes that are sensitive to climate, are identified and thresholds in the form of limits or benchmarks are determined. Scenarios can then be constructed for the key climatic variables affecting these thresholds.

Impact thresholds mark discontinuities that can be grouped into two main categories: biophysical thresholds that mark a physical discontinuity on a spatial or temporal scale, and behavioural thresholds, where reaching a particular state triggers a change in behaviour. This latter definition includes economic impacts.

Examples of biophysical thresholds are:

- climatic thresholds, characterised by distinct changes in conditions such as monsoon onset, or by an agreed standard, such as gale-force winds;

- environmental thresholds, which are important for the diagnosis of impacts because of their dependence on climate variability, e.g. temperature regulated breeding patterns in fauna, fire, drought and floods.

Examples of behavioural thresholds are:

- operational thresholds, which are set by benchmarking a level of performance, e.g. yield per area of a crop in weight, volume or gross income;

- critical thresholds that are used to assess the risk of an impact becoming 'dangerous' (e.g. Parry et al., 1996). This assessment involves placing values on processes and/or outcomes. Dangerous climate change occurs when one, or all systems, are harmed unacceptably.

Thresholds can represent an absolute value or rate of change over time (Figure 27). Threshold A illustrates the ability of impacts to cope with a certain rate of climate change, while Threshold B indicates an absolute level which cannot be exceeded safely. By comparing a threshold with a climate-change scenario as in Figure 27, the probability of that threshold being exceeded at any time in the future can be calculated. This forms the basis for risk analysis under climate change as proposed here.

4.2.2 Stakeholder participation

The involvement of stakeholders in a fully operational risk assessment is important. They have a role in identifying key climatic variables affecting activities to be assessed, identifying thresholds, determining whether the risk posed by the exceedence of a threshold warrants adaptation and in suggesting adaptation options for further analysis. This allows social and economic criteria to be incorporated into risk assessments, either implicitly, e.g. through the use of behavioural thresholds, or explicitly, e.g. through the use of cost-benefit analysis. An example of the implicit incorporation of economic information is in agricultural impacts, where a certain yield is required to maintain an adequate income. By carrying out such assessments on a regional basis, biophysical, economic and cultural differences can be recognised and accommodated in thresholds and adaptation options.

Adaptations can reduce the vulnerability of a threshold to change by making it less likely to be reached, or by raising the threshold itself, but they will involve initial effort and expense with some choices being more effective than others. Adaptation options can be compared within the risk assessment framework. The involvement of stakeholders in this step is needed to both suggest adaptation options for assessment and to perform reviews to determine which are likely to be appropriate on economic, social, cultural or technological grounds.

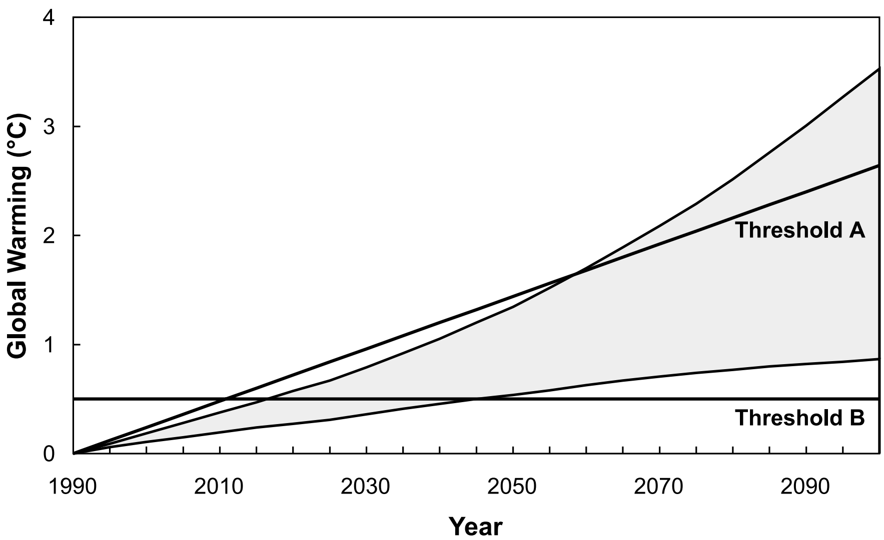


Figure 27: Hypothetical examples of rate-of-change (A) and absolute (B) thresholds related to global warming superimposed on the IPCC (1996) global warming projections

4.2.3 Estimating the effect of the Kyoto Protocol using risk analysis

The broad framework of calculating risk and applying mitigation and adaptation options to reduce that risk is referred to as risk assessment.

The task of calculating the probability of one or more thresholds being exceeded under climate change is risk analysis.

The results of the global scenarios calculated in Section 1 and important coastal-related impacts discussed in Section 3.0 are used to perform a simple risk analysis. Because of the importance of coastal impacts within the SPREP region, risk analysis is demonstrated using two hypothetical thresholds relating to sea-level rise and coral reefs. Further risk analysis is carried out to analyse the effect of the Kyoto Protocol on the risk profile of those thresholds.

It must be stressed that this example is intended to demonstrate how a risk assessment can be applied to impacts, and is not attempting to quantify risk for an actual coastline for use in planning or deciding policy. Likewise, the effect of the Kyoto Protocol is only assessed relative to the simplified example given here, but will vary depending on the sensitivity of impacts to climatic factors reduced through application of the protocol.

After creating climate change scenarios as in Section 1 and 2, and identifying critical variables that affect the impacts in question, the next step of a risk assessment is to identify thresholds for risk analysis. For this example, two phenomena associated with coastal impacts are assessed: sea-level rise and the calcification rates of reef communities.

A hypothetical coastline is backed by important urban and resort development. It is adjacent to a lagoon protected by a fringing reef. The fringing reef protects the coastline from severe wave damage during most storm surge events. Important coast-based industries are tourism and fishing.

Sea-level rise is first addressed as an absolute threshold. In this case, the example coastline has been surveyed and after consultation with planners, engineers, scientists and local land-users, a rise of 50 cm in sea level has been identified as a critical threshold. Factors which may influence such a decision include a high probability that storm surges and/or erosion will damage built infrastructure, freshwater aquifers may become affected by salt-water intrusion, or that important estuarine tidal ecosystems will be permanently flooded with further estuarine development being prevented by human land-use. Adaptation to prevent damage at this level of sea-level rise is expensive and the risk of it being exceeded for a number of dates in the future needs to be assessed.

The example coastline also has a fringing reef. Buddemeier et al. (1998) state that on the basis of current evidence, atmospheric content of CO_2 of 560 ppm, twice the pre-industrial level, leads to a reduction in the calcification of reefs and benthic calcifying communities of 17–35 per cent.

Consequently, there is a concern that reduced rates of calcification may reduce the ability of the reef to:

- grow fast enough to protect the coastline under sea-level rise; or
- that reduced growth rates may result in too little sediment being produced by the fringing reef, thus reducing the rate of island accretion.

A threshold of 560 ppm of atmospheric CO_2 is chosen on the basis of these reduced rates of calcification. Even though the science surrounding these phenomena is not fully understood, the precautionary principle requires us to protect the environment even under the absence of full scientific certainty. In this case, a significant risk has been identified (Buddemeier et al., 1998).

Therefore, we have identified two related hazards that may combine to place our hypothetical coastline under a greater threat than that posed by each risk on its own, and two chosen thresholds based on an assessment of the impacts. These thresholds are an atmospheric concentration of CO_2 of 560 ppm and a sea-level rise of 50 cm.

The next step is to conduct the risk analysis. Firstly, the range of possible outcomes of the critical variables need to be identified for particular dates in the future. In this case, global sea-level rise will be used in the absence of regional estimates of sea-level rise. Tectonic, isostatic and accretion rates affecting sea level are assumed to be zero but can be taken into account if known for a particular location. The range of global sea-level rise from the IS92a–f scenarios is bound by the IS92c and IS92e scenarios with sea-level estimates from MAGGIC as in Warrick et al. (1996). The range of atmospheric CO_2 is from the same model. Both ranges are assumed to have uniform probability, i.e. the probability of the extremes of the range are just as likely to occur as the median value. The ranges can be seen in Table 11.

Monte Carlo sampling (n=5,000) was carried out within these ranges to determine the probability of both thresholds being exceeded at the same time.

The routine was repeated until the statistical error was <1 per cent.

Three periods where chosen for the analysis: 2050, 2075 and 2100.

The results are shown in Table 12. They show that the thresholds are largely unmet by 2050 but by 2075 the sea-level rise threshold is exceeded in over 20 per cent of cases and atmospheric CO_2 in over 60 per cent of cases for a combined risk in 16 per cent of cases. By 2100, the combined thresholds are exceeded in over 40 per cent of cases. The lower limit actually increases under the Kyoto Protocol due to the emission scenario IS92c having lower emissions for Annex I countries than is required by the Kyoto Protocol (see Table 4).

When the Kyoto Protocol modified IS92a–f scenarios are applied, the decrease in risk ranges between 0 and 15 per cent. The risk of both thresholds being exceeded drops by 9 per cent and 6 per cent in 2075 and 2100 respectively.

Note that the risk of exceeding a threshold does not always decrease. For instance, the CO_2 threshold in 2100 remains unchanged despite a reduction in the range of uncertainty. This is due to the range of possible atmospheric CO_2 concentrations reducing at its upper limit and increasing at its lower limit, leaving the overall distribution of uncertainty about the threshold unchanged.

For a climate change risk assessment, there are two principal forms of management available to reduce risks once analysed: mitigation and adaptation. Mitigation reduces both the median value and the uncertainty range associated with climate change, thereby reducing the risk profile.

	Sea level rise (cm)		**Atmospheric CO_2 (ppm)**	
	low	*high*	*low*	*high*
IS92a—f				
2050	6	40	450	565
2075	10	64	465	720
2100	13	94	470	955
IS92a—f Kyoto Protocol				
2050	7	35	455	515
2075	10	58	480	630
2100	14	82	500	805

Table 11: Ranges of sea-level rise for IS92a–f and IS92a–f Kyoto Protocol modified scenarios

	Sea-level rise (50 cm)	Atmospheric CO_2 (560 ppm)	Combined
IS92a—f			
2050	0	3	0
2075	22	63	16
2100	52	81	44
IS92a—f Kyoto Protocol			
2050	0	0	0
2075	15	48	7
2100	47	81	38
Difference			
2050	0	-3	0
2075	-7	-15	-9
2100	-5	0	-6

Table 12: Risk of threshold exceedence for a sea-level rise of 50 cm, an atmospheric CO_2 content of 560 ppm and their combined risk in percentage for 2050, 2075 and 2100 according to the IS92a–f and IS92a–f Kyoto Protocol modified scenarios.

Where the upper limit of possible climate change is reduced through mitigation, risks will tend to be reduced. The principal instrument of mitigation is the Kyoto Protocol. However, the effect of the Kyoto Protocol on risk as simulated above is limited. Although mitigation in the form of reductions of 5.2 per cent from Annex I countries by 2012 and sustained until 2100 does reduce the risk of individual impacts to climate change, that reduction is small. Mitigation analyses have concluded that substantial quantities of greenhouse emissions are likely to be emitted over the coming century so that the risk of substantial climate change occurring remains high. Therefore, a substantial level of adaptation will be required (Parry et al., 1998; Pittock and Jones, 1999).

4.2.4 Adaptation

Adaptation is the other aspect of managing risk under climate change.

Adaptation reduces the vulnerability of impacts to changing climate, rather than limiting climate change itself. The definition of adaptation in the climate change context is to make more suitable (or fit some purpose) by altering (or modifying) (Smit et al., 1998). Under climate change, we are proposing that risk analysis can be undertaken by analysing the probability of a particular threshold being exceeded as demonstrated above.

How the significance of that risk should be defined (i.e. whether to adapt, or not to adapt) is yet to be determined, although we propose that the decision to adapt should be made jointly by the affected stakeholders and the research teams involved, and that this step becomes part of the analysis (Hennessy et al., 1998; Pittock and Jones, 1999; Walsh et al., in press; Whetton et al., in press).

There are a number of reasons why stakeholders are vital within this framework. Stakeholders will have to provide much of the investment in adaptation, or to seek it from another source (e.g. through grants, investment, legislation etc.). Although the choice of whether to adapt or not may be subjective, it does recognise implicitly that some form of valuation needs to take place, where the costs of not adapting are weighed against the costs and benefits of adapting. While there is a role for formal assessments (e.g. cost-benefit analysis, least cost analysis, decision analysis), some of these processes may take place under less formal conditions, utilising 'expert judgement' or experience, where information has been gathered from the assessment process itself.

The need for stakeholder involvement in adaptation has already been recognised in the South Pacific where measures are being taken to reduce existing risks to coastal damage under current conditions. Without community participation, such measures are unlikely to succeed (Kaluwin and Smith, 1997; Primo, 1997). There are many other benefits to local involvement in adaptation, such as increasing the possibility that adaptations will be culturally appropriate, the opportunity to build on existing practices more readily and the valuation of non-monetary costs and benefits, such as culturally held values (e.g. Nunn, 1997).

It is also possible to set limits for risk by fixing a certain probability, requiring adaptation to the level determined by that probability. Titus and Narayan (1997) recommend a 99 per cent probability level for sea-level rise, i.e. that 99 per cent of possible future outcomes for sea-level rise should be allowed for when planning coastal protection measures. This is a more classical engineering approach.

In the framework presented here, hazardous impacts are linked to climate through the use of a critical threshold. Adaptation would be required where such a threshold has a significant probability of being met under climate change. Significance would be decided by the stakeholders in co-operation with the researchers as part of the assessment process. The gap between the current situation and the critical threshold then becomes the window for adaptation. In the example given in the previous section, the two thresholds under assessment may be exceeded sometime between 2050 and 2075, becoming probable by 2100. The window for adaptation therefore becomes the period between now and then, giving stakeholders an idea of the scale and timing required to adapt.

Adaptation can also be applied in less defined circumstances where a degree of potential loss can be identified and adaptation offers shorter term benefits independent of climate change, the so-called 'win-win' situation.

Examples in the SPREP region are coastal protection where there is an existing risk, increasing output and reducing the variability of agricultural yield and ensuring a regular water supply.

The proposed risk assessment framework allows a scientific assessment framework to be integrated with risk management in the form of stakeholder-driven adaptations. As such, it has the potential to be integrated into existing adaptation programmes in the South Pacific such as integrated coastal zone management (see Kaluwin and Smith, 1997).

5. Synthesis and recommendations

5.1 Synthesis

The aim of this study is to survey the effect of the Kyoto Protocol and construct regional climate change scenarios for the South Pacific. The brief for this study from the South Pacific Regional Environmental Programme (SPREP) is as follows:

- Study the implications of the Kyoto Protocol (i.e., at least 5.2 per cent reduction of greenhouse gas (GHG) emissions below 1990 levels by Annex I Parties in the commitment period 2008–2012) and develop regional scenarios of the impacts of climate change in the Pacific island countries (PICs), including GHG (i.e. six main gases) emissions, and provide emission scenarios for the most important GHGs for time horizons of 2050 and 2100.

- Using information from relevant global climate models (GCM) and appropriate to the Pacific islands conditions, develop climate change scenarios from emission reduction commitments for the following:

  – energy-balance/upwelling climate models;

  – important GHG emission scenarios including sulphate aerosol and forcing;

  – regional mean surface and ocean temperature;

  – precipitation in the region (patterns and intensity);

  – sea-level rise trends;

  – El Niño and tropical cyclone linkages to global warming/climate change; and

  – impact of climate change and variability to water, coastal areas, agriculture, forest and health.

This project has also been integrated into the Pacific Island Climate Change Assistance Programme (PICCAP), through the preparation of scenarios of temperature and rainfall for four regions covering Micronesia, Melanesia, and north and south Polynesia for inclusion into the PACCLIM climate scenario generator. CSIRO Atmospheric Research has contributed additional resources so that a comprehensive set of scenarios consistent with the above brief can be developed.

This synthesis combines the information in the report to provide regional scenarios for the variables listed above. Quantitative regional scenarios are provided for changes to average temperature and half-yearly rainfall.

For a number of climate variables, the lack of scientific knowledge allows only qualitative conclusions to be made. A section on risk assessment is also included, as this is a major strategy being developed by CSIRO Atmospheric Research to manage the uncertainty associated with scenario construction and impact assessment.

The results presented here have varying degrees of confidence:

- Low confidence indicates that modelling has produced an impact but the result is either preliminary, i.e. it has not been widely tested, or is only seen in a limited number of simulations. Much caution is warranted when making projections based on these results.

- Moderate confidence indicates that the results have been reproduced a number of times in studies and/or are considered to be well grounded in theory. Moderate confidence is equivalent to an ‘each way’ bet.

- High confidence indicates that the results are considered to be robust. They have been found in a number of studies and occur under a broad range of conditions within model simulations. These outcomes are considered more probable than not.

No change is assumed where there is no information given due to a lack of evidence. Where no change is indicated with a degree of confidence, the status quo is supported by modelling evidence.

5.1.1 Implications of the Kyoto Protocol

The Kyoto Protocol (KP) was studied in detail to determine how it may affect the science of climate change (Section 2.1; Table 1). Several Articles in the protocol allow 1990 greenhouse gas baselines and post-1990 emissions to be altered under specified circumstances. Changes in the mixture of greenhouse gases will also lead to changes in the relative forcings per molecule of those gases, which will alter how they are represented in terms of the equivalent forcing of CO_2 as required by the protocol. These factors will have a very small effect on how climate is represented in simple climate models compared to the effect that mitigation under the Kyoto Protocol has on global warming.

Three greenhouse gas emission scenarios IS92a, c and e were modified according to the Kyoto Protocol with the reductions of 5.2 per cent for Annex I countries being applied until 2100. Concentrations of the three major greenhouse gases in Annex I country emissions, CO_2, CH_4 and N_2O, were reduced by 2 per cent in 2005, 5 per cent in 2010 and 5.2 per cent thereafter (Table 2). Historical CO_2 emissions and IS92a values for CH_4 and N_2O were used for 1990 and 1995 in all three scenarios.

When applied to the upwelling-diffusion energy-balance model, MAGICC, global warming projections from IPCC (1996) increase slightly for the IS92c (low) scenario, and decrease for the IS92a (mid) and IS92e (high) scenarios (Table 13, based on Table 3). The IS92c temperatures increase because Annex I emissions in this scenario are actually lower than those specified by the Kyoto Protocol. An anomaly is also produced by the highest emission scenario IS92e which shows a lower range for 2050. This is due to the extremely high loading of sulphate aerosols which depresses the temperature in MAGICC, due to the unequal timing of cooling produced by sulphate aerosol (instantaneous) and global warming processes (delayed). However, a reduction in the use of fossil fuel would be expected to also reduce sulphate aerosol emissions, a feedback which has not been introduced at this time.

Global sea-level rise projections from IPCC (1996) increase slightly for the IS92c (low) scenario, decrease slightly for the IS92a (mid) and more substantially for the IS92e (high) scenario (Table 14, based on Table 4).

The IS92c sea levels increase for the same reason as for global temperature.

A similar anomaly exists for the KP modified IS92e results for sea-level rise as occurs for temperature.

For the purposes of forward planning or impact research we advise the use of the existing temperature and sea-level projections of the IPCC (1996) until projections based on new IPCC scenarios become available. This is for the following reasons:

- the changes simulated by the KP modified IS92a–f emission scenarios are small (Tables 13 and 14);
- there are a number of uncertainties surrounding possible feedback effects as sulphate aerosol emissions would reduce with the reduced use of fossil fuels under the Kyoto Protocol;
- the IPCC is currently developing new scenarios to replace the existing IS92a–f scenarios.

Scenario	— 2050 —		— 2100 —	
	KP modified range	Difference	KP modified range	Difference
IS92c	0.6 to 1.2	<+0.1	1.0 to 2.2	+0.1 to +0.2
IS92a	0.6 to 1.2	-0.1	1.3 to 2.7	-0.1 to -0.3
IS92e	0.5 to 1.0	-0.2 to -0.3	1.4 to 2.8	-0.4 to -0.7

Table 13: Projected changes in global temperature (°C) due to the Kyoto Protocol for the IS92a, IS92c and IS92e emission scenarios for 2050 and 2100 (from Table 3)

Scenario	— 2050 —		— 2100 —	
	KP modified range	Difference	KP modified range	Difference
IS92c	6 to 37	<+1	14 to 75	+1 to +3
IS92a	7 to 37	-1 to -2	18 to 82	-2 to -5
IS92e	7 to 35	-2 to -4	19 to 82	-5 to -12

Table 14: Projected changes in sea level (cm) for the IS92a, IS92c and IS92e emission scenarios for 2050 and 2100 (from Table 4)

Therefore, recommended projections are 0.5–1.3°C in 2050 and 0.9–3.5°C in 2100 for temperature. Sea level projections are 6–40 cm in 2050 and 13–94 cm in 2100 (Table 15).

Other possible changes that should also be considered but cannot be quantified include rapid climate change and unexpected outcomes from processes not fully understood.

5.1.2 Regional climate change scenarios

Six coupled ocean-atmosphere climate simulations were included in the analysis of regional climate change scenarios: CSIRO Mark 2 GCM with and without sulphates, CSIRO DARLAM 125 km, DKRZ ECHAM4/OPYC3 GCM, Hadley Centre HADCM2 and the Canadian CGCM1 (Table 5).

Regional scenarios for changes to average temperature and half-yearly rainfall have been created from scaled patterns from these models. The scaling technique is a recent innovation that extracts the relationship between each grid box value and global temperature over a transient simulation to obtain a local greenhouse signal as free as possible of extraneous influences such as multi-decadal variability. These individual model patterns have been aggregated into a range of local change per degree of global warming.

These regional scenarios can be considered as projections that represent a sizeable range of possible future climates. However, the probability of these changes relative to other outcomes remains unknown, so they cannot be regarded as forecasts.

Temperature

A comparison between observations and model simulations shows that the GCMs reproduce spatial temperature patterns reasonably well when model resolution is taken into account (Figure 10). The large grid sizes compared to the size of Pacific island countries mean that model output represents changes in marine air temperature, so issues relevant to detailed island climatology have not been solved. Under climate change, regional warming is mostly below the level of average global warming, as would be expected over the ocean.

Higher latitudes of the southern ocean show the least warming in the SPREP region, while the greatest warming tends to occur in the far west and the central and eastern equatorial Pacific (Figure 11).

Regional scenarios are shown in Table 16, with ranges defined by the lowest local warming per degree of global warming multiplied by the lowest global warming in Table 15 and the highest local warming multiplied by the highest global warming for both 2050 and 2100. A high confidence is attached to this range.

Global warming	*Low*	*Mid*	*High*
2050	0.5°C	0.9°C	1.3°C
2100	0.9°C	2.0°C	3.5°C
Sea-level rise			
2050	6 cm*	20 cm	40 cm
2100	13 cm	49 cm	94 cm

* *Note that in Part One of this project (Jones, in press) the low sensitivity outcome for latent sea-level rise was rejected on the basis of evidence from historical sea-level rise. The whole range is included here, but a sea-level rise in the low to mid range is considered less likely than a rise in the mid to high range.*

Table 15: Global warming and sea-level rise scenarios for 2050 and 2100 (IPCC, 1996)

Region	**Local warming per °C of global warming**	**Warming in 2050**			**Warming in 2100**		
		low	*median*	*high*	*low*	*median*	*high*
Micronesia	0.7 to 1.0	0.4	0.8	1.3	0.6	1.6	3.5
Melanesia	0.7 to 0.9	0.4	0.8	1.2	0.6	1.6	3.2
Polynesia N	0.8 to 1.0	0.4	0.8	1.3	0.7	1.6	3.5
Polynesia S	0.7	0.4	0.7	0.9	0.6	1.4	2.5

Table 16: Scenarios of temperature change (°C) for regions defined by the PICCAP Project (from Table 6)

Rainfall

Present rainfall patterns over the Pacific, mainly defined by the ITCZ and SPCZ, were broadly captured by four of the models, with the other two producing less realistic results (Figures 13 and 14). However, all were deemed satisfactory for further analysis. All models show an increase in rainfall over the central and eastern Pacific over both half years: May to October and November to April. Changes over other regions were smaller, and tend towards increases. Movements of both the ITCZ and SPCZ were noted but these were not consistent between models (Figures 15 and 16). Rainfall invariably increases where warming over the ocean is greatest.

Regional scenarios for rainfall were more difficult to construct, given the sometimes wide range of results. For these reasons, a median value has been included. The CSIRO Mark 2 sulphate simulation is omitted from the calculation of the median as its results are very similar to the CSIRO simulation without sulphates, and two very similar simulations from the same model would unduly weight the results. When the range lies on either side of zero, the outliers are multiplied by the highest global warming for 2050 and 2100, while the median is multiplied by the median warming. Where both low and high extremes are on the same side of zero, the value closest to zero is multiplied by the lowest global warming temperature. The results are rounded to the closest 5 per cent.

The May to October results are shown in Table 16 and the November to April results are shown in Table 17. Outliers can be noted in north Polynesia at the high end and south Polynesia at the low end. The November to April results contain no significant outliers. The median values show that most of the scenarios lean towards increases in rainfall, although the possibility of decreases cannot be ruled out in some regions, particularly during the November to April period. The most notable rainfall change is for north Polynesia during both seasons, with substantial changes in the median result. A moderate confidence is attached to the sign of rainfall change in north Polynesia, and a low confidence is attached to the sign of rainfall changes elsewhere.

Rainfall variability

The following types of rainfall variability were surveyed:

Interannual variability: An analysis of a single GCM showed no increase in rainfall variability between 1960 to 2100. Model variability is much lower than historical variability, because the model does reproduce the magnitude of ENSO fluctuations and the rain is simulated in a grid box rather than at a point. A low confidence is attached to this result.

Multi-decadal variability: Multi-decadal variations are produced by the CSIRO and other GCMs. Analysis of their behaviour under climate change has yet to be conducted. However, methods need to be developed so that they can be explicitly incorporated into climate scenarios.

Region	Response per °C of global warming			Change in 2050			Change in 2100		
	low	*median*	*high*	*low*	*median*	*high*	*low*	*median*	*high*
Micronesia	0	6	10	0	5	15	0	10	35
Melanesia	-4	2	6	-5	0	10	-15	5	20
Polynesia N	4	13	43	0	10	55	5	25	150
Polynesia S	-8	1	2	-10	0	5	-30	0	5

Table 17: Scenarios of rainfall change (%) for regions defined by the PICCAP Project for May to October, rounded to the nearest 5% for 2050 and 2100 (From Table 7)

Region	Response per °C of global warming			Change in 2050			Change in 2100		
	low	*median*	*high*	*low*	*median*	*high*	*low*	*median*	*high*
Micronesia	-2	1	4	-5	0	5	-5	0	15
Melanesia	-3	2	6	-5	0	10	-10	5	20
Polynesia N	6	14	20	10	15	26	20	30	70
Polynesia S	-6	1	3	-10	0	5	-20	0	10

Table 18: Scenarios of rainfall change (%) for regions defined by the PICCAP Project for November to April, rounded to the nearest 5% for 2050 and 2100 (From Table 8)

Rainfall intensity: Increases in daily rainfall intensity are expected in many regions where rainfall increases, remains the same or decreases slightly. Even with an appreciable decrease in average rainfall, reductions in daily rainfall intensity can be negligible (Figure 19). The results concerning rainfall intensity come from a number of models and studies, so have a high confidence attached.

Sea-level rise

Historical sea-level rise over the Pacific from tide gauge records adjusted for post-glacial rebound is consistent with global estimates. Due to the preliminary nature of studies of regionally varying sea-level change, it is recommended that projections of global sea-level rise from IPCC (1996) continue to be used in impact and adaptation assessments for the time being.

ENSO

ENSO is the dominant influence on climate variability in the Pacific, producing large oscillations in temperature, winds, rainfall, sea level, surface pressure and a number of other variables. Therefore, knowledge of how ENSO changes under climate change is a key to understanding climate change impacts in the Pacific region. Analysis of model output shows that the ENSO phenomenon is likely to continue out to 2100. Patterns and analyses of temperature and rainfall produced for this study show that the GCMs produce a more El Niño-like mean state over the Pacific under climate change. Rainfall increases are also distributed in an El Niño-like pattern but they generally increase over most of the Pacific, except where warming is least. These analyses need to extend from the NINO3 and NINO4 regions which are ±5° from the equator, to the whole of the Pacific to obtain a broader picture of how ENSO may change.

Tropical cyclones

Numbers: There is no evidence that TC numbers may change.

Intensities: A general increase in TC intensity, expressed as possible increases in wind speed and central pressure of 10–20 per cent at the time of CO_2 doubling now appears likely. How this affects the risk posed by severe storms needs to be determined on a regional basis. A moderate confidence is attached to this result.

Regions of formation: No significant changes in regions of formation were noted in the DARLAM 125 km resolution simulation, although it is possible they may change in response to long-term changes to ENSO.

Regions of occurrence: There appear to be no major changes in regions of occurrence except to note that TCs may track further polewards. A low confidence is attached to this result.

5.1.3 Impacts and risk

Impacts

This summary is restricted to impacts that may be directly affected by the outcomes of the previous section. A more extensive review is in Section 4.1.

Coastal areas will continue to be affected by ENSO variability, tropical cyclones and wave climates. ENSO appears likely to continue, although how it may manifest remains uncertain. Tropical cyclones may become 10–20 per cent more intense at $2\times CO_2$ occurring roughly between 2030 and 2060 which would be likely to increase storm surge heights. There is no information about how wave climatologies may or may not change on a seasonal basis. Sea-level rise is additive to ENSO sea-level variability, surge and swells. Therefore current risks are likely to persist and probably increase at least at the rate determined by sea-level rise.

Over the past 12 months, scientific understanding of the status of coral reefs under climate change has altered. There is evidence of a likely reduction in the calcification rates of benthic calcareous organisms (coral and algae) due to the increased acidity of sea water under higher CO_2 levels in the atmosphere. This has changed the attitude of marine researchers from one of assuming reefs were generally fairly resilient under climate change to one of increased concern (Done, pers. comm.). Calcification may reduce the rate of reef growth, strength, or both, making reefs less resilient to sea-level rise and reducing the sediment supply to beaches and islands.

Together with threats from turbid and nutrient run-off, coral bleaching and overuse, coral reefs are now considered to be more vulnerable than previously believed.

Intense rainfall on a daily, monthly and seasonal basis may increase in the north Polynesian region and further east, in conjunction with large increases in average rainfall. Similar but smaller changes may also occur elsewhere. A positive outcome would be an increase in water supply due to higher recharge in the relevant regions, particularly where groundwater lenses are important. Increases in extreme rainfall frequency would lead to more flooding. This may affect:

- planning of drainage works;
- zoning of industrial and residential areas;
- dam design;
- agricultural risk management;
- sewage disposal;
- terrestrial sediment and nutrient pollution of reefs;
- land degradation; and
- transport and telecommunication infrastructure.

Risk

A risk assessment framework has been introduced and an example of risk analysis is carried out. The framework involves a focus on impact outcomes, through the construction of thresholds. This requires input from stakeholders, i.e. from those who manage and use the resources impacted upon. Stakeholders are also central to the introduction and analysis of adaptation options. The framework proposes a scientifically based risk assessment process for climate change that can be integrated with existing adaptation programmes in the Pacific such as integrated coastal zone management.

In the risk analysis (Section 4.2.3), two thresholds for a hypothetical coastline were introduced: a 50 cm sea-level rise, and an atmospheric concentration of CO_2 of 560 ppm, associated with a possible reduction of calcification rates in reef communities. When analysed for combined risk under the IS92a–f scenarios and KP-modified scenarios derived in Section 2.1, in 2075 and 2100, the risk under the Kyoto Protocol was reduced by 9 per cent and 6 per cent respectively. This shows that for these two impacts, implementing the Kyoto Protocol reduces the risk by <10 per cent, or alternatively, delays it by less than a decade. The results are shown in Table 19.

5.2 Recommendations

A number of recommendations can be made arising from issues raised in this report. We have restricted these recommendations to the major issues affecting the modelling of climate and climate impacts in the Pacific, where knowledge is sparse, or where there is an opportunity to significantly reduce uncertainty.

5.2.1 The behaviour of ENSO under climate change

The behaviour of ENSO under climate change will affect Pacific island countries through oscillations in sea level, rainfall, temperature, winds and tropical cyclone behaviour. One of the most critical aspects of modelling ENSO is model resolution. CSIRO Mark 3 GCM, scheduled for operation later this year will run at a higher resolution than the GCMs used in this study, providing the opportunity to capture ENSO and major Pacific climate features much more realistically.

The following specific recommendations are made:

- undertake climate change simulations using a finer resolution GCM in order to better capture ENSO and other large-scale features such as the ITCZ and SPCZ;
- simulate ENSO using GCMs in order to capture ENSO and its related features under current climate. This is required to validate the climate change simulations;
- carry out fine-resolution nesting of regional climate models using one or more of these techniques:
  - one-way nesting in CSIRO Mark 3 at 60 or 30 km resolution;
  - testing and implementing two-way nesting (a regional climate model running jointly with a GCM);
  - a stretch-grid approach, where the finest resolution is over the area of greatest interest; and

Year	IS92a—f	KP modified IS92a—f	Difference
2075	16	7	-9
2100	44	38	-6

Table 19: Risk of threshold exceedence (%) for the combined risk of a sea-level rise of 50 cm and an atmospheric CO_2 content of 560 ppm 2075 and 2100 according to the IS92a–f and IS92a–f Kyoto Protocol modified scenarios

- further analyse existing model runs to try and better understand the El Niño-like state observed in the models analysed in this report. Analyse important variables across the Pacific at latitudes extending beyond the NINO3 and NINO4 regions.

5.2.2 Regional sea-level rise

The South Pacific Sea Level and Climate Monitoring Project is gathering valuable data from the Pacific, although it will be some years before any useful trends in sea level could emerge. There is a need to continue the work on global and regional sea-level rise currently being carried out by CSIRO Marine Research. Output derived from other GCMs would help to build up a picture of possible regional sea-level rise, allowing the construction of regional scenarios as for rainfall and temperature. However, at the moment there are too few results available to carry out this task.

5.2.3 Development of climate scenarios and projections

The limitations of the IS92a–f scenarios have been noted, although the next generation of emission scenarios becomes available later this year. Concern was also expressed about sulphate aerosol pathways in the IS92a–f scenarios which may be unrealistically high. The potential feedback where reductions in fossil fuel burning reduces sulphate aerosols also needs to be tested, even though the ranges of uncertainty for sulphate aerosol forcing are very high.

The following specific recommendations are made:

- Test the effect of the Kyoto Protocol with different sulphate aerosol scenarios and forcings;

- Prepare projections for new Special Report on Emissions Scenarios (SRES) scenarios as soon as they, and the first climate modelling results using them become available, to replace the current IPCC (1996) projections; and

- Develop probable scenarios for the Pacific region as described in Jones and Pittock (1999).

5.2.4 Extreme events

The following specific recommendations are made:

- analyse high-resolution climate model data for tropical cyclone numbers, intensities, and distribution;

- develop model-based wave climatologies; and

- analyse extreme daily rainfall using high-resolution climate model output.

Storm surge modelling has the potential to develop risk profiles for storm surges. This may be valuable for coastal protection on a location-specific basis under both current and future climate.

5.2.5 Risk assessment

Climate change risk assessment is a relatively new area that has the potential to integrate scientific assessments with stakeholder-driven adaptation. The potential for this framework to introduce scientific assessments into the Pacific region should be studied. The intended outcomes are practical, locally applied adaptation measures suitable for both current conditions and climate change.

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