CLIMATE OF PAPUA

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Climatic Setting of Papua

Papua is the western half of an equatorial island that is the northern extension of the Australian continental plate, together forming a barrier that blocks the flow of surface water from the western Pacific to the Indian Ocean. As a result, surface waters transported across the Pacific and so the warmest on the planet pile up in the western Pacific north of Papua as the vast Western Pacific Warm Pool (WPWP) (Figure 2.3.1). The WPWP is the single largest heat source to the global atmospheric circulation on the planet. To the south of Papua is the shallow epicontinental Arafura Sea which permits only slight water transfer between the oceans and also has high surface temperatures. Although no point in Papua is more than 250 km from the sea, the island is effectively divided in two by a ESE-WNW trending mountain cordillera that exceeds 3,500 m above sea level (asl) in Papua and reaches 5,000 m asl in the highest peak, Mt. Jaya. This pattern is repeated in the west by a lower chain of mountains in the Vogelkop peninsula.

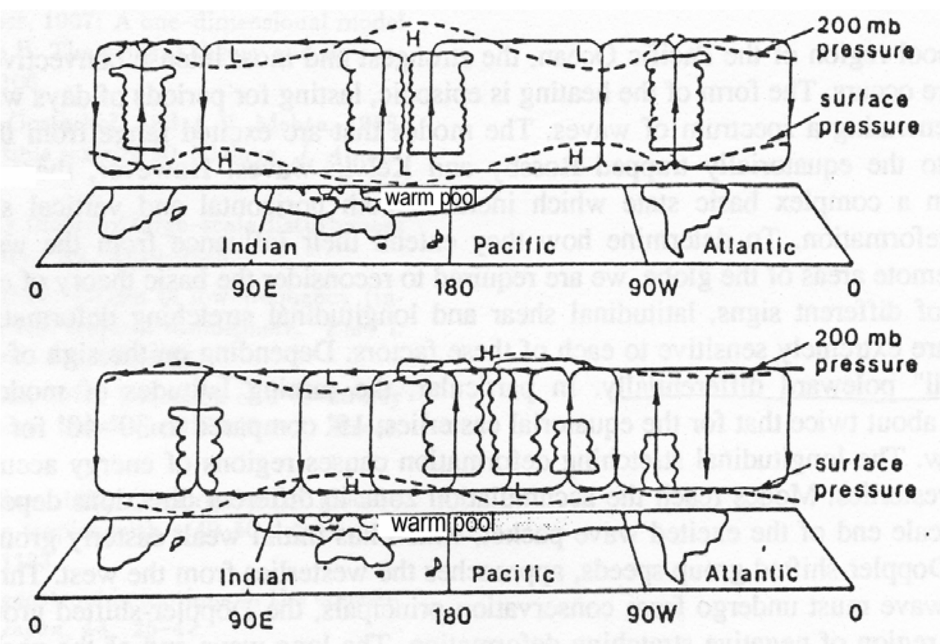

Fig. 2.3.1 Schematic cross-section of the zonal circulation on the equator during La Niña (above) and El Niño (below) phases of the El Niño-Southern Oscillation. (after Webster and Lucas, 1992).

The atmosphere and seasonal changes.

The large-scale atmospheric and oceanic controls on the climate of Papua are discussed by MacAlpine et al. (1983), Nix and Kalma (1972), Hastenrath (1991), and Philander (1990). Three major circulation systems control the weather of Papua. The meridional Hadley Circulation consists of equatorward air flow at surface, rising air over the warmest sea-surface temperatures (SST), poleward flow in the free troposphere, and subsidence on the subtropical high-pressure centers. The zonal Walker Circulation is also thermally driven but by the contrast in SST across the tropical Pacific (Figure 2.3.1). In the western Pacific, these circulations are abstractions as ascent related to the Hadley and Walker Circulations are indistinguishable. Less well known are the oblique, large-scale circulation systems associated with the semi-permanent troughs in the mid-latitude westerlies, the so-called polar troughs. On the western sides of these troughs, surface air is steered equatorward while, on the eastern sides, equatorial air is driven poleward.

These three circulation systems generate two important zones of surface air convergence. The northern and southern hemisphere cells of the Hadley Circulation meet at the Intertropical Convergence Zone (ITCZ) where deep convection is fueled by low-level convergence of moist air. Although narrow in the eastern Pacific, the ITCZ or equatorial trough is up to 1,200 km wide in the New Guinea region. The ITCZ migrates ~15^{o} north and south over New Guinea annually following the warmest surface waters. The ITCZ is bordered by northern and southern monsoon shear lines (N/SMSL). The trough in the mid-latitude westerlies that most affects Papua is the South Pacific Convergence Zone (SPCZ).

The surface wind systems affecting Papua are the trade winds, southeast and northeast, and the equatorial or monsoon westerlies. Depending on the direction of the surface pressure gradient between the two monsoon shear lines and their location with respect to the equator, trade winds over Papua are either southeasterly, northeasterly, or northwesterly and of variable depth. During the southern-hemisphere winter, the SMSL shifts northward to coincide with northern New Guinea (equator) and the principal ITCZ convergence is to the north along the NMSL. The result is that New Guinea is completely embedded in the southeast trades, hence the southeast season. During southern summer, the NMSL shifts south to northern New Guinea and principal ITCZ convergence is along the SMSL. In this case, the northeast trades veer at the equator and become the monsoon westerlies over New Guinea (the northwest season). In the transitional seasons, depending on the direction and strength of the surface pressure gradient in the equatorial trough, the predominant winds commonly differ between the northern and southern New Guinea lowlands with the boundary between the two air masses in the highlands.

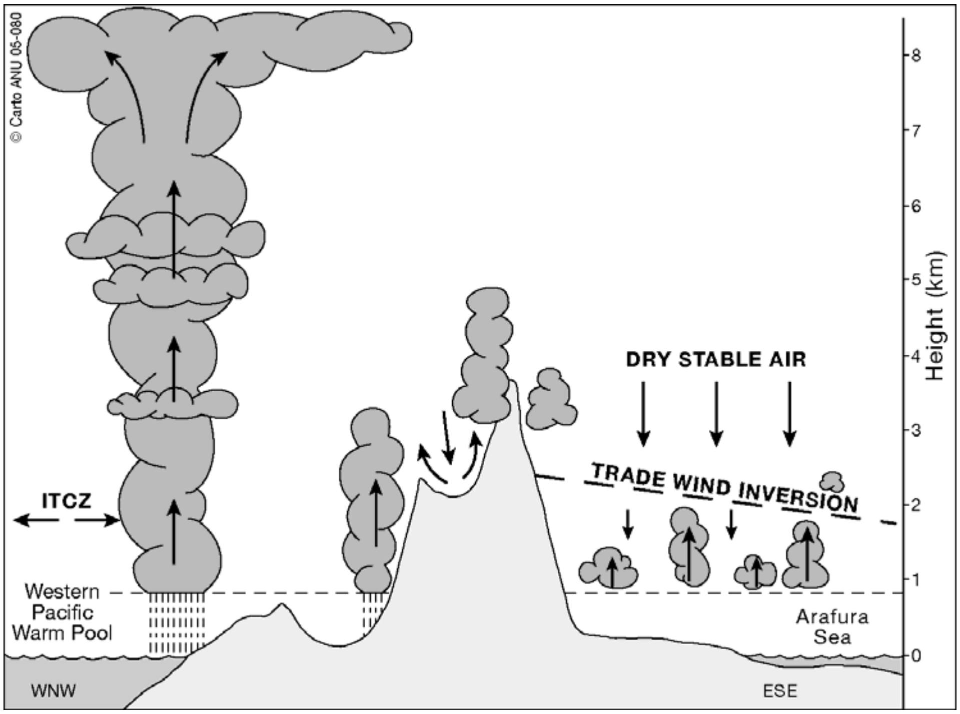

Fig. 2.3.2 A cross section of major climate elements (after MacAlpine, Keig and Falls 1986).

The trade winds and monsoon westerlies are contained with the Atmospheric Boundary Layer (ABL) which is capped by a strong temperature inversion that is at ~2000 m asl but disappears where convection breaks out during daytime along the ITCZ (Figure 2.3.2). This inversion marks the upper limit of the ABL cloud layer. Given their 2000 m stable upper bound, the trade and monsoon winds are commonly blocked by the high central Range with the effect that highland New Guinea climates are dominated by local convection. During daytime, mountain heating generates intense convection and local circulations throughout the year.

The ocean and seasonal changes.

The WPWP exhibits temperatures above 27.5^{o}C, variable salinity (increasing downward), and an average depth of 150 m. It is effectively separated from the cold ocean by the thermocline, a relatively sharp cooling with depth (Figure 2.3.3). In the eastern tropical Pacific, the surface mixed-layer is cooler, on average 25°C with significant seasonal variability, and thinner, with an average thermocline of <50 m (Figure 2.3.1).

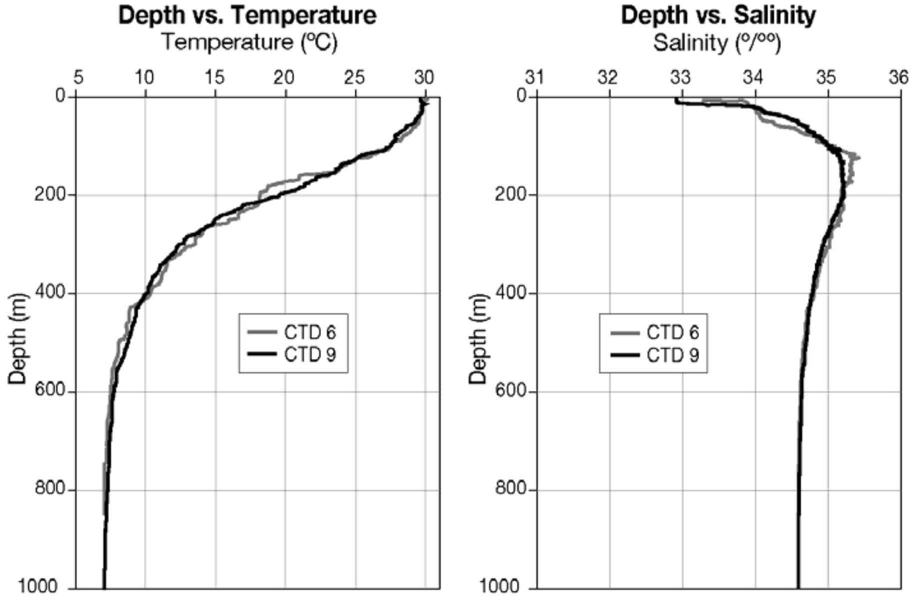

Fig. 2.3.3. Changes in temperature and salinity with depth in the western Pacific. The profiles are from a large bay on the north coast of Papua, Cenderawasih Bay. Two soundings, CTD 6, at the mouth of the bay, and CTD 9, in the centre of the bay, are typical.

The lower SST reflects wind-driven upwelling of cool thermocline water along the equator and also deep cold water along the coast of South America. As a result, eastern Pacific surface waters form a cold tongue which narrows westward. The east-west contrast in Pacific SST is largely responsible for the atmospheric pressure gradient across the Pacific that drives the Walker Circulation (Figure 2.3.1). SST in the central and eastern Pacific is also asymmetrical with respect to the ITCZ or thermal equator which follows the zone of maximum SST between 3 and 10°N.

The Walker Circulation drives the westward-flowing and strong South Equatorial Current (SEC) and the weaker North Equatorial Current (NEC) across the tropical Pacific toward Papua (Figure 2.3.4). Between these two is the North Equatorial Counter Current (NECC) that flows eastward under the influence of equatorial westerlies. Because of the trade winds, the depth of the thermocline decreases from west to east across the tropical Pacific as does the height of the sea surface, the latter by ~50 cm. The zonal slope in the thermocline establishes a pressure gradient in the upper ocean that is directed from west to east and drives a strong subsurface current eastward along the equator within the ocean thermocline, the Equatorial Undercurrent (EUC).

The surface current along the north coast of New Guinea, the New Guinea Coastal Current (NGCC), is strongly influenced by the semi-annual reversal in wind direction (Barmawidjaja et al. 1993; Wyrtki,1961). The NGCC flows to the west during the southern winter, the southeast season, and reverses during northern winter. The NGCC joins the SEC around Halmahera and forms the Halmahera Eddy (HE) (Tomczak and Godfrey 1994; Lukas et al. 1996) (Figure 2.3.4). At a depth of ~200 m, the New Guinea Coastal Undercurrent (NGCUC) flows to the west along the northern New Guinea coast and, after passing the equator, reverses direction and becomes the eastward-flowing EUC. In both hemispheres, surface waters flow away from the equator but recirculate back to it through the ocean thermocline and equatorial upwelling (Dietrich 1970).

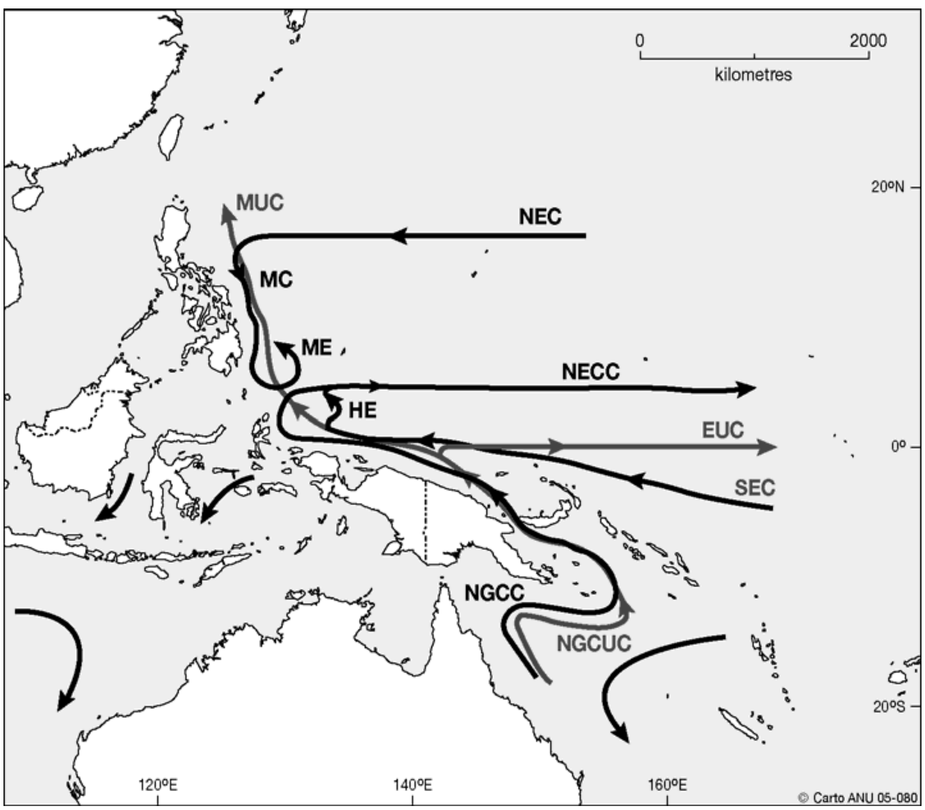

Figure 2.3.4. Climatological schematic of western tropical Pacific currents. SEC: South equatorial current, NEC: North equatorial current, NECC: North equatorial counter current (below 200m), NGCC: New Guinea Coastal Current, NGCUC: New Guinea Coastal Undercurrent.

In Cenderawasih Bay, the surface-mixed layer is 30^{o}C and features a salinity gradient from 33.3 psu (practical salinity units) at the northern end to 30.3 – 32.9 psu in the southern sector (Figure 2.3.3). The depth of this mixed layer is ~40 m in the southern bay and ~60 m in the north. The semi-restricted nature of the basin is clearly indicated by the oxygen content which is low below 200 m depth compared to the open sea. The water mass below 150 m depth has hydrographic properties typical of Western South-Pacific Central Water. The upper water mass has lower salinity and reflects the positive water balance caused by high rainfall and river inflow. This water drains to the ocean and is replaced by cooler, saltier water crossing the 800 m deep sill of the bay.

By contrast, the Arafura Sea has drying winds and a general circulation from east to west during southern winter. At this time, evaporation and clear skies cause high salinities and warming of the shallow water to temperatures over 32^{o} C towards the end of the year. The movement of the ITCZ over the region and development of the Australian Monsoon in December-March reverses the flow and SST falls. Papua is too equatorial to be affected by tropical cyclones (typhoons) which develop south of 6^{o}S but associated rain can affect Papua. To the west of the Arafura shelf, upwelling of cool water from the Indian Ocean occurs off Kaimana, Misool, and the southwestern Vogelkop Peninsula. This leads to more seasonal and lower rainfall. The effect of the different water masses is evidenced by a dramatic vegetational change from a low monsoon forest with winter deciduous elements on Misool Island to tall evergreen rainforest on the northern coast of Salawati Island.

El Niño-Southern Oscillation

The Walker Circulation fluctuates in strength on interannual (3-7 year) timescales resulting in oscillations in surface air pressure over the WPWP that are coherent but out-of-phase with surface pressure over the southeastern tropical Pacific (Philander 1990). This oscillation is known as the Southern Oscillation. When the circulation weakens, surface pressure in the eastern Pacific decreases while, in the western Pacific, surface pressure increases (Figure 2.3.1). The large convective zone over the WPWP shifts eastward which reduces convection in the western Pacific, thereby causing surface pressure to increase. Fluctuations in the Walker Circulation arise due to two-way coupling with changes in tropical Pacific surface and near-surface waters.

The relatively cool eastern tropical Pacific warms on an irregular, interannual period and this causes the trades and, in fact, the entire Walker Circulation to slacken. This permits the WPWP to drain eastwards, further warming the eastern tropical Pacific ocean and atmosphere and slowing the Walker Circulation. These processes lead to a basin-scale warm state referred to as El Niño (Figure 2.3.1). When a sluggish wave in the depth of the ocean thermocline that was launched westward when eastern Pacific SST was cool returns, eastern Pacific SST starts to decrease again which increases the strength of the trades. This, in turn, causes the WPWP to retreat westward, increasing thermocline depth in the west. The relatively cool state of the tropical Pacific is referred to as La Niña. The interannual fluctuations in the coupled Pacific ocean-atmosphere system are referred to as the El Niño-Southern Oscillation or ENSO. Interestingly, ENSO arises due to internal processes without imposed external perturbations.

Radiation and Cloudiness

Solar radiation at the top of the atmosphere is high all year round but the high cloud cover reduces direct radiation at ground level. Areas with daily cycles of rain formation receive the bulk of their radiation in the morning. This leads to aspect differences that might not be expected given the high sun angle experienced through the year. West-facing slopes may receive significantly less radiation than east-facing ones, which can affect crop yields and altitudinal limits. Brookfield (1964) suggested that daily local circulations are responsible for agricultural success in the mountain valleys. Very high cloudiness and waterlogging hinder agriculture on the outer flanks of the mountains but the intermontane valleys are drier and sunnier, supporting dense populations. Heating in the morning causes upslope winds to rise, creating cloud on the slopes. Where a valley is large enough air descends in the center, giving clear skies and abundant sunlight (Figure 2.3.5). In the evening the process is reversed with cold winds descending the slopes and air rising and supporting evening shower activity in the valley center.

Radiation at the top of the atmosphere varies by about 5-10% between November, when the sun is overhead, and July, when it is furthest away. However, cloudiness is lowest during southern winter and increases in the southern summer so the radiation received at the ground varies less than at the top of the atmosphere. Values from Sentani, a relatively sunny location, suggest mean daily solar energy densities of 460-500 milliwatt hours/cm^2 through the year, (McAlpine, Keig, and Falls 1983). Maximum values for Merauke probably match those for Daru

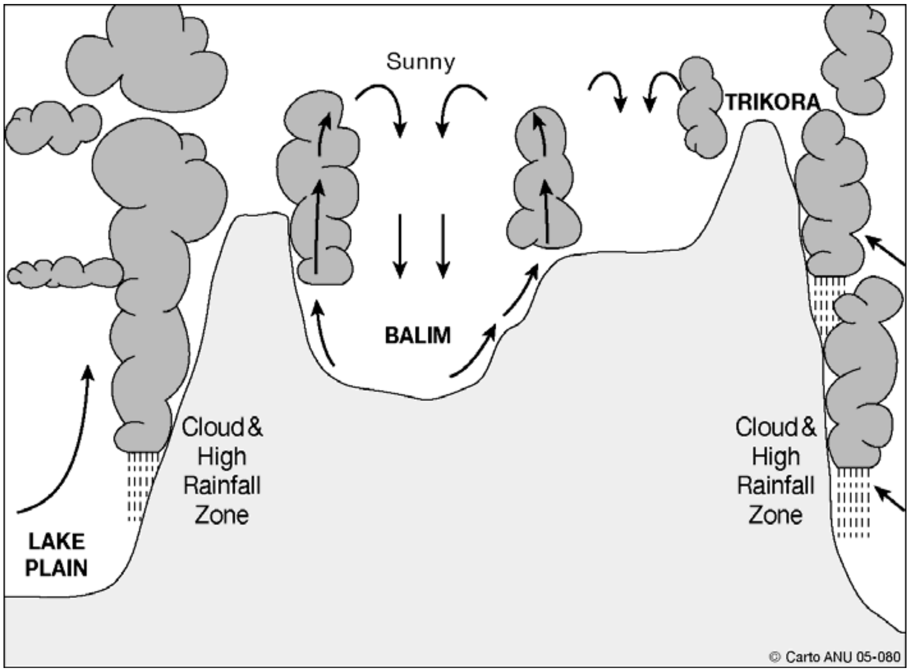

Figure 2.3 5. Daytime circulation in a highland valley. The direction reverses at night.

at 625 milliwatt hours/cm^2. Highland areas are much cloudier and may receive only 450 milliwatt-hours/cm^2. Hence, total inputs of radiation of less than 160 watt-hours/cm^2 per year are found in the mountains compared to annual means of 200 watt hours/cm^2 per year on the north and south coasts. This has consequences for plant growth and evaporation, and may control soil temperature and vegetation limits in the highlands, as suggested by Barry (1978).

Temperature

For New Guinea, meteorological observations, especially in the highlands, are sparse. Therefore, we regard the meteorological data collected at a transect of sites from sea level to 4400 m asl on the south-facing slope of the Central Range at ~137° E and 4° S as representative of the southern half of Papua. This transect of sites is adjacent to the currently glaciated Mt. Jaya massif and is hereafter referred to as the Mt. Jaya transect. These stations show that there is a greater diurnal temperature range below 1500 m asl than above it, particularly in January (Figure

2.3.6). For example, the difference between average January maximum and minimum temperatures is 8-10^{o} C below 700 m asl. However, above 2500 m asl, the average January and July diurnal temperature difference is ~5^{o} C.

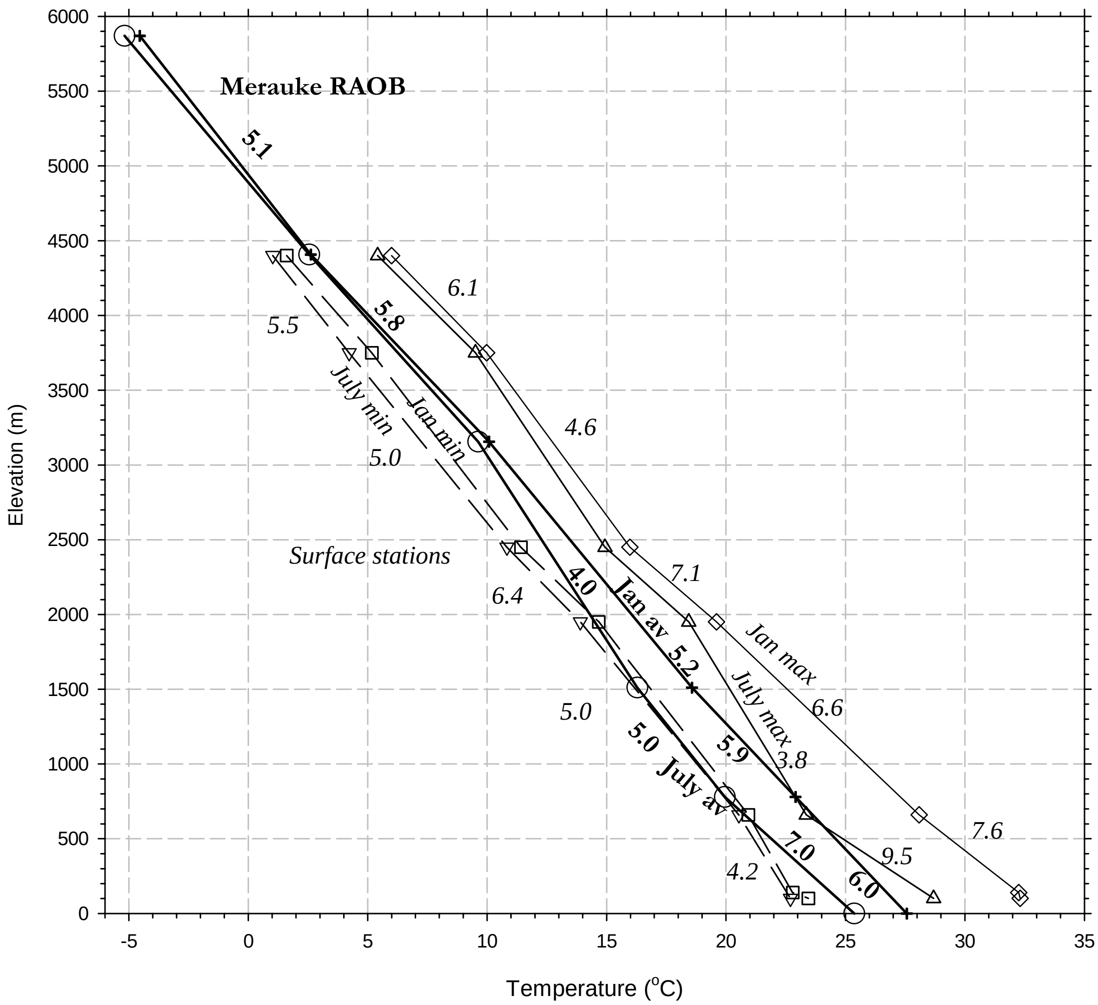

Figure 2.3.6. Average maximum and minimum temperatures for January and July for surface stations on the Mt Jaya transect. Also shown are average January and July temperatures at standard pressure levels (height equivalent) above Merauke recorded by radiosondes launched at 0 GMT. Numbers are lapse rates in ^{o}C/km.

These data also indicate that there is greater seasonality below 2000 m asl than above it (Figure 2.3.6). This seems to be primarily due to a larger difference in average monthly maximum temperatures below 2000 m asl than above. For instance, at 660 m asl, average January and July temperatures differ by 5^{o} C whereas, above 2500 m asl, the difference is about 1^{o} C. January and July minimum temperatures differ by only 1^{o} C across all elevations.

Atmospheric temperatures above Merauke recorded by radiosondes also exhibit increased seasonality below about 2000 m asl (Figure 2.3.6).

The decrease of temperature with altitude above 2500 m asl on the Mt. Jaya transect (surface lapse rate) averages 5.3 ^{o}C/km annually (Figure 2.3.6). This is about the same as the atmospheric lapse rate recorded by radiosondes at Merauke. Surface lapse rates below 2500 m asl are higher, 7^{o} C/km, for maximum monthly temperatures reflecting relatively strong heating of air near the surface during daytime. Surface lapse rates for minimum temperatures are less than the annual average because of radiational cooling at night. Merauke radiosonde data show the same trends. Between 2000 and 2500 m asl, surface lapse rates for maximum monthly temperatures are relatively high. This feature likely reflects the average position of the top of the ABL.

There is strong correspondence between the lapse rate and important biophysical boundaries. Megathermal lowland environments have temperatures of 25-35° C throughout the year with a low diurnal range due to the high heat capacity of saturated air. At about 1,200-1,400 m, the mesothermal zone sees a reduction in tree leaf size and evapotranspiration is reduced. Above this zone, highland climates center around 18° C but have a greater diurnal range from 10-25 °C. Very rare frost can be experienced after periods of dry weather. Above 2,800 m asl, frost becomes more frequent leading to microtherm climates and a change to subalpine forests with microphyll and nanophyll leaves. The treeline, at about 3,900-4,200 m, marks a mean annual isotherm of 6° C and the tropic alpine zone (Barry 1978b, Hnatiuk, Smith and McVean 1970). Daytime temperatures can reach 20° C but, at night, frosts of -2 to -5° C are common. Because humidity remains high, the much colder night-time temperatures of other tropical mountains (-20° C) do not occur (Hastenrath 1991). Due to high snowfall, the snowline occurs at about +1° C at around 4,650 m asl and peaks above this altitude are in the nival zone, where plant life is reduced to snow algae (Peterson and Kol 1976). Snowfalls down to 3,800 m are common but these melt rapidly, usually within a few hours.

Humidity and Rainfall.

Water-vapor mixing ratios (MR) decrease with elevation on the Mt. Jaya transect from 17-20 g/kg at sea level to about 6-10 g/ kg at 4,400 (Fig 2.3.7). The slope of the MR decrease with altitude on the surface transect increases significantly at ~2,000 m asl which we interpret as reflecting the average elevation of the top of the ABL. This is consistent with water vapor content in the atmosphere above Merauke which shows a similar increase in slope between 1500 and 3000 m asl.

On average, there is little seasonality in water-vapor content on the Mt. Jaya transect and what little there is constant with altitude. Water-vapor is at a maximum in January at all elevations. There is greater seasonality in ABL water vapor at Merauke than on the Mt. Jaya transect. The ABL and lower free troposphere at Merauke becomes considerably drier during July than is the case on the Mt. Jaya transect.

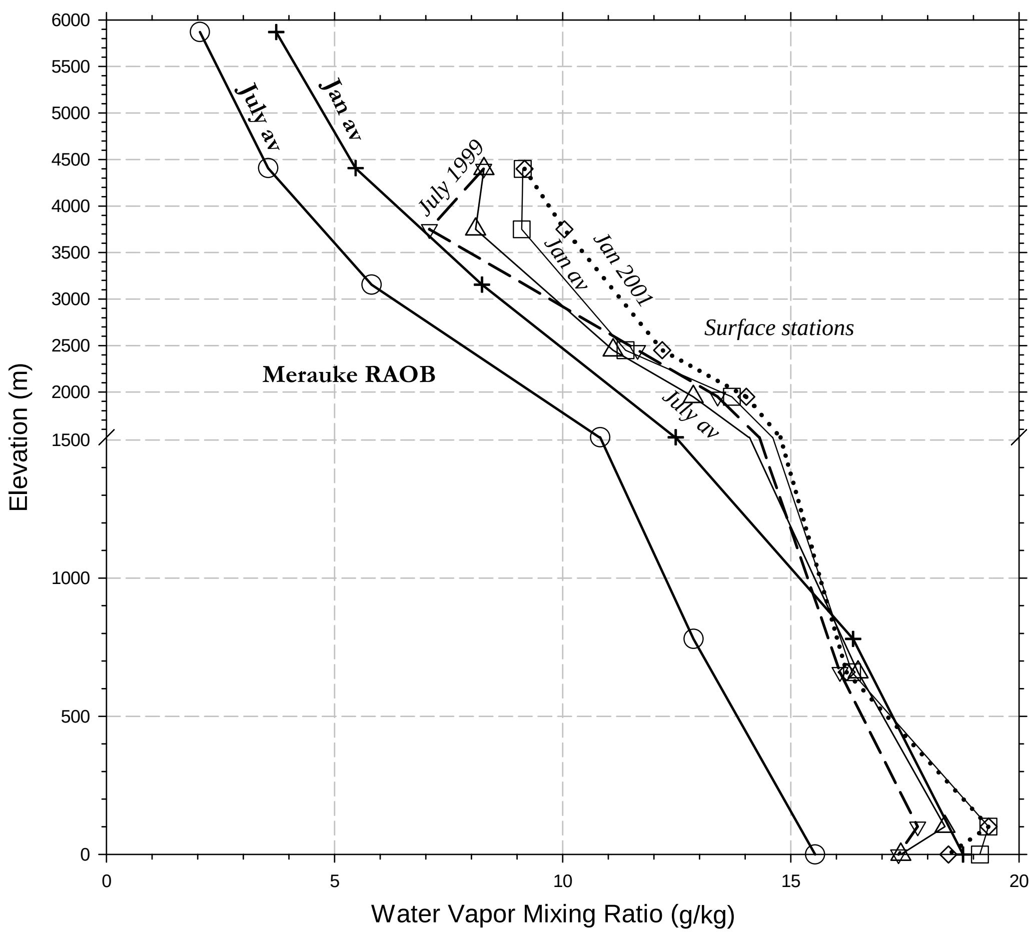

Figure 2.3.7. Average January and July water-vapour mixing ratios for stations on the Mt. Jaya transect (1994-2003, thin solid lines) and also for standard atmospheric pressure levels (height equivalent) above Merauke based on radiosonde observations (+,o RAOB) from 1994-2001. Water vapour profiles are also shown for July 1999 (dashes) and January 2001 (dots) on Mt. Jaya.

Annual rainfall.

Papua is one of the wettest regions on earth. Much of Papua regularly receives 2,500–4,500 mm of rain per year, and a few areas receive over 7,000 mm of rain every year. A map of the provincial pattern of annual rainfall (Figure 2.3.8; Brookfield and Hart 1966) shows that high rainfall occurs along both the north and south sides of the main ranges and around the eastern side of Cenderawasih Bay.

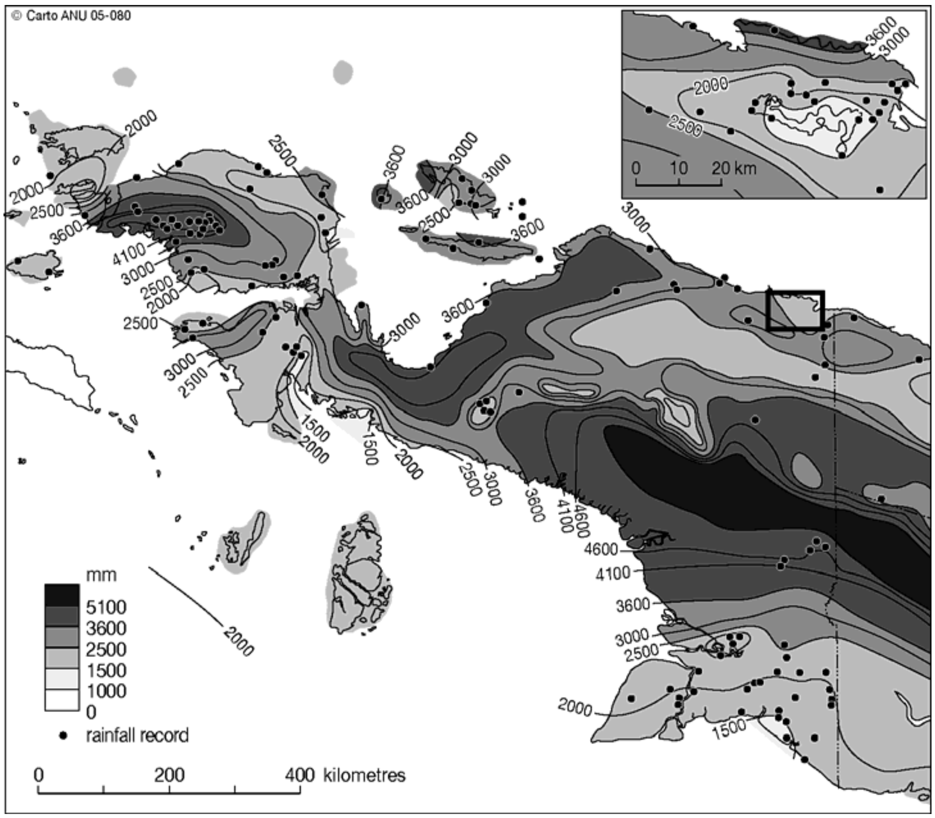

Figure 2.3.8. Annual rainfall for Papua from Brookfield and Hart (1966). Inset shows a rainshadow around Lake Sentani south of the Cyclops Mts.

High rainfall is also received on the south west of the Vogelkop. Very high annual rainfall, over 7,000 mm per year, is received from Ok Tedi in Papua New Guinea, west along the southern flank of the Merauke Range past the Baliem Valley to Timika. No records exist of the highest rainfall point in Papua, but it probably lies at ~1,800 m south of Mt Mandala and has an annual total of ~12,000 mm. Tembagapura at 1950 m asl receives an average of 7000 mm each year. A long-term record of 6,010 mm per year comes from 700 m asl on the southern slopes at Ninati but the mid-altitude southern slopes of the main range are virtually uninhabited. It is not clear why the rainfall in the southeast season diminishes further west because rainfall is also high at this time southwest of the Vogelkop.

Within restricted regions, such as on the south-side of the central range along the Mt. Jaya transect, rainfall increases with increasing altitude (Figure 2.3.9). However, over the whole region, higher annual rainfall is not associated with increasing altitude. The highest alpine areas likely have lower annual rainfall than the mountain flanks. Areas where annual rainfall is below 2,000 mm occur in the Sentani area due to a rainshadow from the Cyclops Mts and the southern third of the Merauke District (Trans-Fly) where the seasonal dry period is longest (Figure 2.3.8). There is also a zone of lower rainfall (< 2,500 mm) in the lowland north of the central range where a rainshadow is developed.

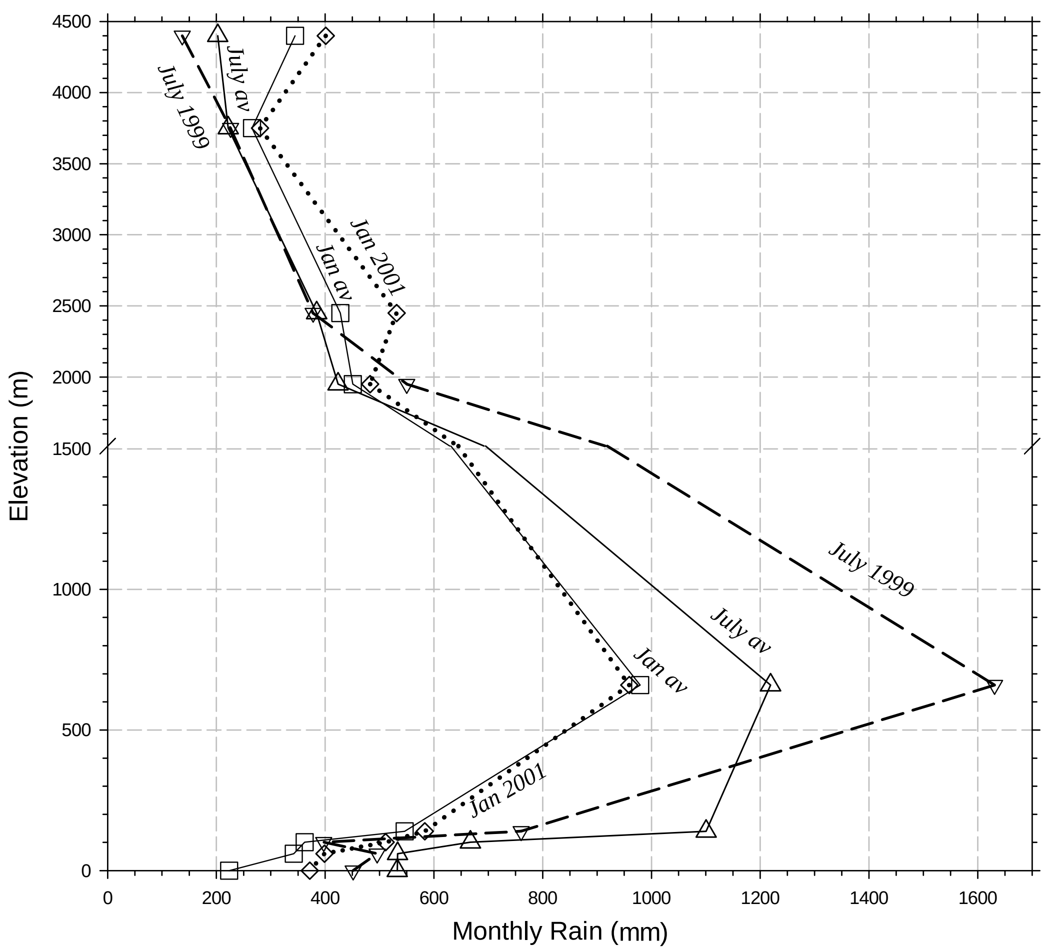

Figure 2.3.9. Average January and July rainfall for stations on the Mt. Jaya transect Rainfall for July 1999 and January 2001 are also shown.

Smaller intermontane basins north of the crest of the main range, such as the Ok Sibil, Baliem, Hitalipa and Wissel lakes areas, are characterized by lower precipitation due to a rainshadow effect. At the Wissel Lakes, the Weyland Mts block moisture from the west while the Merauke Range provides a rainshadow from the east. These upland basins support dense populations. A similar structure provides a drier “neck” southeast of the Arfak Range and east of the Bomberai Peninsula. The driest place in Papua is at Barari in Arguni Bay with a mean annual rainfall of 1,020 mm (Brookfield and Hart 1966). This area is in a rain shadow from the ranges south of Cenderawasih Bay and the Bomberai peninsula.

Rainfall seasonality.

Seasonality is not as marked in Papua as in parts of Papua New Guinea and Maluku because of its near-equatorial setting and position within the WPWP. In many parts of Papua, most rain falls between January and April, the northwest season, with the least falling between May and August, the southeast season. In some parts of the country, this pattern is reversed and most rain is received during the southeast season when the southeast trades strengthen. The southern slopes of the central range provide examples.

Along the Mt. Jaya transect, the phase of the seasonal precipitation maximum changes at ~2000 m asl which is the average elevation of the top of the ABL and upper limit of the southeast trades (Figure 2.3.9). Below 2000 m asl, the precipitation maximum is in the southeast season whereas, above, the precipitation maximum is in the northwest season. Seasonality in precipitation is much reduced above 2000 m asl.

Measured as the relative difference in rainfall between the dry season and the wet season, the most seasonal parts of Papua are the southern half of Merauke regency (Figure 2.3.10). Merauke receives 1,513 mm but no month is completely dry, with falls >100 mm for every month except August.

In many locations, rain is received all year round and there is no seasonal pattern to rainfall amounts. This simply reflects great variability in the sources and intensity of rain. Local, diurnal thunderstorms can provide punishing afternoon and evening showers in winter. Mesocale convective systems bring periods of rain and drizzle for days at a time when the ITCZ moves over the island in southern summer (Tokay and Short, 1999).

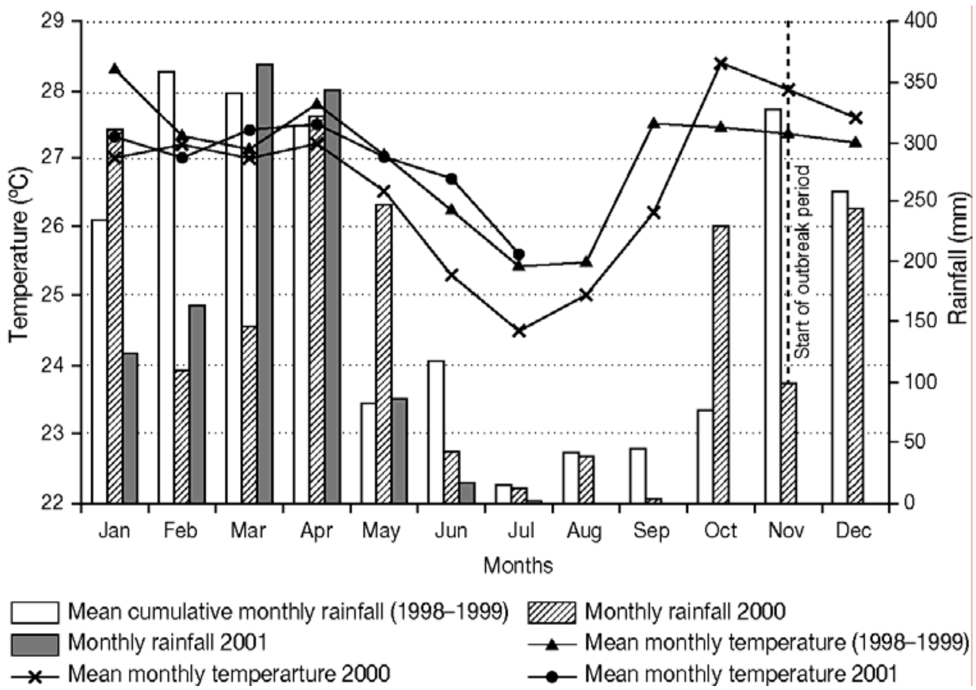

Figure 2.3.10. Monthly temperature and rainfall at Merauke from 1998-2001 (from Sukri et al. 2003).

Interannual Variability in Rainfall

Interannual variation in annual rainfall related to ENSO is low over much of Papua especially in the highlands. Areas where rainfall variability is higher includes the southeastern region near Merauke (MacAlpine et al. 1983). Although variations in annual rainfall related to ENSO are muted, the amplitude of seasonal rainfall can change significantly with ENSO events. For instance, Papua experiences periods of uncharacteristically low rainfall during El Niño warm events of ENSO. These events can seriously disrupt food production in Papua as happened during the 1997-1998 El Niño (Ballard 2000). That drought was associated with heavy frosts at unusually low altitudes and this killed sweet potato and other crops. Fires broke out across the island, burning normally wet forests. Sukri et al. (2003) present data for an extreme dry season at Merauke during the 2002 El Niño when almost no rain fell from June until late November. With the onset of rain, a severe dengue fever attack was experienced.

Soil Water Surpluses and Deficits

The amount of water in the soil available to plants is critical to agriculture. Soil water is measured as "water balance". Different types of soil can hold different amounts of water. Water can be absorbed or lost from different soils at different rates. The amount of water a soil can hold is known as the "field capacity" of a soil. The balance of water in the soil measured in millimeters per day, is the difference between the amount of water entering the soil as rain and the amount of water lost from the soil through evaporation (from the surface), transpired by plants (lost from the leaves) or drained downwards through the soil beyond the reach of agricultural plants. When a soil is capable of absorbing more water (that is, the soil is below field capacity), but no water is supplied by rainfall (or irrigation), the water balance is said to be in *deficit*. When water begins to run off the surface of a soil or is drained beyond the rooting zone of agricultural plants, the soil can absorb no more water and is said to be *saturated* and the soil water balance is said to be in *surplus*. Although different soils absorb and lose water at different rates, the most important determinant of soil water surpluses and deficits is rainfall.

Five patterns of soil water balance can be observed in Papua

- Regular, seasonal, severe soil water deficits occur in the southern part of Merauke regency.
- Irregular, moderate soil water deficits in the central part of Merauke regency, the Sentani-Genyem area, and Misool Island.
- Infrequent, slight soil water deficits over much of the lowlands (below 1,200 m asl).
- Rare deficits with moderate soil water surpluses on the New Guinea mainland at middle altitudes (1,200–1,500 m asl).
- Rare deficits with large soil water surpluses in the highlands.

These patterns of soil water balance have a strong influence on agricultural systems. Where rainfall is high and regular and soils are usually saturated, the digging of drains to remove water from the soil and planting certain crops in mounds to raise their roots above the saturated soil is critical for successful agricultural production. Where regular seasonal soil-water deficits occur, a number of techniques are used to overcome the lack of soil water for part of the year. This is done primarily through the selection of mixes of crops planted. Irrigation in Papua is not common although water management, such as stream diversion, occurs widely. Taro was once a major crop requiring irrigation and the last remnants of such systems are still practiced in a few locations such as the Baliem Valley.

Winds

Winds are generally light and Papua lies north of the zone of destructive tropical storms, although some cyclone tracks approach the south coast with heavy rain reaching land. The southeast trade winds provide a constant light wind of 10-20 km/hr on the southern coast. Light westerly winds occur during the monsoon season. Strong wind is associated with local systems that meet mountain peaks, ascending air being compressed and accelerated. Cold air drainage can also deliver bursts of cold air to valley floors at night.

Powerful thunderstorms with very high rain intensity are common in Papua, due to the extreme height and size of some systems which may exceed 20,000 m and 50 km across. Extreme vertical wind shear occurs in these storms and is a major aviation hazard. Storm fronts are often associated with violent wind gusts that can cause local damage.

Climate and Glacier change:

General modeling suggests that the tropics are experiencing temperature increases both on land and in the sea. This may lead to greater variability in the weather in the future. The subject in the New Guinea region is not well researched but one clear climatic boundary, the extent of glaciers, is well known. The Papuan glaciers are of critical importance because, as proven climate recorders, they are the only glaciers within the vast Indonesian Low pressure system.

Glaciers on Mt. Jaya and elsewhere in Papua (Hope et al.1973) have been shrinking since they were first photographed in 1907 (Ballard et al. 2001) and 1936 (Colijn 1937). Known collectively as the Carstensz Glaciers, the principal Mt. Jaya ice masses of the 1990s consisted of two valley glaciers, the Meren and Carstensz Glaciers in the Meren and Yellow Valleys, respectively, as well as two high-elevation plateau glaciers, the West and East Northwall Firn (Figure 2.3.11). In the first survey of the glaciers in 1936 (Dozy 1938), the West and East Northwall Firn were continuous and the eastern portion of the Northwall Firn provided ice-flow to the West Meren Glacier and so was part of the West Meren Glacier. The east Northwall Firn also fed an eastward-flowing lobe of the Meren Glacier referred to as the East Meren Glacier or the Harrer Glacier. The East Meren Glacier was substantial in the 1930s based on geologic evidence. In the 1950s - 60s, the Northwall Firn separated into western and eastern sections and, by 1987, the East Northwall Firn had effectively separated from the Meren Glacier.

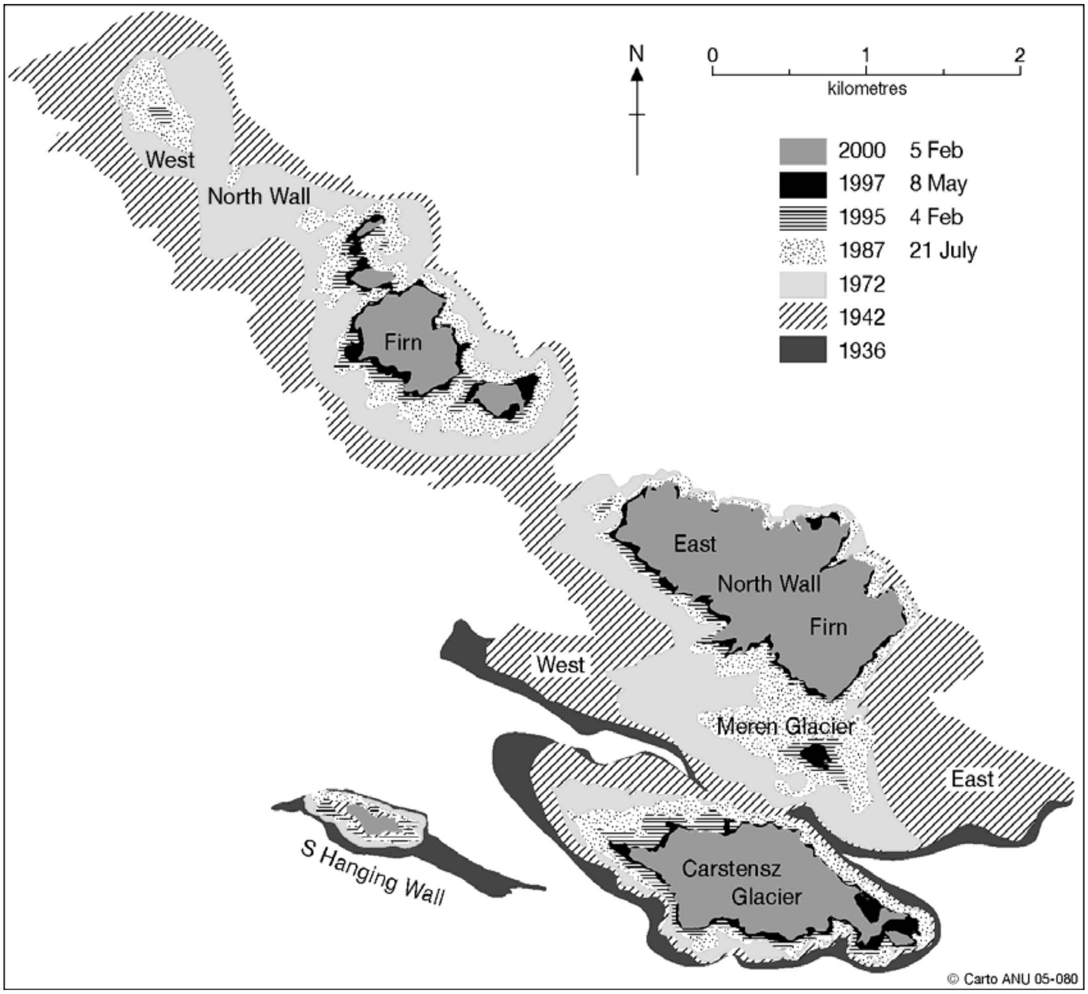

Figure 2.3.11 Map view of the Carstensz Glaciers and changes in their extent since 1936. The 1972, 1942, and 1936 boundaries of the East Meren Glacier are schematic.

The magnitude of recession of the Carstensz Glaciers, its causes, and its implications for local, regional, and global climate change are only qualitatively known (Prentice and Maryuani 2004). Other ice caps occurred at Mt Idenberg, 15 km west of Mt Jaya, on Mt Trikora, and on Mt Mandala. Ice disappeared from Mt Trikora in the 1960’s (Hope, Peterson, and Mitton 1973) and from Mt Idenberg in 1978 leaving a small ice dome on Mt Mandala. These fluctuations have probably been common and the mountains may have been ice free in the early Holocene as no evidence for glaciation is known from this time (Prentice et al. 2005).

Glacier Area and Length Changes

The recession of the Carstensz Glaciers from ~11 km^2 in 1942 to 2.4 km^2 by 2000 represents about an 80% decrease in ice area (Figure 2.3.12). The Carstensz Glacier itself decreased slightly less, ~70%, over this interval. The West and East Meren Glaciers melted away completely between July 1997 and February 1999. Comparison of Meren Glacier area in 1942 to East Northwall Firn area in 2000 yields an 80% decrease in the area of this ice system.

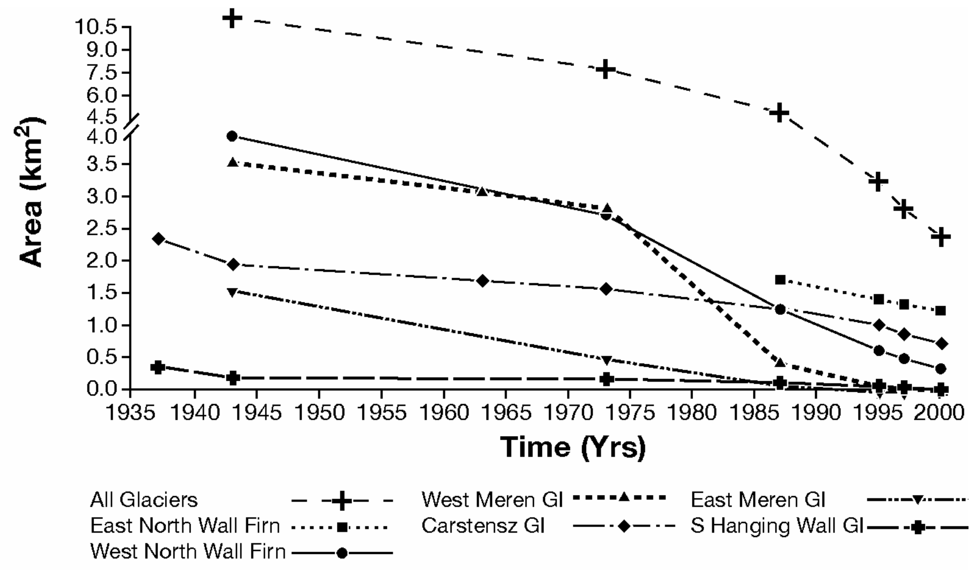

Fig. 2.3.12. Changes in the area of the Mt. Jaya glaciers between 1936 and 2000.

A longer record with more temporal detail is that of the recession of the Carstensz and West Meren Glacier fronts. Between 1936 and 2000, the Carstensz Glacier front receded about 1.2 km, a decrease in length of 46%. Over the same interval, the 2.6 km recession of the West Meren Glacier represents a 100% decrease in length. However, given the separation of the Meren from the East Northwall Firn, it is more reasonable to compare West Meren Glacier length in 1942 to East Northwall Firn length in 2000. The latter comparison yields a 62% decrease in length of the Meren-East Northwall Firn ice system.

Recession rates are useful for resolving the causes of glacier recession. To isolate the impact of climate forcing, we look for the common signal in the recession-rate histories and attempt to factor out the influence of topographic complexity. The dominant pattern of the frontal recession-rate histories for the Carstensz and the West Meren Glaciers appears (this is a low resolution record) to be that the recession rate increased from the 1940s to a peak in 1995-

1996 after which it decreased slightly. On the other hand, the rates of decrease in ice area exhibit considerable discrepancies and so do not support a single climate-forcing hypothesis. The rate of change in Carstensz Glacier area indicates a progressive increase in climate forcing to a peak in 1995-1996. However, the rate in the reduction of area of the other ice masses deviates significantly from the Carstensz pattern.

Glacier Mass Balance and ELA Changes

For 1973-1994, the distribution of the total net snow balance (sum of snow gain and loss) with elevation on the Carstensz and West Meren Glaciers indicates that, for the 22-year period, the balance was zero at ~4,780 m asl. Hence, the altitude of the equilibrium-line (ELA) averaged 4,780 m asl between 1973 and 1994. This is consistent with the ELA that is apparent on the SPOT satellite imagery of July 1987 which is a transient ELA. The average ELA of 4,780 m represents a rise in the ELA of 120 m since 1972 when the ELA was determined to be 4,660 m asl, after correction of the Carstensz Glaciers Expedition topographic map (Hope et al.1976) to the 1995 topographic map.

For the 2-year interval 1995-1996, the total net snow balances, B_n, for the Carstensz Glaciers were determined by photogrammetry to be all quite negative. The two largest glaciers, the Carstensz and East Northwall Firn, had similarly negative budgets which, on an annual basis, were -4,678 x 10^6 kg in water equivalent (average surface lowering of 3.3 m) and -4,739 x 10^6 kg (average surface lowering of 4.5 m), respectively. The distribution of B_n with altitude shows that the average ELA for the two years was above the highest glacier surfaces which were at 4,940 m asl. This balance-based estimate for the ELA contrasts sharply with two different estimates of the ELA based on observation of the transient snow-line. Those observations put the snow-line at 4,650 – 4,750 m asl which is well below our average ELA.

For the period 1997-1999, B_n for the Carstensz Glacier and West Northwall Firn were slightly negative while B_n for the East Northwall Firn was slightly positive. The West Meren Glacier melted away between 1997 and 2000 and so its B_n was negative but not quantifiable. Annual surface lowering of the Carstensz Glacier and the West Northwall Firn averaged 1.3 and 1.6 m/yr, respectively, whereas the surface of the East Northwall Firn gained altitude slightly.

The general glacier retreat between 1972 and 2000 is broadly consistent with the warming trend in the lower free troposphere over western New Guinea over this interval.

Additionally, there appears to be some correlation between interannual specific-humidity variations and glacier fluctuations. By piecing together scattered and discontinuous highland and radiosonde weather observations, we speculate that, at glacier altitudes, mean annual temperature increased ~1^{o} C over this 30 year period. This likely had some effect on the glaciers but, at finer temporal resolution, the relationship is blurred for lack of contemporaneous data. Temperatures at sea-level around western New Guinea increased significantly from a major decadal low in 1992-1993 to highs in 1995/96 and 1996/97 that were the highest since 1988-1989. These highs coincide with the estimates for highly negative mass balances and very high glacier ELA for 1995-1996. In addition to these effects, the frequent intervals of modelled low water-vapor between 1991 and 1995 stand out as a significant series of events that preceded the 1995-1996 negative glacier mass balances. The prolonged near-drought conditions should have negatively impacted the mass balance. From 1997 to 2000, the ELA was probably lowered because mass balances were more positive. This may relate to increased precipitation. Despite this reprieve, the ice is clearly marginal under current conditions of global warming and so may be lost from New Guinea within a few decades.

Discussion

Papua has unusually wet climates in both an Indonesian and global context, with large areas of the mountains too wet and cloudy to support viable agriculture. This moisture buffers the province against dry seasons and maintains thermal equability. The elevation of the high mountains in the late Tertiary set the main climate parameters and the Papuan biota has adapted to the very great thermal range but low local variability since then. Even glacial shifts, while depressing snowline and vegetational boundaries, did not remove the general moisture surpluses except perhaps in the south.

Global warming will similarly be buffered in that rainfall will probably be enhanced, although variability may also increase, with more frequent and severe El Niño and La Niña events possible. This will stress vegetation at boundary positions. The alpine area will be liable to invasion by shrubs and trees but, currently, fire maintains large tracts of open grasslands, so only alpine obligates will be under threat. Likewise, significant changes in alpine lake physical and chemical properties will threaten their flora and fauna. Some retreat of forests and change in forest composition might be expected at the rainforest limit near Merauke. Severe frosts from future El Niño events may kill off frost sensitive species in upper montane forests and limit agriculture.

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