RECONSTRUCTION OF TROPICAL PACIFIC CLIMATE VARIABILITY
FROM PAPUA ICE CORES, INDONESIA

Dissertation

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The glaciers near Puncak Jaya in Papua, Indonesia are the last remaining tropical glaciers in the tropical west Pacific. Three ice cores were recovered from the East Northwall Firn during an ice core drilling expedition in mid-2010. Two ice cores to bedrock measuring 32.13 m (D1) and 31.25 m (D1B) were analyzed for stable isotopes, insoluble dust and major ions concentration to reconstruct the tropical Pacific climate variability. In addition, 1362 rain samples were collected daily within ~2 years from January 2013 to February 2014 and from December 2014 to September 2015 at stations along the altitudinal transect (9 to ~4,000 meter above sea level) on the southern slope of the central mountain ranges in Papua to support the interpretation of ice core records.

The isotope analysis of rainfall demonstrates the nature of altitude effect with a mean isotopic lapse rate for δ¹⁸O (δD) of -2.4‰/km (-18.2‰/km). The results suggest that regional convective activity is a more important factor controlling δ¹⁸O variability of rainfall on daily to interannual timescales, than precipitation amount and local surface temperature. Furthermore, the seasonal convection effect on rainfall δ¹⁸O is interpreted as the temperature (at mean condensation level) effect with a depleted (enriched) rainfall δ¹⁸O during summer (winter) in association with enhanced (suppressed) convection such that precipitation is generated at a higher (lower) mean condensation level with cooler (warmer) corresponding temperatures. In addition, rainfall δ¹⁸O (δD) values at all stations
are generally more enriched by 1.6‰ - 2‰ (11‰ - 15‰), and $d$ values are lower at lower stations during El Niño than during the normal periods.

The age of the D1 (D1B) core is ~90 years (~77 years) with a bottom age of 1920 (1933). The ice core dating is based on tritium analysis which provides an absolute time marker of 1964 at a depth of 23.4 m in Core D1, and $\delta^{18}O$ reference-matching with NINO3.4 sea surface temperature (SST) and additional support from the dust and chemistry profiles. On decadal to interdecadal timescales, dust and major ions concentration in the ice cores are modulated by El Niño-Southern Oscillation (ENSO) activity and Pacific Decadal Oscillation (PDO) phases. On interannual timescale peaks of dust, ammonium and potassium are associated with El Niño-linked drought and biomass burnings.

The D1 $\delta^{18}O$ time series shows an increasing trend from 1920 to 2010 with a slope of 0.012‰ per year and has a significant positive correlation with the 550-mb air temperature anomalies along the tropical bands between 15°S and 15°N. The identical increasing trend of the annual D1 $\delta^{18}O$ and the annual mean temperatures at the 550-mb level and lowland stations over Papua region suggest that atmospheric warming controls $\delta^{18}O$ variation. In contrast, annual precipitation rate shows contemporaneous positive trend that contradicts the amount effect as a major factor.

Radiosonde data indicate that the freezing level (0 °C isotherm) altitude (FLA) has increased (decreased) during El Niño (La Niña) events which are associated with warmer (cooler) conditions at the glacier site. During El Niño, precipitation $^{18}O$ is enriched, aerosol deposition on the glacier surface is enhanced due to the exposure of
more dust sources, and accumulation is low with most precipitation falling as rain rather than snow, particularly in austral winter. During La Niña, precipitation $^{18}$O is depleted, higher precipitation and cooler temperatures prevail in the highland, particularly in austral winter, which results in higher snow accumulation.

The comparison of the Papua ice core records with coral records in the vicinity of Papua, indicates that $\delta^{18}$O in the Papua ice cores is positively correlated with most of the coral $\delta^{18}$O in the Northwest, West and Southwest Pacific Ocean and Indonesian seas, and negatively correlated with coral $\delta^{18}$O in the Central Pacific.

Compilation of previous studies indicates that the total ice area near Puncak Jaya has decreased at a rate of ~0.15 km$^2$ per year since ~1850, which implies that these glaciers will likely disappear by ~2017-18. This is consistent with personal observations made during the 2010 ice core drilling campaign where ice around a tent at the campsite had melted a 30 cm after ~3 weeks camping on the ice field, suggesting a thinning rate of ~5.2 m/yr. A recent stake measurement on the ice surface indicates that the ice thickness has thinned by ~5.26 m or ~1.05 m/yr between 2010 and 2015, which provides the upper bound of the date for the glaciers disappearance by ~2040. The very recent separation of the two ice masses in the East Northwall Firn suggests that the south ice front has experienced an accelerated retreat rate from ~14 m/yr between 1936 and 2006 to ~51 m/yr between 2006 and 2015 which is most likely due to atmospheric warming.

Keywords: Puncak Jaya; Papua; Indonesia; paleoclimate; precipitation; stable isotopes; tropics; ice cores; glacier retreat; monsoon; El Niño-Southern Oscillation
DEDICATION

I dedicate this dissertation to all of my family, especially my beloved wife, son and daughter for their patience, understanding and encouragement during my study. I also dedicate this dissertation to my colleagues in the Indonesian Agency for Meteorology, Climatology and Geophysics (BMKG), particularly the research and development center for supporting my study, as well as in Byrd Polar and Climate Research Center (BPCRC), particularly the ice core paleoclimatology research group (ICPRG) for their support, suggestions and encouragement. I am extremely grateful for what I have achieved today.
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CHAPTER 1
INTRODUCTION

Tropical glaciers and ice caps are highly sensitive indicators and recorders of climate changes. Small glaciers in the Tropics respond rapidly to climatic changes (Francou et al., 2003). Records of past climates and environmental conditions are preserved in the layers of these glaciers and ice caps. Ice cores drilled through these frozen archives allow us to investigate past climatic variations and understand the mechanisms that control these changes (Thompson et al., 1984). Locked in these glaciers is an incredible wealth of information on climatic and environmental changes associated with El Niño-Southern Oscillation (ENSO) variations through time, as well as the monsoon variability. These high-elevation, low-latitude ice fields exist within a fairly stable but sensitive environment where mid-tropospheric temperatures vary within a relatively narrow range.

The warmest sea surface and upper level atmospheric temperatures occur in the Tropics where the associated energy drives intense and deep convective precipitation and is important for the evolution of coupled ocean/atmosphere phenomena such as ENSO and Asian-Australian monsoon system (Fig 1.1) (Thompson et al., 2011). Climate variability in the tropical Pacific is dominated on interannual time scales by ENSO, involving the interplay between central and eastern equatorial Pacific sea surface
temperatures (SSTs), and the atmospheric pressure gradient between the western and eastern Pacific Ocean. ENSO is responsible for meteorological phenomena (e.g., floods, droughts, temperature extremes, Atlantic and Pacific tropical storm intensities, etc.) that directly or indirectly affect much of the tropical and extratropical regions and their populations.

Unfortunately, most of the tropical glaciers are currently in retreat, as observed in the tropical Andes (Thompson et al., 1979; 1984; 1984a; 1985; 1994; 1995; 1998; 2011; 2013; Vimeux et al., 2009), in the tropical Africa (Hastenrath and Kruss, 1992; Kaser and Noggler, 1991; Taylor et al., 2006; Thompson et al., 2002; 2009; 2011a), and on small glaciers near Puncak Jaya in Papua province, Indonesia (known as Irian Jaya before 2002 in New Guinea)(Klein and Kincaid, 2006; Kincaid, 2007; Prentice and Hope, 2007; Prentice and Glidden, 2010). Therefore, ice cores recovered from these high-elevations, low-latitude glaciers are important archives from which we can extract paleoclimate information about the Tropics before they soon disappear.

On the eastern side of the Pacific Ocean, climate history records and reconstruction of climate variability based on tropical ice cores have been relatively well studied in the Andes (Thompson et al., 1979; 1984; 1985; 1994; 1995; 1998; 2011; 2013; Vimeux et al., 2009). ENSO events have been recorded in ice cores from the Quelccaya ice cap, Peru (Thompson, 1993; Thompson et al., 1984a; 2011; 2013), Nevado Huascarán, Peru (Henderson et al., 1999) and cores drilled on Nevado Hualcàn in 2009 (unpublished). On the western side, the 2010 ice cores from glaciated peaks near Puncak Jaya in Indonesia (Permana, 2011; Thompson et al., 2011) provide information on the
effects of ENSO on these glaciers and are useful to assess teleconnections through ENSO processes with the Andean ice fields in order to acquire understanding about large-scale impacts of this phenomenon in the past.

Figure 1.1 The warmest (a) upper-level atmospheric temperatures and (b) sea surface temperatures are located in the tropical latitudes, as are (c) the regions of most intense convective precipitation (modified after Sobel (2002)). ECMWF: European Centre for Medium-Range Weather Forecasts; Figure 1 from Thompson et al. (2011)
Glaciers near Puncak Jaya (4°05’ S, 137°10’ E) are located at an elevation of 4,884 meters above sea level (masl), the highest peak between the Himalayas and the Andes, in the heart of the West Pacific Warm Pool (WPWP), the heat engine for the globe's climate system (Fig 1.1; Fig 1.2a). During the Papua ice core drilling expedition from May to June 2010, three ice cores were recovered from the East Northwall Firn (Fig
1.2b). Two ice cores to bedrock were drilled from the west dome of the ice field (Sumantri peak), measuring 32.13 m (D1) and 31.25 m (D1B) long, and a third core (D2) was drilled from the east dome (Soekarno peak/ Ngga Pulu) measuring 26.19 m (Fig 1.2c) (Stone, 2010; Permana, 2011; Thompson, 2011; Thompson et al., 2011). Drilling in the D2 borehole was terminated when a crevasse was encountered. In addition, in order to support the interpretation of stable isotopes from the Papua ice cores, 1362 rain samples collected daily during the period of January 2013 to February 2014 and December 2014 to September 2015 along the altitudinal transect (9 - 3,945 masl) on the southern slope of the central mountain ranges in Papua were analyzed for stable isotopes. These data, along with the instrumental meteorological measurements, contribute to a better understanding of the factors that control the temporal variability of rainfall stable isotopes.

Here the first tropical ice core-based climate history from glaciers in the WPWP are investigated. The objectives of this study are to: (1) understand factors that control the temporal variability (daily to interannual) of the stable isotopic ratio in rainfall in the Papua region based on ~2-years of daily rain samples and meteorological data at various elevations across the southern slope of the central mountain ranges in Papua; (2) determine the age of the 2010 Papua ice cores using variations in physical and chemical parameters, including stable isotopes of oxygen and hydrogen, tritium (H³), major ions, and dust concentrations; (3) reconstruct the tropical Pacific climate variability associated with the ENSO and the monsoon system as recorded in the Papua ice core records; (4) compare the Papua ice core records with other paleoclimate records in the vicinity of Papua; and (5) examine the contribution of the retreat of the Papua glaciers to the general
understanding of the causes of glacier retreat in the tropical mountain regions over the last century.

The results of this research provide information about ENSO variability on the western side of Pacific Ocean, which allows an assessment of the teleconnection among tropical Pacific SSTs, precipitation and ice core records in Papua and the South American Andes through ENSO processes. This is important to acquire a better understanding of the large-scale impacts of this phenomenon in the past. Furthermore, findings from the stable isotope analysis of both the precipitation and ice core records in Papua may provide observational benchmarks for isotope-enabled climate modeling in assessing the future climate variability.
2.1 Climate of Papua

Papua (8°S–0°S; 130°–141°E) province is the largest and easternmost of the 33 provinces in the Republic of Indonesia. With a land area of 319,036 km² (16.62 % of the Indonesian region), it occupies the western half of New Guinea Island. Papua is bordered on the east by Papua New Guinea (PNG) and on the west by West Papua province, and is surrounded by the Pacific Ocean to the north, the Arafura Sea (AS) to the south and the Banda Sea (BS) to the east. The island is divided from the east-southeast to the west-northwest by mountain ranges that exceed 3,500 meters above sea level (masl), including the highest peak, Puncak Jaya (4,884 masl) (Fig 1.2a). The Papua land mass is covered mostly by tropical lowland and montane rainforest with wetlands in some coastal regions and grasslands in the high mountains (Bowler, 1976).

Due to its equatorial position, climate of Papua is influenced by three major large-scale atmospheric circulations (McAlpine et al., 1983; Prentice and Hope, 2007). The meridional Hadley circulation is composed of equatorward flow near the surface, rising air near the equator where SSTs are warmest, poleward air flow in the upper level atmosphere and descending air in the subtropics. Within the Hadley cell the surface air flows, or trade winds, move towards the equator in both hemispheres and converge and
rise at the Intertropical Convergence Zone (ITCZ), a high precipitation band of thunderstorms. Annually, the ITCZ migrates ~15° north and south in association with the warmest SSTs. This movement is responsible for the seasonal rainfall in Papua.

The South Pacific Convergence Zone (SPCZ), a broad zone of low-level convergence, high cloudiness, and enhanced precipitation, is located in the south Pacific sector of the WPWP (Kiladis et al, 1989; Vincent, 1994). The SPCZ is the largest and more persistent portion of the ITCZ that stretches southeast from the WPWP to the Polynesian Islands. On seasonal timescales, the SPCZ is most active in austral summer, but it is persistent year-round. Along with the Hadley circulation, the zonal Walker circulation, which is thermally driven by the SSTs gradient across the tropical Pacific, influences the climate of Papua by enhancing ascending air in the WPWP (Fig 2.1).

![Figure 2.1 Major large-scale atmospheric circulations influence the Papua climate](image source: http://clb.csp.escience.cn/getCachedDocument.do?docid=13712)

Papua is one of the wettest regions on Earth, with many sites receiving 2,500 – 4,500 mm of precipitation annually (Prentice and Hope, 2007). At the southern slope of the central Cordillera, the annual precipitation ranges between ~5,000 and 12,000 mm.
with the highest amount (~12,500 mm/yr) recorded at Mile50 station (4.28°S; 137.01°E; 617 masl) (Permana, 2011). During austral summer (the “northwest” season) from December to March, the climate is dominated by the monsoon westerlies (3 - 5 m/s) which result from veering of the northeast trade winds over the equator due to heating over the New Guinea land masses (Prentice and Hope, 2007). Most of Papua experiences rainfall maxima and the island is almost continuously wet during this season (Fig 2.2a), with the monsoon westerlies prevailing up to an elevation of about 600 millibars (mb). The weather pattern is dominated by thunderstorms from a low pressure system associated with enhanced deep convective activity which is driven by the ITCZ position at the south of the island.

Figure 2.2 Mean monthly precipitation (shaded, TRMM 3B43V7) and mean 10-m winds (vector, NASA MERRA Reanalysis) from January 1998 – December 2013 for (a) January, (b) April, (c) July, and (d) November
During the austral winter (the “southeast” season) from May to October, the southeast trade winds (~ 7 m/s) flow over the entire Papua region. The southern parts of the island experience drier conditions with minimum rainfall in June - July (Fig 2.2c). Local convection circulations, e.g. trade wind orographic showers associated with high pressure systems, and local wind convergence become dominant during this season. In addition, the Trade Wind Inversion (TWI) layer is present at about 2,000 masl south of the Cordillera as a result of the ITCZ position in the Northern Hemisphere (Fig 2.3) (McAlpine et al., 1983; Prentice and Hope, 2007). The TWI layer, indicated by a strong temperature inversion, is a controlling factor for cloud development which caps local convective circulation and mixing processes. During April and November (transitional season), both of monsoon westerlies and the southeast trade winds influence Papua (Fig 2.2b,d).

In Papua, local conditions (e.g. topography, wind circulation and local sea surface temperature) are occasionally more dominant and overshadow the impact of large-scale circulation on the local climate. For instance, Aldrian and Susanto (2003) identified a reverse seasonality of rainfall in some regions of Papua. One of those regions is located in the lowlands in the southern part of Puncak Jaya peak. The rainfall peaks from June to August and is relatively low from December to February. During the southeast season, lowland meteorological data from this region demonstrate that rainfall at night (6 PM – 6 AM local time) is much higher than in the day (6 AM – 6 PM local time), with differences of up to 250 – 350 mm/month occurring from June to August (Permana,
However, data from highland stations show little seasonality and equal or more rainfall during the day than at night in this season.

Diurnal rainfall variability in this region during the southeast season is mainly caused by local wind regimes and orographic lifting resulting in a shallow local convective system that is limited by the presence and the strength of TWI (McAlpine et al., 1983). During daytime, particularly in late afternoon, the rainfall on the lowland is influenced mainly by southeast trade winds transporting moisture from the Arafura Sea, enhanced by upslope anabatic (valley) winds and sea breezes at the southern coast. As it interacts with the steep topography, orographic lifting of the moisture occurs, producing high rainfall over the slopes and highlands. At night, strong southeast trade winds converge with downslope katabatic (mountain) winds enhanced by land breezes at the southern coast, leading to offshore convergence and producing high nocturnal rainfall on the southern lowlands (Fig 2.3). A similar nocturnal rainfall mechanism also occurs during the northwest season due to offshore convergence of katabatic flow with monsoon westerlies (McAlpine et al., 1983). The only difference between the two seasons is the presence of the TWI, suggesting that more nocturnal rainfall occurs from shallow convection systems during the southeast season. The impacts of trade winds and local wind circulation on the diurnal rainfall have been observed in Papua New Guinea (McAlpine et al., 1983) and the island of Hawaii (Chen and Nash, 1994; Esteban and Chen, 2008) which has an analogous topography to southern Papua. The influence of topography on the diurnal precipitation cycle has been presented by Zhou and Wong (2006) using satellite products and modeling over New Guinea Island. A recent study
suggests that diurnal variation, rather than monsoon variation, is the dominant factor that controls precipitation intensity in this region (Christianto, 2014).

Figure 2.3 A cross section of central New Guinea Island during the austral winter (the southeast season) (after McAlpine et al., 1983; Prentice and Hope, 2007). WNW is west-northwest and ESE is east-southeast wind direction. During the daytime, sea breezes and anabatic winds prevail with similar directions as the southeast trade winds

The temperature regime is typically equatorial with a seasonal range of ~1 to 3°C and a diurnal range of ~6 to 14°C. Moderately high temperatures occur at sea level, varying between 22°C and 34°C. The southern mountain ranges of Papua experience maxima (minima) temperature in December - February (June - August), but the more
equatorward regions experience two maximum (minimum) temperatures in May and November (February and July). Due to high relative humidity in this region, weather station data suggest that the regional surface lapse rate is \(~5.0^\circ\text{C/km}\) (moist adiabatic lapse rate). Below 2,500 masl the lapse rate is \(5.2^\circ\text{C/km}\), while above that altitude the rate decreases to \(4.6^\circ\text{C/km}\) (Permana, 2011).

Besides seasonal variation, Papua is also influenced by large-scale intraseasonal and interannual variability that occur in the WPWP, which are described in the following sections.

### 2.1.1 Impact of Madden Julian Oscillation (MJO)

The most prominent mode of intraseasonal variability in the Tropics is the Madden-Julian Oscillation (MJO) which has 30–60 day periodicity and is characterized by an eastward propagation of both enhanced and suppressed tropical cloud convection across the equatorial Indian Ocean to the western and central Pacific Ocean (Madden and Julian, 1971; 1994). Such patterns have been extensively documented using satellite-measured outgoing longwave radiation (OLR) at the top of the atmosphere as a proxy of tropical convection and rainfall (e.g. Arkin and Ardanuy, 1989). The OLR value represents an integral measurement of the radiative effects from Earth's surface, clouds, and gases in the atmosphere. Low OLR values imply enhanced deep convection (high precipitation), while high OLR values indicate suppressed convection (less precipitation). The impact of the MJO intraseasonal cycle in New Guinea and the western Pacific has been discussed in previous studies (Matthews et al., 2005; 2013; Ichikawa
and Yasunari, 2008; Peatman et al., 2014). Generally, active (wet) phases of the MJO correspond to enhanced tropical rainfall in the tropical Western Pacific, with typical rainfall anomalies between the wet and dry phases in the range of $2 - 6\ \text{mm/day}$ (Matthews et al., 2005). MJO signals are observed to be strong and peak in austral summer (December – March) in the western Pacific and reach their maxima in the SPCZ (Zang and Dong, 2004). Furthermore, the MJO signals systematically modulate the amplitude of the diurnal cycle of precipitation over New Guinea during austral summer (Ichikawa and Yasunari, 2008; Peatman et al., 2014).

### 2.1.2 Impact of El-Niño Southern Oscillation (ENSO)

The ENSO phenomenon, which involves the coupled interaction between atmospheric pressures and SST contrasts along the tropical Pacific Ocean, is a major controlling factor of interannual variability of climate over the globe. ENSO events, which occur every 3 to 7 years, are associated with the Walker circulation such that when the latter weakens (El Niño condition, ENSO warm phase), the trade winds also weaken and the warm SSTs of the WPWP spread to the central and eastern Pacific. Otherwise, when the Walker circulation is enhanced (La Niña condition, ENSO cold phase), trade winds are stronger and SSTs are warmer in the WPWP (Fig 2.4). The impact of ENSO in the Maritime Continent (MC) and western Pacific Islands has been broadly discussed in previous studies (Hendon, 2003; McBride et al., 2003; Smith et al., 2013; Chung et al., 2014; Murphy et al., 2014). Drought conditions typically occur during El Niño in the MC when SSTs in the WPWP are cool, which leads to less evaporation and hence less precipitation. The opposite tends to occur during La Niña.
Figure 2.4 Generalized Walker Circulation (December-February) during (top) ENSO-neutral conditions, (middle) El Niño events and (bottom) La Niña events. The middle and bottom figures are overlaid on a map of average SST anomalies with anomalous ocean warming (cooling) shown in orange (blue).
ENSO events can significantly change the amplitude of seasonal rainfall in Papua (Prentice and Hope, 2007). For instance, lowlands in Papua experience periods of unusually low rainfall during El Niño events (e.g. Sukri et al., 2003 and Fig 22 in Permana, 2011). In the southern PNG, drier (wetter) conditions occur at Port Moresby during El Niño (La Niña) years. At the same time, dry (wet) seasons tend to be cooler (warmer) in El Niño years and the opposite in La Niña years (BoM and CSIRO, 2011). In the New Guinea highlands (~1,500 - 2,500 masl), the occurrences of droughts and frosts were identified to be mostly associated with El Niño events which affected the indigenous people and their crops (Allen, 1989; Allen et al., 1989; Allen and Bourke, 1997). Frost in the tropical highlands is mainly caused by cloudless night skies that allow surface heat to radiate from the ground surface (radiative cooling). These conditions are unusual in the Tropics and possibly occur when a stable air mass with low moisture content descends over the New Guinea highlands, such as during El Niño events. The frost events were identified in El Niño events in the 1972, 1982 and 1997. For example, at Tambul in the Western Highlands, PNG (~2,200 masl), temperatures were below freezing in June and July 1997 and plunged below zero over several nights in August and September, with the coldest night (−2.3 °C) occurring in September 1997 (Allen and Bourke, 1997). At larger spatial scales, maps of possible climatic impact of ENSO extremes on the Asia-Pacific Region were produced in 2009/2010 by the United Nations Office of the Coordination of Humanitarian Affairs (UNOCHA) based on data from the National Oceanic and Atmospheric Administration (NOAA) and the Royal Netherlands Meteorological Institute (KNMI) (Appendix A).
Recent studies, however, suggest that the rainfall impacts of ENSO in the western Pacific region may be different at highlands compared to surrounding regions, most likely because of the interaction of wind and topography (Smith et al., 2013). In addition, ENSO may also have a nonlinear response to the rainfall changes, such as in New Ireland and New Britain regions that tend to be drier during both ENSO extremes (Fig 2.5; Smith et al., 2013; Chung et al., 2014).

Figure 2.5 The relationship between station seasonal rainfall totals and the NINO3.4 SST anomalies over New Guinea region. The four circles correspond to four seasons starting on JFM, AMJ, JAS and OND. Red (blue) indicates a significant negative (positive) correlation ($p < 0.05$), and white indicates no correlation. The blue hatched circles denote a significant second-order polynomial fit to the data; modified from Figure 6 of Smith et al. (2013)

In the Pacific Ocean, there is a long-lived El Niño-like pattern of climate variability which is known as the Pacific Decadal Oscillation (PDO). The PDO phase is defined by the surface water temperature in the Pacific Ocean north of 20°N (Mantua et
During a positive/warm phase, the western Pacific cools, and part of the eastern Pacific warms. The opposite pattern occurs during a negative/cool phase. The PDO has a periodicity of ~20 – 30 years. In the last century, cool PDO regimes prevailed from 1890 – 1924 and from 1947 – 1976, while warm PDO regimes dominated from 1925 – 1946 and from 1977 through at least the mid-1990. The warm (cold) phase of PDO influences the rainfall patterns in the eastern MC by a decrease (increase) in the rainfall (Lee, 2015). The PDO has similar impacts as ENSO to the rainfall in Papua region, but over longer timescales.

2.2 Glaciations in New Guinea Highlands

The tropical glaciation history has been reviewed in several studies (Porter, 2001; Mark et al., 2005; Hastenrath, 2009). Mauna Kea in Hawaii and New Guinea are the only two equatorial Pacific islands that contained glacier cover during the Pleistocene. The only evidence for Mid-Pleistocene glaciation in New Guinea comes from Mt. Giluwe (4,367 masl), PNG which contains the oldest moraines with ages of at least ~158 thousand years before present (kyrs BP) and possibly ~300 kyrs BP (Barrows et al., 2011; Prentice et al., 2011). Most glacial-geologic studies have focused on the eastern half of the island (PNG), which is more accessible than Papua, Indonesia to the west, although currently no glaciers exist in PNG (Löffler, 1972; Brown, 1990; Barrows et al., 2011; Prentice et al., 2005; 2011). However, studies concerning past and recent glaciations have also been performed in the Papua highlands (Peterson and Hope, 1972; Hope et al., 1976; Prentice et al., 2005; 2011; Prentice and Glidden, 2010). There are four areas of Papua that are known to have supported glaciers during the 20th century: Mt
Idenberg (Ngga Pilimsit); Mt Jaya (Puncak Jaya); Mt Trikora (Mt Wilhelm); and Mt Mandala (Puncak Mandala) (Fig 2.6). Glaciers have disappeared from three of these mountains. A small ice cap on Mt. Trikora melted during the period 1939 - 1962 (Mercer, 1967), while glaciers on Mt. Idenberg and Mt. Mandala disappeared by 2003 (Klein and Kincaid, 2006; 2008).

Figure 2.6 The location of areas known to contain glaciers during the 20th century in Papua Indonesia; Figure 73.4 from Prentice et al. (2011)

2.2.1 Glaciological Overview

At least three, and probably four, distinct glacial periods have been recognized based on exposure-age dating on the moraines of Mt. Giluwe, PNG (Barrows et al.,
2011). The Last Glacial Maximum (LGM) in New Guinea has been reported between

$\sim 22 \text{ – 18 kyrs BP}$ based on pollen-derived analysis (Bowler \textit{et al}., 1976) and maximum
depression of vegetation zones (Hope and Peterson, 1976). The LGM glacial limit was
likely higher in the west (Papua), with a snowline at $3,700 \text{ – 4,050 masl}$, than to the east
(PNG) where the snowline was $3,400 \text{ – 3,600 masl}$ with glacial termini as low as $3,100
masl (Mark \textit{et al}., 2005; Prentice \textit{et al}., 2011; Hope, 2014). At the LGM, a recent study
suggested that $\sim 3,770 \text{ km}^2$ (Papua, $\sim 3,270 \text{ km}^2$; PNG, $500 \text{ km}^2$) of New Guinea was
glaciated based on the estimated snowlines and treelines (Hope, 2014). This is more than
twice the previous estimate derived from maps (e.g. Hope and Peterson, 1976).

Temperatures are estimated to have been $\sim 5 \text{ – 8°C lower}$ than present (Brown, 1990;

Earlier deglaciation dates have been reported ($\sim 17 \text{ – 13 kyrs BP}$) in Papua and the
main valley due to the drier climatic regime in the west, and later ($\sim 11 \text{ – 8 kyrs BP}$) in
PNG and the main summit (Hope and Peterson, 1975; 1976; Prentice \textit{et al}., 2011). Small-
scale glacier re-advancement occurred $\sim 13 \text{ – 11 kyrs BP}$ on Mt. Jaya as a possible
response to increased precipitation as the Arafura shelf flooded due to eustatic global sea
level rise during deglaciation; however temperatures remained cold (Hope and Peterson,
1975; Brown, 1990). Complete deglaciation had possibly occurred by 9 kyrs BP and 7
kyrs BP in PNG and Papua, respectively, suggesting that New Guinea may have been
completely ice free by 7 kyrs BP (Brown, 1990). Neoglacial advances have only been
documented on Mt. Jaya at four time periods after $\sim 5$ kyrs BP: $3.5 \text{ – 2.9 kyrs BP}$, $2.5 \text{ – 1.5 kyrs BP}$, within the last $1.5$ kyrs BP with the last cooling episode being the Little Ice
Age ending circa 140 – 170 years ago. After that, continuous glacier retreat has occurred to the present day (Brown, 1990).

The snowline depression of tropical glaciations has been reviewed by previous studies (Porter, 2001; Mark et al., 2005; Hastingrath, 2009). The equilibrium line altitudes (ELA, altitude at which accumulation equals melting) of New Guinea glaciers have increased from ~3,750 masl in the LGM to ~4,850 masl by the end of the 20th century (Prentice et al., 2005; 2011; Prentice and Hope, 2007; Prentice and Glidden, 2010). This suggests ~1,100 m of snowline depression between the LGM and the present. By considering ~125 m lowering of sea level during the LGM in this region (e.g. Yokoyama et al., 2001) and the modern lapse rate of ~5.5°C/km (Permana, 2011), it would suggest that the temperature in the New Guinea highlands during the LGM was at least ~5°C cooler than present (possibly more, as precipitation decreased during the LGM). This highland temperature depression is in contrast to that at sea level, which was about 1 – 3 °C cooler than present (Farrera et al., 1999), implying a steeper present lapse rate than during the LGM.

2.2.2 Glacier Recession in the 20th Century

In 1972, there were seven major ice masses near Puncak Jaya: the East and West Northwall Firn, Meren, Carstensz, Wollaston, Van de Water, and Southwall Hanging Glaciers (Fig 2.7) (Allison, 1974). Recently, some of them have disappeared. By 2005, only smaller parts of the East and West Northwall Firn, Carstensz and Southwall Hanging
Glaciers remained (Kincaid, 2007). In 2010, the last part of the Southwall Hanging Glacier disappeared (Permana, 2011).

Figure 2.7 Map of the glaciers extent near Puncak Jaya in 1972 (Hope et al., 1976), 1987 (Allison and Peterson, 1989), 2002 (Klein and Kincaid, 2006) and 2005 (Kincaid, 2007). The background image is the June 11, 2002 IKONOS image; modified from Figure 18 of Kincaid (2007)
The retreat of glaciers near Puncak Jaya has been occurring from the end of most recent Neoglacial (~1850) to the present day (Peterson et al., 1973; Allison, 1974; Hope et al., 1976; Allison and Kruss, 1977; Peterson and Peterson, 1994; Van Ufford and Sedgwick, 1998; Klein and Kincaid, 2006; Kincaid, 2007; Prentice and Hope, 2007; Prentice and Glidden, 2010; Prentice et al., 2011). The total ice area has decreased from about 19 km$^2$ during the most recent neoglacial to 3 – 4 km$^2$ in 1993/1994 (Peterson and Peterson, 1994; Van Ufford and Sedgwick, 1998). A recent study suggested that the total ice area in ~1850 was more extensive, 30 km$^2$, based on moraines and current topographic divides (Prentice et al., 2012). Another study reported that the ice area receded from ~11 km$^2$ in 1942 to 2.4 km$^2$ by 2000 (Prentice and Hope, 2007). Based on satellite images, the ice area was 2.15 km$^2$ in 2002 (Klein and Kincaid, 2006) and ~1.8 km$^2$ in 2005 (Kincaid, 2007). A recent assessment reported that the glacier area of the East Northwall Firn was about 0.96 km$^2$ in 2006, based on digital aerial photographs (Prentice and Glidden, 2010). In May 2011, the total ice cover was only 1 km$^2$ remained (Prentice et al., 2012). The retreat of the East Northwall Firn has continued in recent years as shown by comparisons of recent photographs from June 2010 and March 2015 (Fig 2.8).

The specific mass balance (the difference between accumulation and ablation rates) for Carstensz and Meren Glaciers were determined to be –0.06 mwe/yr (meter water equivalent per year) and –0.51 mwe/yr, respectively, during the Carstensz Glacier Expeditions (CGE) in 1972/1973 (Allison, 1976). More recently, the specific mass
Figure 2.8 Comparison of aerial photographs of the East Northwall Firn taken in June 2010 (by Endang Budianto) and March 2015 (by Yohanes Kaize). Yellow line indicates the same boulder on both photographs.
balance for Meren Glacier was estimated to be $-1.52$ mwe/yr during 1973 – 1995, $-3.3$ mwe/yr during 1995– 1997 and $-0.55$ mwe/yr during 1997 – 2000. In addition, the specific mass balance for the East Northwall Firn was reported at $-3$ mwe/yr during 1995 – 1997 and only slightly positive (0.34 mwe/yr) during 1997 – 2000 (Prentice and Glidden, 2010).

The controlling factors for tropical glacier retreat in the last century have been widely discussed in the case of the Tropical Andes (Thompson et al., 1994; 2000; 2013; Francou et al., 2003; 2004) and Tropical Africa (Thompson et al., 2002; 2009; 2011a; Taylor et al., 2006; Kaser et al., 2004; Mölg et al., 2003; 2008; 2009). Studies in the New Guinea highlands are older (Peterson et al., 1973; Allison and Peterson, 1976; Allison and Kruss, 1977; Kincaid, 2007). Accelerated atmospheric warming has been a primary driver of the recent tropical glacier recession in Tropical Africa (Thompson et al, 2000; 2002; 2009; 2011a; Taylor et al., 2006).

However, modeling studies suggest that atmospheric drying and decreased precipitation are the main factors for glacier recession in Kilimanjaro, Africa during the 20th century (Mölg et al., 2003; 2009; Kaser et al., 2004). Decreased precipitation causes low cloudiness and increased solar radiation absorption at the glacier surface, thus decreasing the glacier albedo and leading to ablation (sublimation). This argument has been countered by the presentation of the linear relationship between oxygen and hydrogen isotopic ratios for all the summit ice cores drilled on Kilimanjaro, which indicate that neither evaporation nor sublimation, play major roles in modern ice loss on Kilimanjaro (Thompson et al., 2011a).
On the eastern side of the Pacific, glaciers in the tropical Andes respond to changes in both air temperature and atmospheric moisture content related to ENSO events (e.g. Francou et al., 2003; 2004). During El Niño, air temperature increases and precipitation decreases in this region. The warming favors rain over snowfall, while decreased precipitation tends to reduce cloud cover and causes more absorption of solar radiation at the ice surface. Both of these feedbacks induce ablation with melting predominating over sublimation due to high humidity and low wind speeds. The opposite scenario tends to occur during La Niña. Similar to the Papua highlands in the western Pacific, the acceleration of glacier recession has likely been controlled by rising air temperatures, more precipitation as rainfall over snowfall, increased radiation absorption, increased evaporation, or combinations of these factors (Peterson et al., 1973; Allison and Peterson, 1976; Allison and Kruss, 1977; Kincaid, 2007). During the 1972 El Niño, solar radiation absorption may have been the predominant factor controlling ablation on the glaciers, with melting favored over evaporation/sublimation due to relatively humid air and low wind speed (Allison, 1976). This process was enhanced by the existence of cryo-vegetation colonies on the glaciers which reduced the ice albedo (Kol and Peterson, 1976). More recently, an increase in monthly atmospheric temperature by 0.24°C from 1972 to 1987 strongly affected the ice loss on the glaciers. From 1988 to 2005, the glacier retreat has accelerated as precipitation actually increased along with a rising freezing level height, which may have led to a greater proportion of the glacier surface being affected by rain (Kincaid, 2007).
2.3 Chemical and Physical Data from Ice Cores

Ice cores contain valuable information about the climate and environment which can be extracted from physical and chemical characteristics. The relevant parameters in this study, including stable isotopic composition, insoluble dust and several major chemical ions, are described in the following sections:

2.3.1 Stable Isotope Composition

Many elements on Earth exist in radioactive or stable isotopic form in which the isotopes contain equal numbers of protons but different numbers of neutrons in their nuclei. In ice core paleoclimatology, oxygen and hydrogen stable isotopic compositions are important climate proxies that can be derived from glacier ice. Oxygen has three stable isotopes: $^{16}$O, $^{17}$O and $^{18}$O with relative abundances of 99.76%, 0.04% and 0.2%, respectively while hydrogen has two stable isotopes: H and $^2$H (deuterium, also referred to as D) with relative abundances of 99.984% and 0.016%, respectively (Bradley, 1999). Despite nine possible isotopic combinations, only two important "heavier" molecules (H$_2^{18}$O and HDO) are commonly used in paleoclimate research. The oxygen and hydrogen isotopic fractionation (separation of the heavier and lighter isotopes) in water molecules is influenced mostly by evaporation and condensation. Since these "heavier" water molecules have lower vapor pressures than the more abundant lighter H$_2^{16}$O, the lighter evaporates first, and the remaining water becomes relatively enriched in deuterium and $^{18}$O. The opposite tends to occur during condensation, where heavier water molecules are removed initially. Evaporation and condensation are highly dependent upon
temperature. As air cools by rising in the atmosphere or moving toward the poles, the heavier water molecules begin to condense as precipitation, leading to the remaining moisture in the air becoming increasingly depleted in deuterium and $^{18}$O. Hence, the subsequent precipitation will become more depleted in D and $^{18}$O than the initial condensate. The temperature during condensation partially determines the precipitation isotopic composition (Dansgaard, 1964). Stable isotopic ratios ($^{18}$O/$^{16}$O and D/H) are calculated by relative comparison with the Standard Mean Ocean Water (SMOW) and are denoted as $\delta^{18}$O and $\delta$D in unit of "per mil" or ‰.

Globally, the distribution of $\delta^{18}$O and $\delta$D in precipitation is affected mainly by temperature, precipitation amount and moisture sources. As temperature decreases toward higher latitudes (latitude effect) and altitudes (altitude effect), $\delta^{18}$O and $\delta$D in precipitation become more depleted. A linear relationship has been established between $\delta^{18}$O and surface temperature at mid to high latitudes (Dansgaard, 1964; Jouzel et al., 1987). On regional and local scales, precipitation isotopic ratios are controlled by precipitation amount according to Rayleigh distillation, particularly in the Tropics (Dansgaard, 1964). The $\delta^{18}$O and $\delta$D in precipitation become progressively depleted with increased precipitation (amount effect) or distance inland (continental effect), where heavy isotopes in the atmosphere are “rained out”. Moisture source may also control the precipitation isotopic ratio, such that oceanic sources are isotopically more depleted than the terrestrial sources, and along with temperature and precipitation amount govern the seasonal precipitation isotopic composition (seasonal effect).
The linear relationship between $\delta^{18}O$ and $\delta D$ in natural waters is known as the global meteoric water line (GMWL; $\delta D = 8\delta^{18}O + 10$) (Craig, 1961). In an evaporating water body, the slope depends on local relative humidity. Low humidity leads to slopes smaller than eight, while high humidity results in slopes closer to eight (Kendall and McDonnell, 1998). The intercept value of the meteoric water line (MWL), defined as deuterium excess ($d = \delta D - 8\delta^{18}O$), is a useful isotopic tracer of the atmospheric vapor source in precipitation (Dansgaard, 1964). When ocean water evaporates, $d$ values are fixed by the relative humidity of the air mass. The mean $d$ value of GMWL is ~10‰, resulting from evaporation with an average relative humidity of ~85% (Merlivat and Jouzel, 1979; Clark and Fritz, 1997). The $d$ values can also be affected by SSTs and wind speed during evaporation at the oceanic source area (Merlivat and Jouzel, 1979; Benetti et al., 2014; Pfahl and Sodemman, 2014), and can also be used to trace the effect of continental moisture recycling in precipitation (Salati et al., 1979; Gat and Matsui, 1991). For example, the moisture recycling over the Amazon Basin via terrestrial evaporation causes $\delta^{18}O$ enrichment and larger $d$ values in precipitation during the winter (Windhorst et al., 2013). In addition, vapor condensation from ice in clouds can generate an isotopic fractionation that may increase the $d$ value in precipitation (Gonfiantini et al., 2001). Over continents with low relative humidity, raindrop (secondary) evaporation below the cloud base during rainfall is likely to lower $d$ values in the precipitation (Rozanski et al., 1993).
2.3.2 Dust and Chemistry Content

Insoluble microparticle dust has been used to identify seasonal variations in the Quelccaya tropical ice cores (Thompson et al., 1979; 1985; 1994). Generally, higher dust concentrations on glaciers were observed in the dry season, either through wet (raindrops, snow, and fog) or dry (direct aerosol particles) deposition. Several known major explosive volcanic eruptions were also recognized by tephra (volcanic ash) layers in polar ice cores, which also can be used as a time marker in the ice core dating (e.g. Zielinski et al., 1997).

Along with dust content, major ions within the ice provide insights in the overlying atmosphere. The knowledge of ion sources helps in the understanding of atmospheric circulation during deposition. For instance, sodium and chloride ions are a good tracer for sea salt, while calcium and magnesium ions may reflect the terrestrial dust origins due to rock weathering. Ammonium may indicate the biogenic emissions from the environment surrounding the glaciers. Among other sources, increased ammonium ion concentrations may result from the forest fires. Potassium ions may originate from tropical savanna fires (Echalar et al., 1995). Thus both ammonium and potassium ions may serve as biomass burning tracers. In addition, nitrate may originate from biological sources (e.g. Lyons et al. (1985)). High sulfate composition in an ice layer can be used to detect volcanic eruption events in the past, along with volcanic ash (tephra). The presence of nitrate and sulfate has been used to detect anthropogenic emissions from fossil fuel burning during the industrial era (Legrand and Mayewski, 1997).
Radioactive fallout from nuclear bomb tests between the 1950’s and 1970’s can be detected by measuring $^3$H (tritium) content or Beta radioactivity in ice core layers deposited during the mid-20th century. The tritium peaks in an ice core record may serve as reference horizons which provide strong time markers which are useful for ice core dating procedures (e.g. Vimeux et al, 2009).

2.4 Controls on Tropical Rainfall Stable Isotopes

In the Tropics, there is a significant negative correlation between the stable isotope ratios in rainfall and the amount of precipitation, which is known as the stable isotope amount effect (Dansgaard, 1964; Rozanski et al., 1993; Gonfiantini et al., 2001). However, at high altitudes in the Tropics, air temperature remains the primary driver on longer timescales (i.e., longer than annual) (Thompson et al., 2000; 2006). On seasonal timescales, the amount effect results in isotopic values being more depleted in the rainy season than in the dry season (Gonfiantini et al., 2001). The amount effect in rain has been ascribed to fractionation processes that occur on both local (Vuille et al., 2005; Lee et al., 2009) and larger spatial scales (Cobb et al., 2007; Risi et al., 2008; Kurita et al., 2009, 2011; Moerman et al., 2013; Kurita, 2013). Moreover, previous studies also have linked rainfall isotopic composition to moisture sources, moisture transport history and/or prevailing weather patterns (Rhodes et al., 2006; Scholl et al., 2009; Breitenbach et al., 2010; Fudeyasu et al., 2011; Crawford et al., 2013; Suwarman et al., 2013; Windhorst et al., 2013).

The mechanisms that lead to the observed isotope amount effect are poorly understood. The amount effect has been explained as a consequence of the extent of the
rainout process from deep convective clouds in the ITCZ (Rozanski et al., 1993). In this explanation, greater rain intensity during storm events results in lower isotope ratios while less intense rain storms tend to generate precipitation with more enriched isotope ratios. Because heavy water vapor is preferably removed as condensation, the remaining vapor becomes lighter. Thus, subsequent precipitation that forms from a given convective air mass will be more isotopically depleted.

Alternatively, the amount effect has been interpreted by the direct and indirect effects of unsaturated downdrafts in the convective system (Risi et al., 2008; 2008a). The direct effect of the unsaturated downdrafts is associated with processes of rainfall re-evaporation and equilibration through diffusive exchanges. The indirect effect is referred to as downdraft recycling, which decreases the isotopic ratios of atmospheric vapor by injecting isotopically depleted vapor from the downdrafts into the sub-cloud layer which feeds the convective system (Risi et al., 2008a). A recent study suggests that rainfall isotopic variations in the Tropics is not directly controlled by the precipitation amount, but depends on large-scale convective activity (Kurita, 2013) (e.g. Lekshmy et al. (2014) in southern India during the monsoon season). Moreover, moisture convergence also has been suggested to affect the rainfall isotopic composition (Lee et al., 2007; Moore et al., 2014). Results from other studies show that upwind regional convective activity during the monsoon season mainly controls rainfall isotopic variation in the southern Tibetan Plateau (Gao et al., 2013) and over the tropical Andes (Samuels-Crow et al., 2014).

In the Tropics, most precipitation falls from high convective clouds during the summer monsoon, corresponding to intensively deep convection originating from low
pressure systems following the annual movement of the ITCZ. Outgoing longwave radiation (OLR) at the top of the atmosphere is routinely used to locate areas of deep convection and precipitation (Arkin and Ardanuy, 1989). The OLR value represents an integral measurement of the radiative effects from the surface, clouds, and gases in the atmosphere. In general, enhanced deep convection associated with high precipitation are identified by low OLR values ($< 205 \text{ Wm}^{-2}$; Gu and Zhang, 2002), while high OLR values ($> 205 \text{ Wm}^{-2}$) indicate reduced convection with less precipitation or clear-sky condition. Therefore, precipitation amount and OLR are generally inversely correlated on seasonal timescales in the Tropics. In the tropical Maritime Continent, seasonal rainfall isotopic variations have been linked to local or regional precipitation amounts (Kurita et al., 2009), moisture origins (Suwarman et al., 2013) and are influenced by the strength of the regional convective activity (Moerman et al., 2013; Kurita, 2013).

On intraseasonal timescales, rainfall stable isotopes and large-scale convective activity have also been correlated, as reported in tropical South America (Vimeux et al., 2011), Africa (Risi et al., 2008; Tremoy et al., 2012), the southern Tibetan Plateau (Gao et al., 2013) and Southeast Asia (Moerman et al., 2013). For instance, Kurita et al. (2011) showed that rainfall isotopic variation is associated with the intraseasonal MJO cycle. Moreover, large depletions of rainfall $\delta^{18}O$, as indicated by low OLR values, are associated with the wet phases of the MJO (Moerman et al., 2013).

From a different perspective, OLR primarily reflects the temperature at the emission level of infrared radiation. In cloudy conditions, this reflects the altitude of the cloud top. Therefore, in the Tropics, lower OLR values also correspond to higher cloud
top altitudes and are likely associated with cooler temperatures at which precipitation forms during convective storms and vice versa. For example, based on the echo tops (average altitude of precipitation within clouds) data, previous studies in Puerto Rico and Hawaii concluded that rain $\delta^{18}$O seasonality is influenced by atmospheric temperatures that correspond to different cloud heights associated with the seasonal climate patterns (Scholl et al., 2009; Scholl and Coplen, 2010). During summer, deep convection and increased cloud heights occur due to low pressure systems which leads to $^{18}$O depletion in rainfall, while during winter orographic precipitation and high pressure systems are predominant where $^{18}$O enrichment in rainfall is controlled by the limited cloud height under the TWI layer (Scholl et al., 2009). The relationship between $\delta^{18}$O and temperatures at the mean condensation levels has previously been proposed for the interpretation of the tropical ice core records (Thompson et al., 2000; 2003). Modeling studies in the western U.S. have also revealed the relationship between rainfall isotope seasonality and the condensation height in temperate latitudes (e.g., Buenning et al., 2012; 2013).

2.5 Paleoclimate Records in Vicinity of Papua

Several studies have investigated the past climate and environmental history in the vicinity of Papua. Most of these studies were compiled to interpret the climate and environmental change in tropical Australasia over the last 35 kyr (Fig 2.9) (Reeves et al., 2013; 2013a). Only a few climate records have been reconstructed on the island of New Guinea, and none close to the glaciers near Puncak Jaya. Most paleoclimate studies in
this region were conducted at lower elevations close to the ocean (pollen and coral records), with the exception of the glacial extent records from moraines on the Mt. Giluwe, PNG (Barrows et al., 2011).

Figure 2.9 Locality of the sites of previous paleoclimate and paleoenvironmental studies in vicinity of Papua with various proxies. Puncak Jaya is marked by black triangle; Figure 1.A from Reeves et al. (2013)

In general, this region experienced a wetter and cooler climate since at least from ~53 to 30 kyrs BP based on carbon isotope composition of leaf waxes in lake sediment records from central Sulawesi (Wicaksono et al., 2015). In the early glacial period (~30 to 22 kyrs BP), proxy records suggest cooler and dryer conditions which peaked during the LGM (~22 to 18 kyrs BP) during maximum glacial extent across the region when SSTs averaged about 1 – 3 °C cooler than present. The deglaciation period (~18 to 12 kyrs BP) was first noted by warming in the Coral Sea (eastern Australia), then the Indonesian seas, with Indonesia experiencing wetter conditions by 17 kyrs BP, coincident with the beginning of the flooding of the Sunda Shelf. This was followed by wetter conditions in northern Australia after 14 kyrs BP, probably due to the flooding of the
Arafura shelf. In the early Holocene (~12 to 8 kyrs BP) the climate conditions were relatively warmer and wetter, which lasted until the mid-Holocene (~8 to 5 kyrs BP); while in the late Holocene (~5 to 0 kyrs BP) the climate was more variable and associated with the ENSO in El Niño mode (Reeves et al., 2013; 2013a).

On shorter timescales, most of the paleoclimate records from this region (such as coral and high-resolution pollen records) capture the ENSO signal. Most of the coral records from the equatorial western Pacific are good recorders of ENSO variability over the last two centuries (Tudhope et al., 2001; Morimoto et al., 2002; Asami et al., 2005; Quinn et al., 2006). The coral skeleton $\delta^{18}O$ is sensitive to SST due to temperature-dependent fractionation (Epstein et al., 1953; Kim and O'Neil, 1997), and the $\delta^{18}O$ of seawater variations which is directly related to changing seawater salinity due to kinetic fractionation during evaporation/condensation (Cole and Fairbanks, 1990; Fairbanks et al., 1997). In general, coral $\delta^{18}O$ records are positively correlated with sea surface salinity (SSS) and negatively correlated with SST while the coral Sr/Ca is only positively correlated with SST (e.g. Corrège, 2006; Cahyarini et al., 2009). When combined, the coupling of coral $\delta^{18}O$ and Sr/Ca records could yield more reliable SST and SSS reconstructions.

In addition, ENSO events are also recorded by a high-resolution pollen record from the uppermost section of a marine core covering the last 250 yrs from Kau Bay, Halmahera, Indonesia, northwest of Papua (van der Kaars et al., 2010). The high-resolution pollen record suggests that droughts occur during El Niño events, and elevated charcoal levels reflect a greater incidence of fire during these extremely dry periods.
3.1 Meteorological Data

The Indonesian Agency for Meteorology, Climatology and Geophysics (BMKG) has maintained several meteorological stations in West Papua and Papua provinces since the late-1970s or mid-1980s. However, most of those stations are located in lowland and coastal areas. The data are available on daily basis, but sometimes patchy, incomplete and not updated. In addition, PT Freeport Indonesia (PTFI), a gold mining company established in 1967 which is located to the south of the glacier area, has maintained an automatic weather station (AWS) network from the southern coast to the highlands near glaciers since 1997. Although some of the data are incomplete, they provide a relatively high temporally resolved (~15 minutes) dataset as well as data along a transect with increasing altitude from the coast to the highland. In addition, this study utilizes satellite-derived meteorological and reanalysis data to better understand both regional and large-scale variation of the atmosphere in this region.

3.1.1 Station Data

The BMKG is currently migrating the station meteorological data into an online database, which can be accessed at http://dataonline.bmkg.go.id/. There are 17
meteorological stations on the island of New Guinea as shown in Fig 3.1 and described in Table 3.1. Based on the current database, there are 6 stations with data since the late-1970s or mid-1980s (black circles in Fig 3.1) and 7 stations that have data since 2000s (red circles in Fig 3.1), while 4 stations are not used in this study as they are either short or patchy (yellow circles in Fig 3.1). The closest BMKG station to the glacier area is at Timika (37 masl), which has provided precipitation and temperature data since the mid-1980s. The seasonal variation and the monthly time series plots of temperature and precipitation for BMKG stations used in this study are given in Appendix B.

Figure 3.1 Location of the BMKG meteorological stations (No. 1 to 17). Description of BMKG station is given in Table 3.2. The black (red) circle indicates the stations that have data since the late 1970s or the mid-1980s (the 2000s) and the yellow indicates stations not used in the study. The closest NOAA radiosonde station to the glacier site is located in Biak (No. 6). The Global Network of Isotopes in Precipitation (GNIP) stations are located in Jayapura, Indonesia (No. 14) and Madang, Papua New Guinea (No. 18)
Table 3.1 Description of BMKG stations in New Guinea Island

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<th>Latitude</th>
<th>Altitude (masl)</th>
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<tr>
<td>12</td>
<td>Genyem</td>
<td>140.68 E</td>
<td>02.60 S</td>
<td>70</td>
</tr>
<tr>
<td>13</td>
<td>Sentani</td>
<td>140.55 E</td>
<td>02.55 S</td>
<td>99</td>
</tr>
<tr>
<td>14</td>
<td>Jayapura</td>
<td>140.72 E</td>
<td>02.87 S</td>
<td>3</td>
</tr>
<tr>
<td>15</td>
<td>Wamena</td>
<td>138.95 E</td>
<td>04.07 S</td>
<td>1660</td>
</tr>
<tr>
<td>16</td>
<td>Tanah Merah</td>
<td>140.30 E</td>
<td>06.10 S</td>
<td>16</td>
</tr>
<tr>
<td>17</td>
<td>Merauke</td>
<td>140.38 E</td>
<td>08.47 S</td>
<td>3</td>
</tr>
</tbody>
</table>
In addition, the location of the PTFI AWS network is shown in Fig. 3.2 and described in Table 3.2. The weather station at TPR was relocated to M66 on March 2008. Measured meteorological parameters include temperature, precipitation, relative humidity, wind speed and direction, radiation and pressure. The time series for each measurement varies both in time and with patchiness of the data. Data processing was conducted by removing outliers and calculating the climatological data for each parameter. In order to improve the climatological data, datasets from the previous study (Permana, 2011) were updated. The updated seasonal variation of temperature and precipitation and the monthly time series plots of meteorological variables for PTFI stations are given in Appendix C.

Figure 3.2 Weather station network along the south altitudinal transect of the glacier area in Papua maintained by PTFI. Description of stations is given in Table 3.2
Table 3.2 Description of stations in PTFI AWS network

<table>
<thead>
<tr>
<th>Station No</th>
<th>Station Name</th>
<th>Station Code</th>
<th>Lon</th>
<th>Lat</th>
<th>Alt (masl)</th>
<th>Period</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Portsite</td>
<td>PORT</td>
<td>136.84 E</td>
<td>04.83 S</td>
<td>9</td>
<td>01/1997 – 09/2015</td>
</tr>
<tr>
<td>2</td>
<td>Mile 21</td>
<td>M21</td>
<td>136.91 E</td>
<td>04.62 S</td>
<td>27</td>
<td>07/1997 – 09/2013</td>
</tr>
<tr>
<td>3</td>
<td>Timika</td>
<td>TMK</td>
<td>136.89 E</td>
<td>04.53 S</td>
<td>37</td>
<td>01/1997 – 12/2014</td>
</tr>
<tr>
<td>4</td>
<td>Kuala Kencana</td>
<td>KK</td>
<td>136.86 E</td>
<td>04.41 S</td>
<td>67</td>
<td>01/1997 – 12/2013</td>
</tr>
<tr>
<td>5</td>
<td>Mile 50</td>
<td>M50</td>
<td>137.01 E</td>
<td>04.28 S</td>
<td>617</td>
<td>01/1997 – 10/2011</td>
</tr>
<tr>
<td>6</td>
<td>Tembagapura</td>
<td>TPR</td>
<td>137.09 E</td>
<td>04.14 S</td>
<td>1900</td>
<td>01/1997 – 02/2008</td>
</tr>
<tr>
<td>7</td>
<td>Mile 66</td>
<td>M66</td>
<td>137.10 E</td>
<td>04.15 S</td>
<td>2350</td>
<td>03/2008 – 09/2015</td>
</tr>
<tr>
<td>8</td>
<td>Ridgecamp</td>
<td>RCMP</td>
<td>137.13 E</td>
<td>04.10 S</td>
<td>2410</td>
<td>01/1997 – 09/2013</td>
</tr>
<tr>
<td>9</td>
<td>Mile 74</td>
<td>M74</td>
<td>137.12 E</td>
<td>04.09 S</td>
<td>2750</td>
<td>05/2000 – 03/2013</td>
</tr>
<tr>
<td>10</td>
<td>Dispatch Tower</td>
<td>DISP</td>
<td>137.12 E</td>
<td>04.08 S</td>
<td>4109</td>
<td>01/1999 – 12/2014</td>
</tr>
<tr>
<td>11</td>
<td>Grasberg</td>
<td>GRS</td>
<td>137.11 E</td>
<td>04.07 S</td>
<td>3945</td>
<td>01/1997 – 09/2015</td>
</tr>
<tr>
<td>12</td>
<td>Alpine</td>
<td>ALP</td>
<td>137.12 E</td>
<td>04.04 S</td>
<td>4400</td>
<td>01/2000 – 04/2009</td>
</tr>
</tbody>
</table>

3.1.2 Radiosonde Data

The closest radiosonde data from glaciers near Puncak Jaya were downloaded from the NOAA Earth System Research Laboratory (ESRL) which is located at Biak Island in the northern Papua (No 6. in Fig 3.1). The observed variables include air temperature, geopotential height, wind speed and direction. The radiosonde data are
available from April 1994 to May 2011. However, the data are patchy and incomplete. The dataset can be downloaded from http://esrl.noaa.gov/raobs/.

3.1.3 **Reanalysis and Satellite Data**

The mean monthly anomalies of surface temperature and precipitation are from the NOAA Global Historical Climatology Network (GHCN) that is available from January 1880 (1900) for temperature (precipitation) to May 2015 with 5° grid resolution (http://www.esrl.noaa.gov/psd/data/gridded/data.ghcngridded.html). At higher resolution, meteorological data were downloaded from the NOAA - National Center for Environmental Prediction/ National Center for Atmospheric Research (NCEP/NCAR) reanalysis data that are available since 1948 with 2.5° grid resolution (http://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis.derived.html) and the NOAA 20th century reanalysis with 2° grid resolution that are available from 1871 to 2012 (http://www.esrl.noaa.gov/psd/data/gridded/data.20thC_ReanV2.html). Over shorter time periods and higher resolutions, meteorological values were acquired from the NASA- Modern Era Retrospective-analysis for Research and Applications (MERRA) reanalysis data, which are available from 1979 to present with a resolution of 0.67° longitude and 0.5° latitude (Bosilovich, 2008; NASA, 2011) (http://disc.sci.gsfc.nasa.gov/mdisc/data-holdings).

Regional precipitation data were downloaded from the National Oceanic and Atmospheric Administration (NOAA) - Global Precipitation Climatology Project (GPCP) version 2.2 with 2.5° grid resolution that are available since January 1979 to July 2015 (http://www.esrl.noaa.gov/psd/data/gridded/data.gpcp.html), and the National
Aeronautics and Space Administration (NASA) - Tropical Rainfall Measuring Mission (TRMM) satellite data 3B43 V7 with 0.25° grid resolution (http://disc2.nascom.nasa.gov/opendap/TRMM_L3/TRMM_3B43/contents.html). While, regional daily precipitation data were obtained from TRMM 3B42 V7 product (http://disc2.nascom.nasa.gov/opendap/TRMM_L3/TRMM_3B42_daily/contents.html).

The TRMM 3B42 products are available from January 1, 1998 to present and will continue to be produced through mid-2017. The vertical profile of monthly latent heating in the troposphere is derived from the TRMM product 3A12 V7 with 0.5° grid resolution (http://disc2.nascom.nasa.gov/opendap/TRMM_L3/TRMM_3A12/contents.html).

Unfortunately, the TRMM 3A12 products are only available up to March 2015 as the instruments on TRMM were turned off on April 8, 2015.

Regional convective activity data were obtained from the NOAA interpolated OLR dataset on a 2.5° grid resolution (Liebmann and Smith, 1996) (http://www.esrl.noaa.gov/psd/data/gridded/data.interp_OLR.html). Unfortunately, the NOAA OLR data are only available from June 1, 1974 to December 31, 2013 due to limited available funding. At higher resolution, OLR data were derived from MERRA reanalysis as a variable of upward longwave flux at the top of the atmosphere from 1979 to present with a resolution of 0.67° longitude and 0.5° latitude (http://goldsmr2.sci.gsfc.nasa.gov/dods/MATMNXRAD for monthly data; http://goldsmr2.sci.gsfc.nasa.gov/dods/MAT1NXRAD for hourly data).
3.2 Climate Indices

In this study, the regional climate indices from the NOAA National Weather Service (NWS) Climate Prediction Center (CPC) were used as the SST-based ENSO indicators. In order to obtain long-term climate indices, the NOAA Extended Reconstructed Sea Surface Temperature version 4 (ERSSTv4) dataset was used to provide monthly SSTs from January 1854 to May 2015 with a 2°x 2° grid (Huang et al., 2015; Liu et al., 2015) (http://www.esrl.noaa.gov/psd/data/gridded/data.noaa.ersst.v4.html). These indices were calculated based on the monthly average SST values for specific regions of the tropical Pacific Ocean, which are 5° N–5° S and 150°–90° W for NINO3, 5° N–5° S and 170°–120° W for NINO3.4, and 5° N–5° S and 160° E–150° W for NINO4 (http://www.cpc.ncep.noaa.gov/data/indices/).

The PDO is a long-lived El Niño-like pattern of climate variability in the Pacific Ocean. The monthly PDO indices were provided by the Joint Institute for the Study of the Atmosphere and Ocean (JISAO) at University of Washington and calculated based on Mantua et al. (1997). Their data sources were the UK Met-Office (UKMO) Historical SST dataset for 1900–1981, Reynold’s Optimally Interpolated (OI) SST (V1) for January 1982 to December 2001, and the OISSTV2 for January 2002 to the present (http://research.jisao.washington.edu/pdo/PDO.latest).

3.3 Rainfall Stable Isotope Data

Rainfall stable isotope data from New Guinea are available from stations of the International Atomic Energy Agency (IAEA) and the World Meteorological Organization
(WMO) through the Global Network of Isotopes in Precipitation (GNIP) (IAEA, 2006). The stations are located in Jayapura, Indonesia (2.53°S; 140.72°E; 3 masl) and Madang, PNG (5.22°S; 145.80°E; 49 masl) (No 14. and No 18. in Fig. 3.1). The monthly isotope data are available from 1957 to 1991 for Jayapura and from 1968 to 1982 for Madang. However, the data are incomplete at both stations. The isotopic data consist of δ^{18}O, δD, d and tritium (³H) concentration. Meteorological data are also available from these stations, including precipitation, temperature and vapor pressure.

During the ice-core drilling expedition in June 2010, 83 rain samples from different elevations were collected daily over a two week period (Permana, 2011). For this study, a total of 1362 samples were collected daily at 7 - 8 AM local time over two periods. The first period was from January 2013 to February 2014 which represents an ENSO-normal year condition. In this period, 923 samples were collected at five locations in close proximity to AWS operated by PTFI (PORT, TMK, KK, TPR and GRS, see Fig. 3.2) (Thompson et al., 2014). Only 899 samples were used in this study because 16 samples lacked proper documentation of the collection dates, and 8 samples from TMK were invalidated by sampling errors (Table 3.3). Details of the sample collection process and description of the rain collector are described in Permana (2011). Sample collection at TMK, KK and TPR began on January, 29 2013, while at PORT it began on January 26, and at GRS it began on January, 30. The fewest samples were collected at GRS from May to June 2013 due to a tunnel collapse at the PTFI training facility in mid-May 2013.

The second collection period was conducted from December 2014 to September 2015 during which 439 samples were collected at three stations (PORT, TPR and GRS)
with 1 sample from TPR and 5 samples from GRS were invalidated by sampling errors (Table 3.3). Collection at PORT and TPR started on December, 11 2014, while at GRS it started on December, 13 2014. The collection during this period was designed to capture the signature of the strong 2014/2016 El Niño event in the Papua rainfall isotopic values. Therefore, the second period represents an El Niño year condition.

Table 3.3 Number of rain samples collected at some PTFI stations. The heavy line splits the two collection periods (ENSO-normal and El Niño)

<table>
<thead>
<tr>
<th>Month</th>
<th>PORT</th>
<th>TMK</th>
<th>KK</th>
<th>TPR</th>
<th>GRS</th>
<th>Total</th>
</tr>
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<td>Feb 2013</td>
<td>15</td>
<td>18</td>
<td>24</td>
<td>18</td>
<td>18</td>
<td>93</td>
</tr>
<tr>
<td>Mar 2013</td>
<td>15</td>
<td>8</td>
<td>25</td>
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<td>6</td>
<td>70</td>
</tr>
<tr>
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<td>20</td>
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<td>28</td>
<td>25</td>
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</tr>
<tr>
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<td>19</td>
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<tr>
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<td>25</td>
<td>24</td>
<td>2</td>
<td>82</td>
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<tr>
<td>Jul 2013</td>
<td>25</td>
<td>16</td>
<td>31</td>
<td>27</td>
<td>22</td>
<td>121</td>
</tr>
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<td>27</td>
<td>11</td>
<td>79</td>
</tr>
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<td>29</td>
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<td>0</td>
<td>65</td>
</tr>
<tr>
<td>Jan 2014</td>
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<td>0</td>
<td>2</td>
<td>20</td>
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<td>0</td>
<td>0</td>
<td>5</td>
<td>0</td>
<td>5</td>
</tr>
<tr>
<td>Dec 2014</td>
<td>11</td>
<td></td>
<td></td>
<td>18</td>
<td>7</td>
<td>36</td>
</tr>
<tr>
<td>Jan 2015</td>
<td>18</td>
<td></td>
<td></td>
<td>25</td>
<td>18</td>
<td>61</td>
</tr>
<tr>
<td>Feb 2015</td>
<td>10</td>
<td></td>
<td></td>
<td>22</td>
<td>24</td>
<td>56</td>
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<tr>
<td>Mar 2015</td>
<td>11</td>
<td></td>
<td></td>
<td>20</td>
<td>16</td>
<td>47</td>
</tr>
<tr>
<td>Apr 2015</td>
<td>19</td>
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<td>20</td>
<td>23</td>
<td>62</td>
</tr>
<tr>
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<td>15</td>
<td>3</td>
<td>34</td>
</tr>
<tr>
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<td>23</td>
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<td>44</td>
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<tr>
<td>Jul 2015</td>
<td>14</td>
<td></td>
<td></td>
<td>13</td>
<td>10</td>
<td>37</td>
</tr>
<tr>
<td>Aug 2015</td>
<td>11</td>
<td></td>
<td></td>
<td>10</td>
<td>9</td>
<td>30</td>
</tr>
<tr>
<td>Sep 2015</td>
<td>10</td>
<td></td>
<td></td>
<td>16</td>
<td>0</td>
<td>26</td>
</tr>
<tr>
<td>Total</td>
<td>333</td>
<td>91</td>
<td>296</td>
<td>437</td>
<td>175</td>
<td>1332</td>
</tr>
</tbody>
</table>
The $\delta^{18}$O and $\delta$D of rain samples were measured at the Byrd Polar and Climate Research Center (BPCRC) stable isotope laboratory using a cavity ring-down spectrometer (Picarro L2120-i water isotope analyzer). The reported estimates of uncertainty attributed to $\delta^{18}$O and $\delta$D measurements are $\pm0.2\%$ and $\pm1.5\%$, respectively. An uncertainty of $\pm1.5\%$ for the $d$ value is calculated from the quadratic combination of the uncertainties of $\delta^{18}$O and $\delta$D.

3.4 Ice Core Data

The Papua ice cores were cut into discrete samples, and the melted ice was analyzed for oxygen and hydrogen isotope compositions, concentrations of insoluble dust particles of major anions and cations. All of these measurements were conducted at BPCRC laboratories. The oxygen and hydrogen isotopic ratios were measured using Thermo Finnigan mass spectrometer and Picarro cavity ring-down spectroscopy. Results of identical samples analyzed in both machines were compared and were found to be in good agreement (Permana, 2011). The concentrations of insoluble dust particles greater than 0.63μm in diameter were measured using a Beckman-Coulter Multisizer 4. The concentrations of the major ions (cations: ammonium ($NH_4^+$), sodium ($Na^+$), potassium ($K^+$), magnesium ($Mg^{2+}$), and calcium ($Ca^{2+}$) and anions: sulfate ($SO_4^{2-}$), nitrate ($NO_3^-$), fluoride ($F^-$), and chloride ($Cl^-$)) were determined using a Dionex ICS-3000 ion chromatography system. Descriptions of the sample handling procedures and data collection are provided in Permana (2011).
The longest core (D1) was sectioned into 1156 co-registered samples (~2.8 cm resolution) which were analyzed for stable isotopes, dust and soluble aerosols, and 136 lower-resolution samples that were prepared for a preliminary stable isotope survey (Thompson et al., 2014). In addition, 68 samples from Core D1 were measured for tritium concentrations at the Division of Climate and Environmental Physics, Physics Institute, University of Bern, Switzerland to locate the 1960s bomb horizons. Also, 6 samples were measured for trace and rare earth elements. Core D1B contains 1607 co-registered samples (~2 cm resolution) which were also analyzed for stable isotopes and dust, and 644 samples of ~5 cm length were analyzed for major ion concentrations. Initially, 208 samples were prepared and measured for a preliminary stable isotope survey. Stable isotopes have been measured on 325 samples from Core D2 (~8 cm resolution).
As discussed in Section 2.4, precipitation amount and OLR (a parameter indicating convective activity) are generally inversely correlated on seasonal timescales in the Tropics. In the Maritime Continent region, seasonal rainfall isotopic variations are linked to local or regional precipitation amounts (Kurita et al., 2009), and to moisture origin (Suwarman et al., 2013) and are influenced by the strength of regional convective activity (Moerman et al., 2013; Kurita, 2013). However, most of those studies were conducted at locations where precipitation amount and OLR values are negatively correlated on seasonal timescales.

The southern slope of the central mountain ranges of Papua has a unique climate regime because the precipitation amount and OLR are positively correlated on seasonal timescales (Fig 4.1), in contrast to the typical relationship in the Tropics. Due to its location in the South Hemisphere, the annual movement of the ITCZ causes low (high) OLR values in the region during austral summer (winter) from December to March (May to October). However, Aldrian and Susanto (2003) identified the rainfall seasonality in this region as anti-monsoonal, marked by a distinct peak during austral winter and relatively low rainfall during austral summer. Therefore, it is important to directly
evaluate the influence of precipitation amount and convective activity on the stable isotope ratios of tropical rainfall in the region.

Figure 4.1 Correlation between seasonal NOAA OLR and GPCP precipitation (long term mean 1981 - 2010) over Maritime Continent. Red lines mark the boundaries of the 95% significance levels

As discussed in Section 3.3, 1362 rain samples that were collected daily during the period of January 2013 to February 2014 (ENSO-normal year) and December 2014 to September 2015 (El Niño year) at different elevations near the PTFI weather stations were analyzed to investigate the controls on temporal variations of stable isotopes from daily to interannual timescales. This chapter discusses the stable isotope analyses of these rain samples and their relationship to local surface temperatures, local and regional precipitation, and convective activity based on *in situ* and satellite data. The isotope results from the first and second periods are then compared to investigate the impact of El
Niño on rainfall isotopes in Papua on interannual timescales. The influences of moisture origins and transport history on rainfall isotopic values are examined based on air mass trajectories. Finally, the influence of atmospheric temperature at mean condensation level on rainfall isotopic values is examined based on latent heat (LH) release and cloud fraction in the troposphere.

4.1 Stable Isotopic Variations

Daily rainfall $\delta^{18}$O, $\delta$D and deuterium excess ($d = \delta$D– 8$\delta^{18}$O) values for each site from two collection periods are shown in Fig 4.2. Detailed statistics of rainfall isotopic values are described in Table 4.1. The results show that the more depleted $\delta^{18}$O and $\delta$D values occur with increasing elevation, which demonstrates the nature of the altitude effect. The mean isotopic lapse rates are -2.4‰/km for $\delta^{18}$O and -18.2‰/km for $\delta$D. These values are very comparable to the mean isotopic lapse rate of -2.3‰/km for $\delta^{18}$O and -18.0‰/km for $\delta$D during the ~2 week collection period in June 2010 (Permana, 2011), as well as to the global isotopic lapse rate of approximately -2.8‰/km for $\delta^{18}$O (Poage and Chamberlain, 2001). This altitude effect is equivalent to a temperature effect of 0.48‰/°C for $\delta^{18}$O and 3.64‰/°C for $\delta$D, based on the surface lapse rate of 5.0°/km (Permana, 2011). In general, all sites show a similar temporal variation of daily rainfall $\delta^{18}$O, which suggests a common regional controlling factor of the temporal variability of stable isotopes of rainfall in the region.
Figure 4.2 Timeseries of daily rainfall stable isotopes for (a) δ^{18}O, (b) δD, and (c) d for each station. Station codes are described in Table 3.2. Dashed lines in (a) represent the linear trends over particular time intervals.
Figure 4.2 Continued from previous page

(b)
Figure 4.2 Continued from previous page
Table 4.1 Descriptive statistics of daily rainfall $\delta^{18}$O, $\delta$D and $d$ values for each collection station and period. Station codes are described in Table 3.2

<table>
<thead>
<tr>
<th>Station Code</th>
<th>Jan 2013 - Feb 2014 (ENSO-normal year)</th>
<th>Dec 2014 - Sep 2015 (El Niño year)</th>
<th>All samples</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Min</td>
<td>Mean</td>
<td>Amount-Weighted</td>
</tr>
<tr>
<td>$\delta^{18}$O (%)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PORT (9 masl)</td>
<td>-15.35</td>
<td>-7.32</td>
<td>-6.78</td>
</tr>
<tr>
<td>TMK (37 masl)</td>
<td>-14.99</td>
<td>-7.89</td>
<td>-7.84</td>
</tr>
<tr>
<td>KK (67 masl)</td>
<td>-17.50</td>
<td>-8.11</td>
<td>-8.28</td>
</tr>
<tr>
<td>TPR (1900 masl)</td>
<td>-21.72</td>
<td>-12.07</td>
<td>-12.02</td>
</tr>
<tr>
<td>$\delta$D (%)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PORT (9 masl)</td>
<td>-107.30</td>
<td>-44.33</td>
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</tr>
<tr>
<td>TMK (37 masl)</td>
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<td>KK (67 masl)</td>
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<tr>
<td>TPR (1900 masl)</td>
<td>-158.62</td>
<td>-81.66</td>
<td>-81.19</td>
</tr>
<tr>
<td>GRS (3945 masl)</td>
<td>-199.02</td>
<td>-119.15</td>
<td>-124.34</td>
</tr>
<tr>
<td>$d$ (%)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PORT (9 masl)</td>
<td>5.88</td>
<td>14.24</td>
<td>14.30</td>
</tr>
<tr>
<td>TMK (37 masl)</td>
<td>6.78</td>
<td>14.52</td>
<td>15.62</td>
</tr>
<tr>
<td>KK (67 masl)</td>
<td>4.17</td>
<td>13.43</td>
<td>13.35</td>
</tr>
<tr>
<td>TPR (1900 masl)</td>
<td>7.48</td>
<td>14.92</td>
<td>14.94</td>
</tr>
<tr>
<td>GRS (3945 masl)</td>
<td>10.10</td>
<td>17.57</td>
<td>17.39</td>
</tr>
</tbody>
</table>
The mean weighted \(d\) values range from 13.35‰ at KK to 17.10‰ at GRS (Table 4.1; see Table 3.2 for station description), which demonstrate an increase of the \(d\) value with elevation at a rate of 0.68‰/km. An increase of \(d\) value with elevation has also been reported in tropical montane rainforest of Amazon basin (e.g. Windhorst et al., 2013). The \(d\) value tends to increase with elevation at high relative humidity (Gonfiantini et al., 2001), which prevails in the Papua region. The mean \(d\) value of all samples is ~15‰, which is greater than the GMWL intercept of 10‰ (Craig, 1961). The higher \(d\) value in this region may be due to a land surface recycling through the surrounding tropical rainforests during transport to the collection sites (Salati et al., 1979; Windhorst et al., 2013). As with \(\delta^{18}\)O, the daily \(d\) variations at all stations show a similar pattern suggesting relatively similar moisture sources and transport pathways into the region (Fig 4.2c).

In the first collection period (ENSO-normal year), the most enriched \(\delta^{18}\)O (\(\delta\)D) of -0.71‰ (6.22‰) was recorded at KK, while the most depleted \(\delta^{18}\)O (\(\delta\)D) of -27.10‰ (-199.02‰) was recorded at GRS. The mean weighted \(\delta^{18}\)O ranges from -6.78‰ at the lowest elevation (PORT) to -17.72‰ at the highest elevation (GRS), while the weighted \(\delta\)D ranges from -39.93‰ to -124.34‰ at these same stations. The ranges of \(\delta^{18}\)O and \(\delta\)D among stations are 11.93‰ to 17.96‰ and 104.23‰ to 149.18‰, respectively (Table 4.1). The linear trends of daily rainfall \(\delta^{18}\)O at the four lowest stations are relatively stable between February and April 2013, positive from May to August 2013, negative from September to December 2013 and positive again from January to February 2014. Due to lack of data, daily rainfall \(\delta^{18}\)O variation at GRS appears to follow this similar
trend during the first collection period, which was later confirmed during the second collection period. In the meantime, there is a little linear trend in daily $d$ values on seasonal timescales (Fig 4.2c).

In the second collection period (El Niño year), the amount-weighted stable isotope signatures ($\delta^{18}O$ and $\delta D$) in rain samples are higher than during the first period (ENSO-normal) at all stations. While, the $d$ values are lower than during the ENSO-normal period, they are insignificant because the differences are within the $\pm 1.5\%$ uncertainty. In addition, the ranges of $\delta^{18}O$ and $\delta D$ values are similar for El Niño and ENSO-normal periods (Table 4.1).

### 4.2 Meteoric Water Lines (MWLs)

By considering all samples, the local meteoric water lines (LMWLs) for each station are given in Fig 4.3. The slopes of the LMWLs range between 7.90 and 8.54, which are comparable to the GMWL slope of 8. However, the intercept values ($d$) range between 14.27‰ and 19.04‰, which are higher than the GMWL value of 10. By considering samples from all stations, the regional MWL (RMWL) is calculated as

$$\delta D = 7.98 \delta^{18}O + 14.38 \ (n = 1332, R^2 = 0.99).$$

Fig 4.4 compares the LMWLs at PORT, TPR and GRS stations, and all together during ENSO-normal period and El Niño period. There are no significant changes in the MWL slopes and intercepts observed at these three stations during El Niño period as compared to the ENSO-normal period. As a result, considering all samples from these
three stations, the RMWL slope (intercept) remains similar for the El Niño period 7.94 (14.38‰) and for the ENSO-normal period 7.99 (14.86‰).

Figure 4.3 Linear relationships between $\delta^{18}$O and $\delta$D values at each station, and from all stations. Black and red lines represent the LMWL and GWML, respectively.
Figure 4.4 Comparison of MWLs at PORT, TPR, GRS stations and of all samples from those stations between ENSO-normal period (left) and El Niño period (right). Black and red lines represent the LMWL and GWML, respectively.
4.3 Controls on Rainfall $\delta^{18}$O

4.3.1 Daily Variation

Due to a similar daily isotopic variation among stations, daily rainfall $\delta^{18}$O from all stations are amount-weighted, composited and then linearly interpolated to obtain the daily regional rainfall $\delta^{18}$O (Fig 4.5). Based on statistical analyses, there are no significant correlations between daily rainfall $\delta^{18}$O and daily local precipitation amount at three low elevation stations (Fig 4.6a-c), while at the mid and high elevation stations, the correlations are significantly negative but weak (Fig 4.6d-e). At the regional scale, no correlation exists between the daily regional rainfall $\delta^{18}$O and daily regional precipitation amount (mean precipitation from TRMM at 4.375 - 4.875°S and 136.875°E) ($R = -0.091$, $p = 0.02$; Fig 4.6f). This suggests that the local/regional precipitation amount does not exert a major control over isotope ratios in the area on a daily basis except at the mid and high elevation stations.

On the other hand, the correlation coefficients between daily rainfall $\delta^{18}$O and daily local surface temperature at all stations are significantly negative but weak (Fig 4.7a-e). These inverse relationships suggest that local surface temperature is also not the main driver of rainfall $\delta^{18}$O in the region. Instead, a statistically significant, but weak, positive relationship ($R = 0.46$, $p < 0.001$; Fig 4.7f) is observed between daily regional rainfall $\delta^{18}$O and daily regional convective activity (mean OLR values from MERRA at 4.5 - 5.0°S, and 136.67°E), suggesting that convective activity is likely the main control of rainfall $\delta^{18}$O in the region.
Figure 4.5 The composite of daily rainfall $\delta^{18}O$ during the two collection periods. The interpolated daily rainfall $\delta^{18}O$ composite (blue) represents the daily regional rainfall $\delta^{18}O$ in this study.

Figure 4.6 Relationships between daily rainfall $\delta^{18}O$ and daily local precipitation amount at (a) PORT, (b) TMK, (c) KK, (d) TPR and (e) GRS. The relationship between daily regional rainfall $\delta^{18}O$ and daily regional precipitation (mean precipitation from TRMM at 4.375-4.875°S, 136.875°E) is shown in (f). Regression lines are in black.
Figure 4.7 As in Fig 4.6, but with daily local surface temperature for (a) to (e), while (f) shows the relationship between daily regional rainfall $\delta^{18}O$ and daily regional convective activity (mean OLR values from MERRA at 4.5-5.0°S, 136.67°E).

Fig 4.8 compares the time series of daily regional rainfall $\delta^{18}O$, regional precipitation amount and regional convective activity for the two collection periods. The positive correlation between daily regional rainfall $\delta^{18}O$ and OLR values increases from 0.22 ($p < 0.01$) during the ENSO-normal period (2013/2014) to 0.61 ($p < 0.01$) during the El Niño period (2014/2015). However, the daily regional rainfall $\delta^{18}O$ only has a significant negative correlation with regional precipitation amount during El Niño period ($R = -0.25; p < 0.01$) and, not during the ENSO-normal period ($R = 0.08; p = 0.13$). This negative relationship is likely due to extreme dry conditions over Papua during the El Niño event, particularly in austral winter (Sukri et al., 2003; Prentice and Hope, 2007).
Figure 4.8 Comparison between (top) time series of daily regional rainfall δ^{18}O and regional convective activity (mean OLR values from MERRA at 4.5-5.0°S, 136.67°E), and (bottom) regional precipitation amount (TRMM values at 4.375-4.875°S, 136.875°E)

To investigate the relationship between δ^{18}O and climate variables on longer timescales, the correlations between running means of daily local/regional rainfall δ^{18}O and convective activity (OLR) as well as with precipitation amount are given in Fig 4.9a.
and 4.9b during the first and second collection periods, respectively. With increased day averaging, correlations between the regional δ¹⁸O and regional convective activity are stronger during the two collection periods, suggesting a major control of convective activity on rainfall δ¹⁸O at longer timescales. In contrast, negative correlations between rainfall δ¹⁸O and local/regional precipitation amount are stronger only during the El Niño event (second period), whereas during the first period, there are no correlations between them up to 10-day running means. Interestingly, at greater than 10-day running means weak significant positive correlations are found which range from 0.11 (for 10-day running means) to 0.35 (for 60-day running means). Furthermore, during the first period, correlations between local rainfall δ¹⁸O and local precipitation remain positive at lowland stations, are insignificant at middle elevation station and are negative at highland station at longer timescales (up to 60-day running means).

In addition, convective process may have the temporal integration effect on rainfall δ¹⁸O over several previous days, as observed in other monsoon regions (e.g. Risi et al., 2008; Vimeux et al., 2011; Tremoy et al., 2012; Moerman et al., 2013; Gao et al., 2013). To assess the temporal integration effect of convective activity and precipitation, Fig 4.9c and 4.9d show the correlations between daily local/regional δ¹⁸O and running means of OLR and precipitation amount during the two collection periods. Daily regional rainfall δ¹⁸O have a stronger positive correlation with OLR averaged over the previous 5–7 days ($R = 0.43, p<0.01$ in the first period and $R = 0.78, p<0.01$ in the second period). The peaks of positive correlation are also observed with OLR averaged over the previous 20 and 60 days in the first collection period but not in the second collection period.
Figure 4.9 (a) Correlations between 'n'-days in running means of daily rainfall $\delta^{18}O$ and daily climate variables for 2013/14 collection period. (b) As in (a), but for 2014/15 collection period. (c) Correlations between daily rainfall $\delta^{18}O$ and 'n'-previous days average of climate variables for 2013/14 collection period. (d) As in (c), but for the 2014/15 collection period. Climate variables include local precipitation, regional precipitation (TRMM-derived) and regional convective activity (OLR data).
On the other hand, the negative correlations are also stronger and significant between daily rainfall $\delta^{18}$O and regional precipitation averaged over the previous 5 – 9 days ($R = -0.18, p< 0.01$ in the first period and $R = -0.52, p< 0.01$ in the second period). The most significant negative correlations between daily local $\delta^{18}$O and running means of local precipitation at most stations occur over the previous 3 to 15 days averaging in both collection periods, with the exception of PORT station which has a significant positive correlation with local precipitation over the previous 6 or more days averaged during the first collection period.

These analyses agree with the time-integrative nature of rainfall $\delta^{18}$O and support the hypothesis of the week-long 'memory' of rainfall $\delta^{18}$O, which may reflect the approximately week-long atmospheric residence time of water vapor in the region (Moerman et al., 2013). The integrative convection property of $\delta^{18}$O has been explained by the progressive isotopic depletion of local water vapor by the downdraft recycling process in the convective system. Furthermore, the shorter memory effect during the rainy season has been related to the more intensive and frequent convective activity in the region (Vimeux et al., 2011).

In addition, the spatial integrative effects of convective activity and precipitation on rainfall $\delta^{18}$O variation are examined based on air mass trajectories. Daily air mass back trajectories during January - December 2013 and January - August 2015 were calculated using the Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPLIT) back trajectory model (Draxler and Rolph, 2014) combined with NCEP reanalysis (Fig 4.10 and Fig 4.11) and Global Data Assimilation System (GDAS) (not shown) data at 12-
hour time steps back to 5 previous days at 1,000, 2,000 (not shown) and 4,000 (not shown) meters above ground level (magl).

To detect the influence of spatial-integration of regional convection and precipitation on rainfall $\delta^{18}O$, the correlation coefficients between daily regional rainfall $\delta^{18}O$ and daily average OLR (MERRA), as well as with the cumulative regional precipitation (TRMM) were calculated along the back trajectories. Fig 4.12 depicts the correlation coefficients of spatial-integrated convective activity and cumulative precipitation along the back trajectories for 2013 and 2015 with different driving meteorology data. Both parameters have significant effects on rainfall $\delta^{18}O$ variation, but convective activity is more strongly correlated than precipitation. The correlation coefficients over the 5 previous days are comparable to the temporal integrative effect of convective activity and precipitation on rainfall $\delta^{18}O$ (Fig 4.9b,d), suggesting a significant spatio-temporal integrated effect of convective activity and precipitation in the region. Although the correlation between daily rainfall $\delta^{18}O$ and precipitation is stronger when precipitation data are accumulated along the back trajectories from the upwind region, it is still weaker than the spatially integrated effect of convective activity on rainfall $\delta^{18}O$, indicating that convective activity is more important than precipitation amount in controlling rainfall $\delta^{18}O$. The results are consistent with previous studies (Gao et al., 2013; Samuels-Crow et al., 2014; Lekshmy et al., 2014) which indicate the upwind regional convective activity is an important controlling factor on rainfall $\delta^{18}O$ variation in the Tropics.
Figure 4.10 Daily air mass back trajectories from 5 previous days to the rain collection stations simulated by HYSPLIT model combined with NCEP reanalysis data at 1,000 magl from January to December 2013 (a-l)
Figure 4.11 As of Fig 4.10, but for period of January to August 2015
Figure 4.12 Correlation coefficients between daily regional rainfall $\delta^{18}O$ and mean daily OLR (MERRA), and with cumulative precipitation (TRMM) along the back trajectories over 'n'-previous days with different starting elevations during 2013 using (a) NCEP Reanalysis and (b) GDAS data. (c) As of (a) and (d) as of (b), but during 2015
Consistently positive relationships between rainfall $\delta^{18}$O and OLR on daily and longer timescales during ENSO-normal and El Niño conditions indicate that regional convective activity explains a larger portion of rainfall $\delta^{18}$O variability. Furthermore, spatio-temporal integrative convection over 5 – 7 days seems to be the strongest control in the region. In contrast, inconsistent temporal relationships between local/regional precipitation amount and rainfall $\delta^{18}$O indicate that precipitation amount is not a major control on rainfall $\delta^{18}$O variability. These results also suggest that large-scale atmospheric processes are important factors that influence the rainfall $\delta^{18}$O variability, with convective activity is more important than precipitation along the trajectories to the precipitation sites.

### 4.3.2 Intraseasonal Variation

Previous studies (Risi et al., 2008; Vimeux et al., 2011; Tremoy et al., 2012; Gao et al., 2013; Moerman et al., 2013) have demonstrated correlations between tropical rainfall $\delta^{18}$O and large-scale convective activity on intraseasonal timescales. Kurita et al. (2011) showed that rainfall isotopic variations in the Maritime Continent Indonesia are associated with the Madden-Julian Oscillation (MJO) intraseasonal cycle. The regional $\delta^{18}$O is highly correlated with regional convective activity (OLR) over increasing time averages such that the $^{18}$O depletions are mostly associated with low OLR values (Fig 4.8a,c). The regional $\delta^{18}$O time series is characterized by peak-trough variations with an average amplitude of ~5‰ with a periodicity of ~20 - 60 days which resembles the MJO periodicity in the region (Fig 4.13). This is consistent with the findings of Moerman et al. (2013) in northern Borneo (4°N, 114°E), northwest of Papua. In their study, the authors
concluded that the MJO strongly influences the intraseasonal variability of rainfall $\delta^{18}O$, with major depletion events (with $\delta^{18}O$ shifts up to $\sim 10\%$) coinciding with the active (wet) phases of the MJO.

Because Papua is also located within the MJO influenced area, major $\delta^{18}O$ depletion events in this region are also likely associated with the active phases of the MJO. Fig 4.14 shows that major regional $\delta^{18}O$ depletion events in the region mostly coincide with the enhanced regional convective activity (low OLR values) during the eastward propagation in the tropical region.
4.3.3 Seasonal Variation

The seasonal amount-weighted $\delta^{18}$O values during the two collection periods at most stations are generally marked by austral summer depletion (December - April) and austral winter to spring (June - October) enrichment; however, an enrichment is also observed in March (Fig 4.15a-e). The seasonal correlation between monthly $\delta^{18}$O and
local precipitation amount varies among the stations; they are insignificantly positive at TMK ($R = 0.51, p = 0.16$) and KK ($R = 0.46, p = 0.14$), while there is no correlation at PORT ($R = -0.02, p = 0.95$), and they are negatively significant at TPR ($R = -0.76, p = 0.004$), but insignificant at GRS ($R = -0.29, p = 0.44$). On the other hand, the correlations with surface temperatures are negative and significant at all stations (PORT ($R = -0.65, p = 0.02$); TMK ($R = -0.57, p = 0.11$); KK ($R = -0.59, p = 0.04$); TPR ($R = -0.60, p = 0.04$); and GRS ($R = -0.51, p = 0.16$)). Regionally, the seasonal variation of the monthly regional amount-weighted $\delta^{18}O$ resembles the seasonal variation of regional convective activity with a significantly positive correlation ($R = 0.80, p = 0.002$; Fig 4.15f).

Meanwhile, insignificant negative relationship is observed between monthly regional amount-weighted $\delta^{18}O$ and TRMM regional precipitation on seasonal timescales ($R = -0.30, p = 0.35$; Fig 4.15f). These results suggest that regional convective activity play more important key role in controlling rainfall $\delta^{18}O$ than local/regional precipitation amount on seasonal timescales.

In addition, a comparison among three different elevation stations (PORT, KK and TPR) that lie within a range of ~80 km indicates that the annual precipitation amount is not correlated with the mean annual $\delta^{18}O$ in 2013; instead the altitude effect associated with the temperature effect is predominant (Fig 4.16).
Figure 4.15 Seasonal amount-weighted $\delta^{18}O$, precipitation amount and temperature at (a) PORT, (b) TMK, (c) KK, (d) TPR, and (e) GRS stations. (f) is seasonal regional amount-weighted $\delta^{18}O$, regional convective activity (OLR), and regional precipitation (TRMM) during the two collection periods.
The annual movement of the ITCZ is the main driver of the seasonal convective activity in the Tropics. Its position in the austral summer (monsoon season) causes a low pressure system in the corresponding hemisphere which leads to the formation of deep convection that generates high precipitation amount. The results of this study are consistent with previous studies that suggest that local/regional precipitation amount is not the main control on seasonal rainfall $\delta^{18}$O variation, but rather depends on large-scale convective activity (Kurita, 2013; Lekshmy et al., 2014). For instance, mean monthly rainfall at KK in July and August (austral winter) during collection period was 1,533 and 1,360 mm, respectively, with associated monthly weighted $\delta^{18}$O of -7.6 and -5.2 ‰, respectively, and associated OLR values of 220 Wm$^{-2}$ and 225 Wm$^{-2}$, respectively. Meanwhile, in January and December (austral summer) mean monthly rainfall was 320
and 575 mm, respectively, with associated monthly weighted $\delta^{18}O$ of -12.9 ‰ and -12.2 ‰, respectively, and associated OLR values of 182 Wm$^{-2}$ and 186 Wm$^{-2}$, respectively (Fig 4.15c). Large amounts of rainfall that occur from June to October (austral winter) at lowland stations are likely not generated by deep enhanced convection with higher and lower temperature clouds in low pressure system, but are likely produced by shallow convection due to local wind circulation and topographic setting where the cloud formation is capped by the Trade Wind Inversion layer in high pressure system.

### 4.3.4 Interannual Variation

The comparison of monthly amount-weighted isotopic values between the ENSO-normal period and the El Niño period shows that the $\delta^{18}O$ ($\delta D$) has increased by ~1.6‰ (~11.4‰) in El Niño period at PORT during December to September. At TPR, $\delta^{18}O$ ($\delta D$) has also increased by ~1.8‰ (~14.8‰) during December to September, while at GRS, $\delta^{18}O$ ($\delta D$) has increased by ~2‰ (~15‰) during February to August (Fig 4.17, $\delta D$ are not shown). In contrast, the $d$ value has decreased by ~1.2‰ at PORT in El Niño period with a decrease of ~2.6‰ during austral summer (December to April), whereas the $d$ values at TPR and GRS are relatively constant with differences of 0.6‰ and -0.6‰, respectively (Fig 4.17).

On the other hand, meteorological data comparison between 2013 and 2015 suggest that temperatures at low (PORT) and midland (TPR) are relatively cooler during El Niño than during ENSO-normal period due to the cooling SSTs in the western Pacific (Fig 4.18). Meanwhile, at high altitude (GRS), temperature is cooler in summer, but warmer in winter during El Niño than during ENSO-normal period. Warmer winter in
Figure 4.17 Comparison of monthly amount-weighted rainfall $\delta^{18}$O and $d$ values between ENSO-normal period (January 2013 - February 2014) and El Niño period (December 2014 - September 2015) at PORT (December to September), TPR (December to September) and GRS (February to August) stations.
Papua highland is possibly due to heat circulation at the tropical troposphere which warms the tropospheric temperatures, usually for up to 6 months, after El Niño begins (Sobel et al., 2002; Sobel, 2014). As expected, precipitation at all elevation and local/regional scales are lower during El Niño than during ENSO-normal period with much drier condition occur during the winter (Fig 4.18). At the same time, the convective activity over Papua are also suppressed during El Niño, as shown by relatively higher OLR values in 2015 than in 2013 (Fig 4.18). This is likely the main cause for the observed enrichment of isotopes ratios in rainfall at all stations (Fig 4.17).

A decrease of $d$ values during El Niño period in lowland station (Fig 4.17), particularly during summer, may be due to cooler SSTs around Papua as the warm pool
shifts to the central Pacific. This leads to a decrease in the kinetic fractionation of evaporation in the ocean and thus, decreases $d$ values of evaporated water vapor and precipitation (Merlivat and Jouzel, 1979).

Although only covering a nearly two-year collection period, comparisons among monthly amount-weighted $\delta^{18}$O, temperature, precipitation and OLR values over the two collection periods are considered to represent the interannual variability over Papua region. Over this period, the correlation between $\delta^{18}$O and local precipitation is negative and insignificant at PORT ($R = -0.14, p = 0.53$), but significant at TPR ($R = -0.56, p = 0.005$) and GRS ($R = -0.53, p = 0.04$). On the other hand, the $\delta^{18}$O are significantly negatively correlated with temperature at PORT ($R = -0.67, p < 0.001$) and TPR ($R = -0.61, p = 0.002$), but insignificant at GRS ($R = -0.33, p = 0.22$). At regional scales, although significant and negative correlation exists between $\delta^{18}$O and precipitation ($R = -0.40, p = 0.05$) but stronger significant positive correlation exists between $\delta^{18}$O and OLR values ($R = 0.74, p < 0.001$) (Fig 4.19). Again, this suggests regional convection, rather than precipitation, plays a dominant role on rainfall $\delta^{18}$O variation at interannual timescales.
Figure 4.19 Time series comparison of monthly amount-weighted $\delta^{18}O$, temperature, precipitation and OLR values over the two collection periods at local and regional scales.
4.4 **Moisture Origins and Transport Pathways**

On seasonal timescales, moisture origins and transport paths have been shown to influence isotope ratios in precipitation by determining the initial stable isotopic ratios of water vapor and additional moisture evaporated from the land surface (land surface recycling) during transport, respectively (Kurita *et al.*, 2009; Breitenbach *et al.*, 2010; Crawford *et al.*, 2013; Suwarman *et al.*, 2013). Initial isotopic compositions ($\delta^{18}O$ and $d$ value) of evaporating water vapor can be affected by the SSTs and surface level relative humidity and wind speed at the site of evaporation (Dansgaard, 1964; Clark and Fritz, 1997; Merlivat and Jouzel, 1979).

For simplicity, it is assumed that air mass back trajectories in Fig 4.10 and Fig 4.11 provide information about the moisture origins and transport paths. From December to March (austral summer), moistures are originating from the seas to the northwest of Papua which are transported to the collection sites by the monsoon westerlies (Fig 4.10a,b,c,l; Fig 4.11a,b,c). During transport pathways, water vapor moved across the Banda Sea, the Arafura Sea and/or the northern Papua tropical rainforest before reaching the collection sites. In contrast, from May to October (austral winter) the southeast trade winds transported moisture from the seas to the southeast of Papua across the southern Papua tropical rainforest to the collection sites (Fig 4.10e-j; Fig 4.11e-h). The moisture sources during April and November (transitional) are combination of both seas to the northwest and southeast of Papua (Fig 4.10d,k; Fig 4.11d).

During austral winter, water vapor originates from the seas to the southeast of Papua that have the long-term mean SSTs of 25 – 29°C, which are in comparison with...
the long-term mean SSTs of 27 – 29°C over the seas to the northwest of Papua where the initial water vapor originates during the summer (Fig 4.20a,b). In addition, mean surface relative humidity over the seas to the southeast Papua during the winter (72 – 82%) is comparable to that over the seas to the northwest Papua during the summer (74 – 84%). Moreover, the mean surface wind speed of the southeast trade winds during winter is about 7 m/s over the seas to the southeast Papua, while during summer the mean surface wind speed over the seas to the northwest Papua decreases from 7 m/s as the northeast trade winds to ~3 – 5 m/s as the monsoon westerlies (Fig 4.20c,d).

The nearly identical meteorological conditions over the moisture oceanic origins during both seasons in Papua region may indicate that the isotopic composition in evaporating water vapor for each season are also nearly identical. A previous study conducted in Australia discusses the moisture oceanic sources in the south of Papua and suggests that there was no significant difference in the rainfall δ¹⁸O when the majority of the moisture originated from oceanic sources (Crawford et al., 2013).

The additional moisture evaporated from land surface recycling during transport from the ocean moisture sources to the precipitation sites can have considerable impact on isotopic composition in precipitation (Salati et al., 1979). The land surface recycling has been shown to yield precipitation with higher d values (e.g. Gat and Matsui, 1991). The mean d values at each station range from 13.43‰ to 17.57‰ during the first collection period and from 13.03‰ to 16.28‰ during the second collection period (Table 4.1), suggesting the major influence of land surface recycling in precipitation during transport to the collection sites. During the summer, additional moisture during transport
likely originates from terrestrial evaporation over the northern tropical rainforest, while during the winter the recycled moisture likely comes from evaporation over the southern tropical rainforest.

Figure 4.20 The long term mean SSTs (°C) from NOAA ERSST (Jan 1979 - Apr 2011) throughout the WPWP for (a) February and (b) August. The climatology of relative humidity (shaded, %) and wind (vector, m/s) at 1,000-mb from NASA MERRA Reanalysis (Jan 1979 - Apr 2011) for (c) February and (d) August
Fig 4.21 depicts the regional MWL and seasonal MWLs for summer and winter during the two collection periods. The slopes and intercepts of MWLs are similar during the ENSO-normal period (Fig 4.21a). The nearly identical intercept \((d)\) values at both seasons suggests that the impact of land surface recycling are likely equal for both seasons and may have caused only a minor seasonal effect on rainfall \(\delta^{18}O\) in the region. While during El Niño, the slope and intercept of summer MWL is lower than the winter's (Fig 4.21b), but is still comparable to the GMWL. This may be caused by lower summer \(d\) values in the lower station due to cooler SSTs at the moisture oceanic sources during El Niño. However, the intercept of 11.71‰ (> 10‰) indicates that land surface recycling still plays an important role on rainfall isotopic composition in Papua during summer 2015. The similar condition also occurs during winter 2015.

In addition, moisture convergence has been shown to affect the isotopic signature of tropical rainfall (Lee et al., 2007; Moore et al., 2014). Moore et al. (2014) suggests that a parameter \(E/P\) (\(E\) is surface evaporation and \(P\) is precipitation), which represents the relative contribution of the converged vapor in precipitation, has a positive correlation with \(\delta D\) of precipitation. However, Fig 4.22 shows that there is no correlation \((R = -0.04, p = 0.90)\) between monthly regional rainfall \(\delta D\) and \(E/P\) derived from MERRA reanalysis in the study area during the first collection period, whereas a positive but insignificant correlation \((R = 0.46, p = 0.18)\) revealed during the second collection period. This result indicates that moisture convergence is not a major control and has insignificant effect on seasonal rainfall \(\delta^{18}O\) in the region.
Figure 4.21 (a) Regional meteoric water lines (RMWLs) during the first collection period. The summer 2013 covers the period of January to March 2013 and December 2013 to February 2014, while the winter 2013 covers the period of May to October 2013. (b) As of (a), but during the second collection period. The summer 2015 covers the period of December 2014 to March 2015, while the winter 2015 covers the period of May to September 2015.
Figure 4.22 Comparison of monthly regional rainfall δD and mean E/P parameter derived from MERRA at 4.5-5.0°S, 136.67°E during the two collection periods

Another factor that may affect the rainfall isotopic composition is a post-condensational effect such as raindrops (secondary) evaporation at the sub-cloud base during rainfall due to surface low humidity, which can decrease d values (Rozanski et al., 1993) and results in ^18O-enriched rainfall. However, this condition is unlikely to occur in humid region such Papua. This is supported by the observed mean relative humidity at all stations across elevation during 2013, which are generally above 80% (Fig 4.23).

Figure 4.23 Mean monthly relative humidity at sample collection stations across elevation during 2013
In summary, moisture origin and land surface recycling along the transport pathways contribute only a minor seasonal effect on rainfall isotopic composition in the region. Other factors such as moisture convergence and raindrop (secondary) evaporation during rainfall also have insignificant influence on rainfall isotopic composition on seasonal timescales.

4.5 The Convection Effect in Association with Temperature at Mean Condensation Level

On seasonal timescales, no or insignificant correlations exist between local/regional precipitation and rainfall $\delta^{18}$O suggesting precipitation amount is not the main controlling factor on seasonal rainfall $\delta^{18}$O. On the other hand, significant positive correlation exists between OLR and rainfall $\delta^{18}$O indicating convective activity plays an important role on seasonal rainfall $\delta^{18}$O. In the Tropics, the lower OLR (< 205 Wm$^{-2}$) values may reflect higher clouds associated with storm system where precipitation forms at lower temperatures. Hence, positive correlations between rainfall $\delta^{18}$O and OLR values can also be interpreted as the temperature (at the mean condensation level) effect.

Instead of rainfall amount, rain condensation mechanisms that correspond to prevailing weather patterns and in-cloud properties when precipitation forms may have greater influence on rainfall isotopic composition. These effects have been suggested by previous studies in Costa Rica (e.g. Rhodes et al., 2006), in Maui, Hawaii (e.g. Scholl et al., 2007), in Puerto Rico (e.g. Scholl et al., 2009), in Ecuador (e.g. Windhorst et al., 2013) and in Sydney, Australia (e.g. Crawford et al., 2013). In general, those studies concluded that the more depleted $^{18}$O precipitation originates in low pressure systems.
associated with tropical storms and large-scale deep convection which are the prevailing weather patterns during the summer (monsoon). In contrast, isotopically enriched precipitation originates in high pressure systems associated with trade wind showers and/or orographic rainfall during the winter (dry) season. The seasonal weather patterns and rainfall δ\textsuperscript{18}O on the southern slope of Papua are consistent with those conditions, suggesting that condensation mechanism is an important factor in controlling rainfall δ\textsuperscript{18}O. By considering the minor and insignificant effect of moisture sources, land surface recycling along the transport paths, moisture convergence and raindrop evaporation on seasonal rainfall δ\textsuperscript{18}O, the relationship between rainfall δ\textsuperscript{18}O and convective activity on seasonal timescales is interpreted as a temperature (at the mean condensation level) effect.

It is difficult to explicitly demonstrate this temperature effect without the echo top data (a measure of the maximum altitude of rainfall within the clouds) from the region (e.g. Scholl et al., 2009). However, the mean echo top altitudes also correspond to the mean condensation levels, which can be estimated by the latent heat (LH) release in the troposphere (e.g. Thompson et al., 2000). The latent heat in the troposphere is released mainly by the condensation process when precipitation forms in clouds. The latent heating datasets have been retrieved by several different algorithms from TRMM satellite measurements (Tao et al., 2006). For simplicity, the monthly LH vertical profile derived from TRMM product 3A12 V7 is used as a proxy of the condensation level in the troposphere. Unfortunately, the LH products are available only over the ocean. Because the collection sites are relatively close to the Arafura Sea to the south, the closest gridded
LH data at 136.25 °E; 5.25 °S is used to represent the collection sites in this study. The vertical profile of monthly LH release from January 2013 to March 2015 is shown in Fig 4.24a. Generally, it exhibits two peaks in the troposphere, at ~0.5 – 2.5 km and at ~3 – 8 km. This is consistent with the double-peak structure of latent heating in the Tropics suggested by previous studies (Zhang et al., 2010; Liu et al., 2015). The shallow mode of the LH peak is associated with shallow convection systems that generate warm rain and low-level heating from shallow cumulus clouds (echo tops are less than 5 km), and partially from cumulus congestus clouds (echo top between 5 and 8 km). The deep mode of the LH peak corresponds to large, deep, organized mesoscale convective systems (MCSs) with echo tops greater than 10 km (Zhang et al., 2010; Liu et al., 2015).

As shown in Fig 4.24a, the deep mode of LH predominates over the shallow mode during the summer (monsoon) season in 2013 (except in March), while the dominant mode was reversed during the winter season (except in July). In September 2013 the troposphere was mainly dominated by the shallow mode of LH. In this study, the maximum altitude of LH release is defined as the highest altitude with heating rate greater than 1.6 °C/day which represents the average maximum latent heating per rainfall rate of 1 cm/day during the onset of the Indian monsoon that generates relatively high precipitation (Magagi and Barros, 2004). The corresponding temperature at the maximum altitude of LH release may indicate the lowest temperature at condensation level at which precipitation forms. An estimate of these corresponding atmospheric temperatures were calculated based on the monthly temperature lapse rates and mean surface temperatures derived from PTFI meteorological data (Fig 4.25).
Figure 4.24 (a) The vertical profile of latent heat release at 136.25°E, 5.25°S from January 2013 to March 2015. (b) The relationship between monthly regional δ^{18}O and the maximum altitude of latent heat release and (c) its corresponding atmospheric temperature. (d) The vertical profile of large-scale cloud fraction over the collection sites at 136.875°E, 4.375°S
The strong relationship between monthly regional $\delta^{18}O$ and the maximum altitude of LH release ($R = -0.66, p = 0.003$; Fig 4.24b) and its corresponding atmospheric temperatures ($R = 0.64, p = 0.004$; Fig 4.24c) from January 2013 to March 2015 compares well with the relationship between monthly regional $\delta^{18}O$ and convective activity (OLR; $R = 0.68, p = 0.002$) over the same period. This result confirms that the effect of convection is associated with the temperature (at the mean condensation level) effect on seasonal rainfall $\delta^{18}O$.

The results are consistent with previous studies in Puerto Rico (Scholl et al., 2009) and in Hawaii (Scholl and Coplen, 2010), which suggests that the seasonal isotopic composition of tropical rainfall is correlated with cloud height and its corresponding (atmospheric) temperature. One common feature that exists among these oceanic islands...
(Puerto Rico, Hawaii and this study area) is the appearance of the Trade Wind Inversion (TWI) layer during the winter season which effectively caps the vertical motion from low elevation and limits the height of cloud development. The prominent shallow mode of LH during austral winter 2013 evinces the TWI layer in the study area (Fig 4.24a). In this season, local convection and the interaction between strong southeast trade winds and diurnal winds result in intense precipitation and mostly nocturnal below the TWI layer. Since the TWI layer limits the development of cloud heights, the precipitation is mainly warm rain which formed from shallow cumulus clouds with a latent heating peak in the 1 – 3 km range (Liu et al., 2015). This is supported by a higher fraction of low and mid-level clouds during June – August 2013 over the collection sites (Fig 4.24d).

Therefore, the δ¹⁸O enrichment of rain during winter 2013 was due to the predominant strong shallow convection system that generated a greater amount of precipitation from relatively lower cloud heights and warmer temperatures. In contrast, during summer 2013, a large-scale deep convection generated precipitation from higher and colder clouds, leading to the depletion of rainfall δ¹⁸O. The most enriched δ¹⁸O was observed in September 2013 which was likely due to the formation of precipitation at lower and warmer clouds with maximum altitude of LH release at ~1.5 km with a corresponding temperature of ~17.5 °C (Fig 4.24b,c), while the low rainfall amount during this month (Fig 4.19) may have been due to the weakest latent heat release in 2013 (Fig 4.24a).

Unlike the lowland stations (PORT, TMK and KK) that receive greater rainfall during the winter than the summer, TPR station shows little seasonality in precipitation
TPR (~1,900 masl) is located just below the TWI layer and is influenced by trade wind orographic showers during the winter. Therefore, the observed $\delta^{18}$O upslope gradient at TPR from May to September 2013 was identical to the $\delta^{18}$O upslope gradients at lowland stations (Fig 4.2a). This may indicate that these stations have a similar gradient of mean condensation level of rainfall during this period as they are located below the TWI layer. On the other hand, the $\delta^{18}$O upslope gradient at the highland GRS station is less steep, suggesting that rainfall was generated from higher mean condensation levels with cooler temperatures. This is likely because GRS is located above the TWI layer.

Scholl et al. (2009) provides information about the mean echo top altitudes and their associated atmospheric temperatures that were categorized by weather patterns (Table 3 in Scholl et al., (2009)). By applying this information, it could be inferred that the prevailing weather patterns during winter at GRS are dominated by troughs, thunderstorms, and easterly waves (tropical easterlies). For instance, showers with thunderstorms were often experienced by the BPCRC team during the ice core drilling operation on the Northwall Firn during June – July 2010. Furthermore, radiosonde records from New Guinea suggest that tropical easterlies are predominant and stronger from June to August at sea level up to at least 6,000 masl (Permana, 2011). These observations support that rainfall at GRS during winter 2013 was likely resulted from higher mean condensation levels with cooler temperatures than other stations below the TWI layer.
During the second collection period (El Niño event), Papua region experiences a high pressure system due to the shift of the warm pool to the Central Pacific. The convective activity is suppressed as the SSTs at surrounding Papua are cooler, as observed in Fig 4.18. This El Niño condition resembles the winter condition but on interannual timescales, where the TWI layer limits the development of cloud heights and precipitation is formed at lower mean condensation level with the corresponding warmer temperature. Fig 4.24 a,b,c indicates that the maximum altitudes of LH release and its corresponding temperatures in 2015 are generally lower and warmer, respectively, than in 2013. For instances, the maximum altitude of LH release (its corresponding temperature) has decreased (increased) from 7.5 km (-11.5°C) to 6 km (-4°C) in January, from 8 km (-14.7°C) to 7 km (-9.5°C) in February.

### 4.6 Comparison with GNIP/IAEA station data

The vertical profile of long-term LH release from January 1998 to December 2013 at 136.25°E, 5.25°S (closest to the collection sites) shows higher altitude of LH release occurring during December to April (summer) and lower altitude of LH during July to October (winter) (Fig 4.26). The peak of shallow LH release on May is possibly due to the chosen grid location that is slightly to the south of the collection sites over the ocean. Instead, the shallow LH peak should occur on June - July at the collection sites when precipitation is maximum.
In order to examine the consistency of the temperature effect as the main control on rainfall $\delta^{18}O$, the long term mean of LH release is derived from the closest grid point to the GNIP/IAEA stations in Jayapura and Madang (see locations in Fig 3.1) and compared to the long term mean of rainfall $\delta^{18}O$ and rainfall amount at both stations (Fig 4.27). Strong negative correlation between rainfall $\delta^{18}O$ and rainfall amount is revealed in Jayapura ($R = -0.66$) and Madang ($R = -0.91$). However, the temperature effect is also apparent, as the more enriched $\delta^{18}O$ months (July to October) are associated with the lower maximum altitudes of LH release in the troposphere with corresponding warmer temperatures, and vice versa.
Figure 4.27 The long-term mean of precipitation $\delta^{18}O$ (connected red triangles) and precipitation amount (connected blue circles) at GNIP/IAEA stations in (a) Jayapura at 140.72°E, 2.53°S during 1961 - 1991 and (b) Madang at 145.80°E, 5.22°S during 1968 - 1982. The long term mean of vertical latent heat release close to (c) Jayapura at 141.25°E, 2.25°S and (d) Madang at 145.75°E, 4.25°S during 1998 - 2013

4.7 Summary

This chapter evaluates the influence of precipitation amount and convective activity on rainfall $\delta^{18}O$ in the southern part of the central mountain ranges of Papua. The results show regional convective activity is a more important factor on controlling rainfall $\delta^{18}O$ variability on daily to interannual timescales than precipitation amount. On intraseasonal timescale, the $\delta^{18}O$ variability at the collection sites resembles the large
scale MJO cycle, with major $\delta^{18}O$ depletion events associated with the active (wet) phases of MJO. The results also demonstrate the nature of the altitude effect.

Air mass back trajectories were simulated to obtain the moisture sources and transport pathways. The findings also suggest that both the temporal and spatial integrative effects of convective activity on rainfall $\delta^{18}O$ are more important than the rainout processes along the transport paths to the collection sites.

Moisture to the collection sites mainly originates from the Pacific Ocean with the significant impact of land surface recycling along the transport paths (as indicated by higher $d$ values) throughout the year as well as during summer and winter in 2013 (ENSO-normal year). However, neither moisture sources nor transport paths appear to have significant seasonal effect on rainfall $\delta^{18}O$ in 2013, as indicated by identical $d$ values between summer and winter seasons in 2013. Moisture convergence and raindrop (secondary) evaporation also has a minor effect on seasonal rainfall $\delta^{18}O$ in 2013.

In conclusion, the convection effect on rainfall $\delta^{18}O$ is likely associated with the temperature (at mean condensation level) effect, which corresponds to the altitude of latent heating release in the troposphere. The more depleted $\delta^{18}O$ values during summer are associated with enhanced deep convection such that rainfall is generated from higher condensation levels (clouds) and cooler temperatures. During winter, local wind circulation is predominant and causes strong shallow convection at lower elevation that generates greater amount of precipitation. However, this convection is limited by the TWI layer, so that precipitation is likely generated from lower condensation level and warmer clouds leading to more enriched rainfall $\delta^{18}O$. 
CHAPTER 5
ICE CORE DATING

The $\delta^{18}$O and $\delta$D profiles of the D1 (32.13 m) and D1B (31.25 m) ice cores show similar variations, attesting to the reproducibility of two records with significant $\delta^{18}$O and $\delta$D variability (5 to 6‰ and 45 to 50‰, respectively) (Fig 5.1; Table 5.1). There are significant depletions of $^{18}$O and D at depths of 26, 17, 12 and 8 meters. Stable isotopic ratios in both cores show gradual but significant enrichment in the top eight meters and smoothing in the top four meters. Both cores are composed entirely of ice with no firm observed. Vertical elongated, or otherwise deformed, bubbles were not observed in either core. The entire high-resolution Core D1 record of stable isotopes, dust and major ions illustrates high aerosol events from 20 to 29 m depth and in the top eight meters (Fig 5.2). Considering that the isotopes are so smoothed in the upper meters and the aerosol concentrations are so low from 8 to 20 m depth and in the lower meters, it seems that there has been substantial post-depositional alteration of these cores. The post-depositional processes that play an important role on low latitude glaciers include evaporation, melting and rain which may cause a re-distribution and/or a washout of isotopic and chemical tracers (Schotterer et al., 2004). However, the stable isotope and major ion variations indicate that perturbations due to melt water percolation did not occur in the ice core, and suggest that the past climate and environmental record remains
preserved in the ice. This is supported by the fact that the effect of melting would remove most of the soluble ions in the ice column through melt water flow, as observed on the Furtwängler Glacier in Kilimanjaro, Africa (Thompson et al., 2011a). Core D1 is the cleanest tropical ice core recovered to date by BPCRC, even compared with the Quelccaya Summit Dome (QSD) core (Thompson et al., 1985) and the Kilimanjaro Northern Ice Field 3 (NIF3) core (Thompson et al., 2002). This is likely because Papua is surrounded by oceans with fewer sources of dust and soluble aerosols (Permana, 2011).

Figure 5.1 The $\delta^{18}$O and $\delta$D profiles of Cores D1 and D1B by depth. Sample resolution of Cores D1 and D1B are ~2.8 cm and ~2 cm, respectively. These profiles demonstrate the reproducibility of the record.
Table 5.1 Descriptive statistics of $\delta^{18}$O, $\delta$D and $d$ values in Cores D1 and D1B

<table>
<thead>
<tr>
<th></th>
<th>Min</th>
<th>Mean</th>
<th>Max</th>
<th>Range</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Core D1</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\delta^{18}$O (‰)</td>
<td>-19.68</td>
<td>-16.07</td>
<td>-13.62</td>
<td>6.06</td>
</tr>
<tr>
<td>$\delta$D (‰)</td>
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<td>-116.36</td>
<td>-96.46</td>
<td>50.19</td>
</tr>
<tr>
<td>$d$ (‰)</td>
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<td>12.21</td>
<td>19.80</td>
<td>11.79</td>
</tr>
<tr>
<td><strong>Core D1B</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\delta^{18}$O (‰)</td>
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<td>-16.41</td>
<td>-14.06</td>
<td>6.26</td>
</tr>
<tr>
<td>$\delta$D (‰)</td>
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<td>-116.87</td>
<td>-96.56</td>
<td>51.21</td>
</tr>
<tr>
<td>$d$ (‰)</td>
<td>10.09</td>
<td>14.41</td>
<td>19.22</td>
<td>9.13</td>
</tr>
</tbody>
</table>

Constructing the timescales of Papua ice cores has been challenging because of the lack of annual resolution of the measured parameters and absence of chronological markers. For instance, little seasonality of precipitation in the highland region suggests no clear distinction between wet and dry seasons and leads to a lack of seasonal variation of the dust preserved in the ice. In addition, little seasonality of temperature is also identified in the highland region with two maximum peaks (April-May and November-December) reflecting the influence of the Sun passing twice over this equatorial region (see in Appendix C for the seasonality at highland GRS, DISP and ALP stations).

Moreover, the expectation that the Papua ice cores might contain organic material that could be used for radiocarbon dating, as the glaciers site is in the middle of tropical rainforest, proved to be unfounded. However, efforts have been made to construct the timescale of the cores, including the use of tritium analysis, $\delta^{18}$O reference matching, and some potential evidence that supports the $\delta^{18}$O reference matching. These procedures are discussed in detail in the following sections.
Figure 5.2 Stable isotopes, insoluble dust and major ion concentrations of Core D1 (1156 samples) by depth. Dust concentrations are plotted with two different scales.
Figure 5.2 Continued from previous page.

Continued.
Figure 5.2 Continued from previous page.
5.1 Tritium Analysis

Tritium is the radioactive form of hydrogen that is produced naturally at low concentrations in Earth's upper atmosphere by cosmic ray bombardment. It has a half-life of $\sim 12.32$ years, and transforms to helium (Lucas and Unterweger, 2000). It is measured in Tritium Units (TU), where 1 TU is defined as the ratio of 1 tritium atom to $10^{18}$ hydrogen atoms. In the mid-1950s and early 1960s, tritium was widely dispersed during above-ground atomic bomb testing. The quantity of tritium in the atmosphere peaked in 1962 – 1963 and has been decreasing thereafter. This 1962/63 peak is useful as a chronological marker in dating ice cores from around the world.

In Papua, the GNIP IAEA/WMO station at Jayapura located about 435 km to the northeast of the ice core drilling sites (Fig 3.1) recorded high tritium concentration in precipitation in 1964 (Fig 5.3). Thus, it is likely that the glaciers near Puncak Jaya preserved this signal within their ice layers.

![Figure 5.3](image-url)

*Figure 5.3* The tritium peak concentration in 1964 ($\sim 33$ TU) in precipitation samples collected by the GNIP IAEA/WMO at Jayapura station ($2.53^\circ$ S; $140.72^\circ$ E; 3 masl; Fig 3.1) which is located northeast of the ice core drilling sites.
Sixty eight samples from Core D1 were analyzed at the University of Bern, Switzerland for $^3$H concentrations to identify this possible bomb horizon. The result shows a peak (~3 TU) recorded at a depth of 23.4 m (Fig 5.4). The differences in concentration between the ice below 26 m and above 22 m suggest that the natural tritium background value has been disturbed by the bomb tests. This peak provides an absolute time marker of 1964 CE (Common Era) at a depth of 23.4 m in Core D1.

Figure 5.4 The absolute time marker (tritium peak, circle with error bars) at 23.4 m in the D1 core represents the 1964 layer associated with thermonuclear bomb testing.

5.2 $\delta^{18}$O Reference Matching

The 13-month running means of monthly isotopic compositions of precipitation from the GNIP stations at Jayapura (1960 - 1991) and at Madang (1968 - 1979) (see location in Fig 3.1) are compared with the NINO3 SST in Fig 5.5. There was a peak enrichment of rainfall $\delta^{18}$O recorded from 1972 - 1973 in Madang and from 1982 – 1984
in Jayapura, which may have been associated with the strong El Niño events in 1972-73 and 1982/83, respectively (Permana, 2011). This is supported by a positive correlation of 0.32 ($p < 0.01$) between the NINO3.4 SST and rainfall $\delta^{18}O$ in Jayapura from 1979 to 1991 (Fig 5.6). This feature has also been observed in rain samples collected during the 2013 ENSO-normal and 2015 El Niño years (Fig 4.17), as discussed in Chapter 4.

Considering this relationship, the NINO3.4 SST was used as a matching time reference to construct the timescale for the Papua ice cores. $\delta^{18}O$ at thirty points in Core D1 were paired with the 13-month running means of NINO3.4 SST by assigning the $^3$H peak at 23.4 meters depth as the 1964 bomb horizon, and assuming the top layer represents the time when the ice cores were collected (May 2010). The reference matching suggests that the $\delta^{18}O$ profile in Core D1 covers ~90 years so that, the bottom of the core is dated at 1920 CE. A similar timescale reconstruction was conducted for Core D1B using twenty six $\delta^{18}O$ matching points, suggesting that the core contains a record of ~77 years and the bottom of the core is dated at 1933 CE (Table 5.2; Fig 5.7). Between the matching points, a time series was developed for each core using linear interpolation. Furthermore, the annual signal was calculated by summing parameter values within the same year for dust and major ions and by averaging for stable isotopic ratios. The calculations were conducted for both annual calendar and thermal year resolution (i.e., thermal year = August of previous year to July of the current year). For instance, thermal year 1998 represents the period from August 1997 to July 1998). Based on this reconstructed timescale, the depth-age relationships for the Papua ice cores are shown in Fig 5.8.
Figure 5.5 Comparisons between the 13–month running means of NINO3.4 SST and rainfall δ¹⁸O at GNIP station in Jayapura and Madang

Figure 5.6 The correlation between the 13–month running means of NINO3.4 SST and rainfall δ¹⁸O at GNIP station in Jayapura
Table 5.2 Reference matching points of $\delta^{18}O$ in Cores D1 and D1B with NINO3.4 SST. The 23.4 meters depth was assigned as the 1964 bomb horizon (bold)

<table>
<thead>
<tr>
<th>Depth (meter)</th>
<th>Year</th>
<th>Depth (meter)</th>
<th>Year</th>
<th>Depth (meter)</th>
<th>Year</th>
</tr>
</thead>
<tbody>
<tr>
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<td>D1B</td>
<td></td>
<td>D1</td>
<td>D1B</td>
<td></td>
</tr>
<tr>
<td>0.00</td>
<td>0.00</td>
<td>2010.4</td>
<td>12.59</td>
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<td>8.37</td>
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</tr>
<tr>
<td>9.80</td>
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<td>12.06</td>
<td>1989</td>
<td>27.90</td>
<td>27.96</td>
<td>1952</td>
</tr>
</tbody>
</table>

Figure 5.7 Thirty points of $\delta^{18}O$ in Core D1 were paired with 13-month running means of NINO3.4 SST by setting the 23.4 meter depth as the 1964 horizon and assuming the top layer represents the time when ice cores were collected (May 2010). The reference matching suggests that the $\delta^{18}O$ profiles in Cores D1 and D1B extend back to 1920 CE and 1933 CE, respectively.
Identification of ENSO Variation from Dust Records

As discussed in Section 2.1.2, Papua interannually experiences drier conditions during El Niño and wetter conditions during La Niña. The concentration of insoluble dust in ice layers would likely increase under drier conditions as more dust is transported by winds from surrounding areas. In addition, higher insoluble dust (microparticle) concentrations in ice cores may also be indicators of higher tropospheric turbidity and wind speeds (Bradley, 1999). The potential sources of insoluble dust particles for these glaciers are likely from crustal inputs around terrestrial New Guinea, forest fires, and dust or ash from volcanic eruptions.
The annual dust concentrations of Cores D1 and D1B were compared with the 13-month running means of NINO3.4 SST, the 2 to 7-yr variance of NINO3.4 SST, the PDO index and the 5-yr running means of wind speed at 500-mb over the New Guinea region (130°E - 150°E; 10°S - 0°S) obtained from NCEP reanalysis datasets (Fig 5.9). It is worth noting that the ice core timescale was constructed based on the matching between δ¹⁸O and NINO3.4 SST, so it is independent of the dust concentration profiles. On interdecadal timescales, Fig 5.9a shows that the ice cores are almost free of dust from 1920 to 1940 show a slight dust increase at the beginning of 1940s, and dust levels remain low until 1950. Since 1950, the dust concentration has been higher and more variable than during the period 1920 to 1940. A possible explanation for this is that the ENSO parameters showed low variance from 1920 to 1960 and high variance from 1960 to 2000 (Torrence and Webster, 1999). During 1920 to 1950 period, ENSO activity was relatively calm with low amplitude (Fig 5.9c), which may have caused the low dust levels in the Papua ice cores. Increased dust in 1950 may have been caused by increased ENSO activity, and coincided with the PDO shift from warm to cool phase (Fig 5.9d). During the cool PDO phase from 1947 to 1976, the WPWP was warmer and more active which led to stronger tropical easterly wind speeds (Fig 5.9e). After 1960, ENSO activity occurred at higher amplitudes which enhanced the troposphere turbidity and wind speed over Papua. This may be marked by the relatively high dust concentration in the ice cores during this time.
Figure 5.9 (a) Annual mean dust concentrations in Cores D1 and D1B. (b) The 13-month running means of NINO3.4 SST. (c) The 2-7-yr wavelet variance time series for NINO3.4 SST after 2-7-yr bandpass filtering (thick black curve with gray shading); the thin curve is 0.5(NINO3.4 SST after 2-7-yr bandpass filtering)². The black (white) shading indicates positive (negative) peaks in the filtered time series, i.e., before squaring, and represents El Niño (La Niña) condition (Torrence and Webster, 1999). (d) The PDO index with red (blue) shading indicates the warm (cool) phases. (e) The 5-yr running means of wind speed at 500-mb over the New Guinea region (130°E - 150°E; 10°S - 0°S) from NCEP reanalysis data.
On interannual timescales, at least 7 out of 9 of El Niño events between 1920 and 2008 (1940-41, 1957-58, 1965-66, 1972-73, 1982-83, 1986-88, 1997-98) were recorded as peaks in dust in the ice core layers (Fig 5.9a,b,c). Only the strong El Niño events in 1925-26 and 1930-31 were not associated with high dust content in the cores, possibly due to the calmer and less variable ENSO activity during that period. The magnitude of dust concentration varies for each El Niño event, as they may also be affected by additional dust from forest fires and volcanic eruptions (Section 5.5). In contrast, strong La Niña events (1942-43, 1954-56, 1970-71, 1973-76, 1988-89, 1998-2000) are marked by relatively low dust content. The relationship between ENSO variation and dust in the Papua ice cores provides evidence to support the reference matching timescale of Core D1 (Section 5.2).

5.4 Identification of Atmospheric Circulation from Major Ion Records

Similar to microparticle profiles, major ion (soluble dust) profiles are also independent of the timescale reconstruction. Major ions in the glaciers are also affected by atmospheric circulation and wind strength. Fig 5.10 shows the comparisons among the annually averaged concentrations of major ion species in Cores D1 and D1B, the PDO index, the average wind speed at 500-mb, and surface pressure over the New Guinea region (130°E - 150°E; 10°S - 0°S) from NCEP reanalysis data. The similar temporal variations among the major ion species suggest that most of the major ions have common sources or similar atmospheric transport pathways.

On interdecadal timescales, the concentrations of major ions are generally correlated with the PDO phases. During a cool phase of the PDO from 1947 to 1976, the
warmer SSTs in the western Pacific Ocean caused lower surface pressures over New Guinea region. Regional pressure gradients enhanced the strength of easterly winds, which is a primary factor in dust transport. In addition, low surface pressures led to ascending vertical motion and increased the tropospheric turbidity. These factors may explain the increase in aerosols deposition on the glaciers during this period. The opposite conditions may have occurred during the warm phases of the PDO from 1925 to 1946 and from 1977 to 1998. During a typical warm phase, the SSTs in the WPWP are cooler, causing slightly higher pressures over Papua which weakens the easterlies and may reduce aerosols transport to the glaciers. This possible connection between the aerosol concentrations in the Papua ice cores and the PDO phases agrees with a previous study of a Tibetan ice core, which demonstrated a significant decrease of aerosol concentrations in the middle to late 1970's, accompanied by regional increases in pressure and decreases in wind strength, all of which coincided with the major shift of the PDO from a cool to a warm phase in 1976 - 1977 (Grigholm et al., 2009).

The stronger easterlies during the cool PDO phase suggests that the potential sources of soluble aerosols are located mainly in the eastern part of the glacier site which consists of the marine source from the Pacific Ocean and the terrestrial sources including the eastern half of New Guinea island and small islands in the west Pacific. Potential sources of soluble calcium (Ca$^{2+}$) and magnesium (Mg$^{2+}$) are sedimentary rocks (e.g. limestone and dolomite) in highlands along the cordillera. Meanwhile, the soluble sodium (Na$^+$) and chloride (Cl$^-$) are likely originated from sea salt aerosols in the Pacific Ocean.
Figure 5.10 Comparisons between the annual concentration of major ion species in Cores D1 and D1B and the PDO index, and the 5-yr running means of wind speed at 500-mb and surface pressure over the New Guinea region (130°E - 150°E; 10°S - 0°S) from NCEP reanalysis data.
5.4.1 Potential Identification of Biomass Burnings

Due to the glacier location in the middle of tropical rainforest, the signature of biogenic emissions are recorded as increased ammonium (NH$_4^+$) and nitrate (NO$_3^-$) ions in ice layers. Both chemical species are generated as part of the biogeochemical cycles. They show high concentrations between the beginnings of the 1950's to the late 1970's, coinciding with a cool phase of the PDO when wind velocity was stronger (Fig 5.10). The ammonium signal is persistent throughout the ice core record suggesting the constant input of biogenic emissions into the glaciers. The potassium (K$^+$) signal resembles the ammonium profile, with high concentrations from 1950 to 1970 and 2000 to 2008, and relatively low concentrations from 1920 to 1949 and 1971 to 1999. In addition, some peaks of ammonium and potassium are also identified at interannual timescales.

Ammonium and potassium are major ions that have been used to identify past forest and land fire activity (Legrand et al., 1992; Echalar et al., 1995; Dibb et al., 1996).

Historical evidence of rainforest fires in PNG have been published in the literature since the 1890s (Johns, 1989; Allen, 1989; Allen et al., 1989). The major forest fire events seem to relate to the recorded drought periods in 1906, 1914/1915, 1941, 1972, 1982/1983 and 1988 (Johns, 1989). Subsequent forest fire events were also reported in 1991 and 1997/1998 (Shearman et al., 2008; Filer et al., 2009). Most fires occurred during El Niño events when this region experiences drought. In the highlands, drought induces cloudless and clear nights which causes frost damage to the crops and leads to increased flammability of the vegetation (Allen and Burke, 1997).
Biomass burnings in the Papua ice cores are identified by peaks in ammonium and potassium ions, which are associated with El Niño-linked drought (Fig 5.10). From the mid-1950s to the mid-1960s, both frost and drought conditions were extensive and severe which lead to food shortages (Allen et al., 1989). From 1972 to 2002, most fires were associated with logging activities and the clearance of land for subsistence agriculture which frequently involved burning that spread into the adjacent forests. At high altitude, montane forest and grasslands in PNG, which are used as wild food sources when crops fail due to drought and highland frost damage, are relatively flammable during the extensive drought period (Shearman et al., 2008). High potassium concentration in the ice cores may have originated from the biomass burning of savannas and grasslands in the highland area. Overall, potential identification of biomass burning from ammonium and potassium records strongly supports the reference matching timescale of D1 core (Section 5.2).

5.4.2 Potential Identification of Volcanic Records

Identification of potential sources of sulfate ($SO_4^{2-}$) ions in the glaciers near Puncak Jaya has been made in order to refine the timescale of Core D1. The natural sources of sulfate in glaciers include terrestrial dust, volcanic activity, sea salt, and marine biogenic emissions (Legrand and Mayewski, 1997). Sulfate levels in the Papua ice cores are noticeable only from 1940 to the beginning of 1970's and after 2000 (Fig 5.10). The sulfate signal is sporadic and lasts only a few years, suggesting that the volcanic input from the vicinity of Papua is likely the potential source.
New Guinea is affected by easterly trade winds in winter and the equatorial/monsoon westerlies in summer. However, the westerly winds that affect Papua are actually the northeast trade winds that veer at the equator. The upper level wind circulation over the New Guinea region is dominated mainly by tropical easterlies (Prentice and Hope, 2007). This regional wind system may explain why the 1963 Mt. Agung volcanic eruption in Bali, Indonesia and the 1991 Mt. Pinatubo eruption in Philippines were not recorded as high peaks of sulfate in the Papua ice cores. These volcanoes are located further to the west and northwest, respectively, from the ice core sites. In reality, the sources of sulfate that are recorded in the Papua ice cores may be volcanoes in mountain ranges to the east of the glaciers.

Figure 5.11 The map of major volcanoes of Indonesia and Papua New Guinea. Volcanoes in black circles are the potential sources of sulfate in the Papua ice cores. A black cross indicates the glaciers near Puncak Jaya (image source: Global Volcanism Program Smithsonian Institution - http://www.volcano.si.edu)
The Global Volcanism Program of the Smithsonian Institution has compiled a record of volcanic eruptions around the world, including the mountain ranges in Papua New Guinea (http://www.volcano.si.edu/). One of the measurements of activity is the Volcanic Explosivity Index (VEI), which provides a quantitative value that encompasses explosive intensity, volume of ash (tephra), and the height of the ejecta into the atmosphere. There are five active volcanic eruptions in PNG from 1920 to 2010: Rabaul, Ulawun, Manam, Langila and Bagana (Fig 5.11). These volcanoes were known active sources of volcanic SO$_2$ from 2005 - 2008 (McCormick et al., 2012).

Figure 5.12 Comparison between sulfate concentrations in Cores D1 and D1B, and the VEI composite of five active volcanoes in Papua New Guinea
Fig 5.12 shows a comparison between sulfate concentrations in Cores D1 and D1B and the VEI composite of five active volcanic mountains in Papua New Guinea. These volcanoes were more active between 1950’s – 1970’s and after 2000, coinciding with the cool phase of the PDO. During these periods, stronger tropical easterlies may have transported a greater amount of volcanic dust and soluble aerosols to the glacier site. These potential sources of volcanic dust may also explain the high concentration of insoluble dust in ice cores during the period from 1950 to the late 1970's (Fig 5.9a). Sulfate peaks in 1969/1970 may be associated with a persistent VEI composite between 1968 and 1972 and a peak in 1970. High concentrations of sulfate after 2000 could be explained by a persistently high VEI composite during that period. However, detailed comparisons by specific eruption remain elusive. In general, the VEI composite time series of five active volcanoes in Papua New Guinea, which may have been the potential sources of sulfate in both cores, supports the reconstructed timescale of the Papua ice cores.
6.1 Comparison of Ice Core δ¹⁸O with Global and Regional Temperatures

Stable isotopic ratios in ice cores, including δ¹⁸O, are commonly used as proxies of temperature. In the Tropics, there is a negative correlation between the rainfall δ¹⁸O and precipitation amount at lower altitude and coastal regions on seasonal timescales (Dansgaard, 1964; Rozanski et al., 1993; Gonfiantini et al., 2001). However, at high mountain regions in the Tropics, air temperature remains the main driver of δ¹⁸O changes on longer timescales (Thompson et al., 2000; 2006). The annual δ¹⁸O in Cores D1 and D1B are compared with the annual mean anomalies of global surface temperature from the NOAA Global Historical Climatology Network (GHCN) and the tropical air temperature (30°S - 30°N) at the surface and at the 550-millibar (mb) levels (elevation of the glacier site) from the NCEP 20th century reanalysis data (Fig 6.1a,b). The D1 δ¹⁸O time series shows an increasing linear trend from 1920 to 2010, with a slope of 0.012‰ per year which mimics the increasing trends of global and tropical temperature anomalies. There is a positive correlation between the annual D1 δ¹⁸O and the annual mean anomalies of global surface temperature ($R = 0.38$, $p < 0.01$), but no correlation after the datasets were detrended ($R = 0.08$, $p = 0.47$). The significant positive correlations reveal between the annual D1 δ¹⁸O and tropical air temperature at the surface
\( R = 0.56, p < 0.01; \ R = 0.41, p < 0.01 \) after detrended) and with temperature at the 550-mb level \( R = 0.53, p < 0.01; \ R = 0.39, p < 0.01 \) after detrended). Over smaller spatial scales, the positive correlations between the annual D1 \( \delta^{18}O \) and the mean temperature anomalies also hold over the region surrounding the glaciers (136 - 138 °E; 4 - 6 °S),

Figure 6.1 (a) The annual \( \delta^{18}O \) of Cores D1 and D1B with the linear trend of D1 core (light line); (b) the annual mean anomalies of global surface temperature (GHCN) and tropical (30°S - 30°N) air temperature at the surface and the 550-mb level (NCEP 20th century reanalysis data); (c) the annual mean anomalies of SST (ERSSTv4) and air temperature at the surface and at the 550-mb level in the Papua glacier region (136 - 138 °E; 4 - 6 °S). Anomalies are based on the climatology period of 1981 to 2010.
both at the surface \((R = 0.46, p< 0.01; R = 0.24, p = 0.02\) after detrended) and the 550-mb level \((R = 0.51, p< 0.01; R = 0.32, p< 0.01\) after detrended). While, the correlation between the annual D1 \(\delta^{18}O\) and SST close to glacier region is significantly positive \((R = 0.26, p = 0.01)\), but is slightly negative when the trend was removed \((R = -0.1, p = 0.52\) after detrended) (Fig 6.1c).

Fig 6.2 illustrates spatial correlations between the annual D1 \(\delta^{18}O\) and extended reconstructed sea surface temperature version 4 (ERSSTv4; Huang et al. (2015); Liu et al. (2015)) data between 70°S and 70°N from 1920 to 2010. The \(\delta^{18}O\) values are positively correlated with equatorial SSTs in the mid to eastern tropical Pacific Basin. However, this result is expected given that the Core D1 was dated by matching the NINO3.4 SSTs to the \(\delta^{18}O\) in the ice core dating process. A similar relationship has also been observed in the \(\delta^{18}O\) record from the Quelccaya Ice Cap, Peru on the eastern side of the Pacific Ocean (Thompson et al., 2013). The similar spatial distribution of the correlations is also shown between the annual D1 \(\delta^{18}O\) and the surface air temperatures over the same period (Fig 6.3). Interestingly, the correlations between the annual D1 \(\delta^{18}O\) and the 550-mb air temperatures (where the glaciers are located) are significantly positive along the tropical bands between 15°S and 15°N during the same time period (Fig 6.4). This could be an indication that the tropical ice core \(\delta^{18}O\) records are good proxies of the tropical upper-level air temperatures. This may also explain the linkage between the \(\delta^{18}O\) records from Papua and the Quelccaya ice core climate records, which are influenced by the tropical Pacific climate variability in the Pacific Basin.
Figure 6.2 The spatial distribution of the correlations between the annual D1δ¹⁸O and the ERSSTv4 anomalies from 1920 to 2010 with the 1981 - 2010 base climatology. The thin black lines enclose areas with \( p < 0.001 \). The glacier site is marked by a blue triangle.

Figure 6.3 As of Fig 6.2, but with the surface (sigma level of 0.995) air temperature anomalies from the NCEP 20th century reanalysis data.
Generally, SSTs over the western Pacific are inversely correlated with the SSTs on the central and eastern Pacific. A comparison of the average SSTs close to Papua glacier region (136 - 138 °E; 4 - 6 °S) and NINO3.4 SST indicates a slightly positive correlation ($R = 0.10, p<0.01; \text{Fig} \ 6.5$) which is possibly due to a trend of the SST warming. However, when the time series were detrended, the correlation becomes significantly negative ($R = -0.26, p<0.01; \text{Fig} \ 6.5$) which is as expected.

In contrast, the detrended upper atmosphere (550-mb level) air temperatures over Papua glacier region are positively correlated with the detrended NINO3.4 SST ($R = 0.35, p<0.01; \text{Fig} \ 6.6$). Maximum correlation, however, is revealed with NINO3.4 SST leading the 550-mb air temperature over Papua glacier region by 6 months ($R = 0.52, p<0.01; \text{Fig} \ 6.6$). This suggests a linkage between the equatorial Pacific SST and the
tropical atmospheric temperatures (Sobel et al., 2002; Chiang and Sobel, 2002) and may explain the impact of ENSO variation on the Papua ice core record in the western Pacific.

Figure 6.5 Comparison of the 13-month running means of average SST anomaly close to glacier region (136 - 138 °E; 4 - 6 °S) and NINO3.4 SST anomaly from 1920 to 2010 with the 1981 - 2010 base climatology

Figure 6.6 Comparison of the 13-month running means of 550-mb air temperature anomaly over glacier region (136 - 138 °E; 4 - 6 °S) and NINO3.4 SST anomaly after the time series were detrended from 1920 to 2010 with the 1981 - 2010 base climatology
6.2 Comparison of Ice Core $\delta^{18}O$ with More Recent Records and Station Data

Over shorter time periods, the annual ice core $\delta^{18}O$ time series are compared from 1979 to 2010 with 550-mb air temperatures, NINO3.4 SSTs, precipitation rate, and convective activity (OLR) at the closest data point to the glaciers near Puncak Jaya, and also with the composite of mean temperatures and precipitation at 11 lowland BMKG stations (Sta No. 2, 3, 5-10, 12, 14 and 15 in Fig 3.1) (Fig 6.7). The correlations among those parameters before and after removing the long-term trends (detrended) are provided in Table 6.1 and Table 6.2, respectively.

Table 6.1 shows that the annual D1 $\delta^{18}O$ has a significant positive correlation with the convective activity data (OLR) from 1979 to 2010, which agrees with the conclusion in Chapter 4 that the convective activity associated with the temperature at mean condensation level is the main control of rainfall isotopic ratios on daily to seasonal timescales. This provides evidence that the OLR may also be the main driver of $\delta^{18}O$ on interannual timescales. During this period, the D1 $\delta^{18}O$ is insignificantly negatively correlated with the precipitation, suggesting the amount effect applies on interannual timescales. On the other hand, the D1 $\delta^{18}O$ is also insignificantly positively correlated with temperatures at the 550-mb level and the surface lowland. This may also suggest the influence of the temperature effect on interannual timescales.
Figure 6.7 (a) The annual δ^{18}O of Core D1 (blue) and D1B (red); (b) the 550-mb air temperatures from MERRA (blue) and NCEP 20th century (red) reanalysis; (c) the composite of mean temperature at 11 lowland BMKG stations; (d) the NINO 3.4 SST; (e) precipitation rates from GPCP (blue) and TRMM (red) data; (f) as in (c) but for precipitation (g) the NOAA OLR data. All data points (except (c) and (e)) are from the closest grid point to the glacier site from January 1979 to June 2010, except TRMM data which extends back only to January 1998.
Table 6.1 The correlation coefficients among parameters that are plotted in Fig 6.5. Darker shadings indicate more significant correlations

<table>
<thead>
<tr>
<th>Correlation</th>
<th>D1 δ¹⁸O</th>
<th>MERRA 550-mb temp</th>
<th>BMKG temp</th>
<th>NINO3.4 SST</th>
<th>GPCP precip</th>
<th>BMKG precip</th>
</tr>
</thead>
<tbody>
<tr>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BMKG temp</td>
<td>R = 0.25</td>
<td>R = 0.56</td>
<td>p = 0.000</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>NINO3.4 SST</td>
<td>R = 0.54</td>
<td>R = 0.31</td>
<td>R = 0.27</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GPCP precip</td>
<td>R = -0.18</td>
<td>R = -0.18</td>
<td>R = -0.01</td>
<td>R = -0.52</td>
<td></td>
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<tr>
<td>BMKG precip</td>
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<td>R = 0.16</td>
<td>R = -0.16</td>
<td>R = -0.60</td>
<td>R = 0.53</td>
<td></td>
</tr>
<tr>
<td>NOAA OLR</td>
<td>R = 0.31</td>
<td>R = 0.28</td>
<td>R = 0.27</td>
<td>R = 0.86</td>
<td>R = -0.52</td>
<td>R = -0.62</td>
</tr>
</tbody>
</table>

Table 6.2 The correlation coefficients among parameters that are plotted in Fig 6.5 after removing the long-term trends. Darker shadings indicate more significant correlations

<table>
<thead>
<tr>
<th>Correlation</th>
<th>D1 δ¹⁸O</th>
<th>MERRA 550-mb temp</th>
<th>BMKG temp</th>
<th>NINO3.4 SST</th>
<th>GPCP precip</th>
<th>BMKG precip</th>
</tr>
</thead>
<tbody>
<tr>
<td>MERRA 550-mb temp</td>
<td>R = 0.16</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BMKG temp</td>
<td>R = 0.20</td>
<td>R = 0.29</td>
<td>p = 0.11</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>NINO3.4 SST</td>
<td>R = 0.55</td>
<td>R = 0.38</td>
<td>R = 0.35</td>
<td>R = 0.35</td>
<td></td>
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</tr>
<tr>
<td>GPCP precip</td>
<td>R = -0.22</td>
<td>R = -0.38</td>
<td>R = -0.19</td>
<td>R = -0.53</td>
<td></td>
<td></td>
</tr>
<tr>
<td>BMKG precip</td>
<td>R = -0.12</td>
<td>R = 0.08</td>
<td>R = -0.34</td>
<td>R = -0.61</td>
<td>R = 0.52</td>
<td></td>
</tr>
<tr>
<td>NOAA OLR</td>
<td>R = 0.31</td>
<td>R = 0.33</td>
<td>R = 0.33</td>
<td>R = 0.86</td>
<td>R = -0.54</td>
<td>R = -0.63</td>
</tr>
</tbody>
</table>
In addition, the linear trends of the annual D1 $\delta^{18}O$, 550-mb air temperatures from MERRA, and the composite of mean temperature from BMKG lowland stations, with slopes of 0.014‰ per year, 0.013 °C per year, and 0.017 °C per year, respectively (Fig 6.7), suggesting atmospheric warming. The NINO3.4 SST trend is relatively neutral. Although the OLR trend is slightly positive over this period (0.021 Wm$^{-2}$ per year), it has small changes compared to its values and is considered as relatively neutral. The precipitation rate has also a positive trend during this period, as shown by data from the GPCP and BMKG lowland stations (0.031 mm/day per year and 0.019 mm/day per year, respectively). This contradicts the amount effect, which would suggest a negative trend as the $\delta^{18}O$ increases. When these parameters were detrended from 1979 to 2010, the correlation between the composite of mean temperature from BMKG lowland stations and the 550-mb temperature became less significant (Table 6.2), while both parameters were more negatively correlated with precipitation and more positively correlated with the OLR data.

Comparison of the annual D1 $\delta^{18}O$ with the climate parameters obtained from PTFI station network would also suggest that the temperature effect dominates over the amount effect. Fig 6.8 illustrates the comparisons from 1997 to 2010 of the annual D1 $\delta^{18}O$ with the composite of annual mean temperatures at highland, midland and lowland stations from the PTFI AWS network. Fig 6.9 depicts similar comparisons, but with the annual average precipitations. The highland composites include stations at elevations of ~4,000 masl and higher (GRS, DISP and ALP; see Fig 3.2; Table 3.2). The midland
values represent stations at elevations of ~2,000 to 3,000 masl (M66, RCMP and M74), while the lowland composites consist of stations near the coast (PORT, M21, and TMK).

The strongest positive relationship is revealed between the annual D1 $\delta^{18}O$ and the mean temperatures at the highland stations ($R = 0.79, p<0.001$; Fig 6.8b), while the correlation of the D1 $\delta^{18}O$ with the mean temperatures at the midland region is insignificantly positive ($R = 0.18, p = 0.53$; Fig 6.8c). In contrast, the correlation of D1 isotopic values with lowland mean temperatures is insignificantly negative ($R = -0.24, p = 0.41$; Fig 6.8d). The lowland mean temperatures resemble the SSTs close to the glaciers region and broadly western Pacific, which generally show negative correlations with the SSTs in the central and eastern Pacific (Fig 6.5). On the other hand, the correlations during the period from 1997 to 2009 between the annual D1 $\delta^{18}O$ and the annual average precipitation at the lowland ($R = -0.20, p = 0.50$) and the midland ($R = -0.23, p = 0.43$) stations are negative but insignificant, while no correlation is shown with the highland stations ($R = -0.04, p = 0.88$) (Fig 6.9).

These results suggest that annual mean temperatures, and not the precipitation amount, in the highlands seem to be the main driver of the annual-scale $\delta^{18}O$ in Papua ice cores. This is supported by the freezing level (0 °C isotherm) altitude (FLA) derived from the closest radiosonde station in Biak (~345 km to the northwest of the glaciers site; see location in Fig 3.1), which indicates an increase (decrease) of the FLA during El Niño (La Niña) events, causing warmer (cooler) conditions at the glacier site (Fig 6.10a). The FLAs are in agreement with the estimated equilibrium line altitude (ELA) which was about 4,950 masl for 1995-1997 and about 4,820 masl for 1997 – 2000 (Prentice and
Glidden, 2010), and was above 4,950 masl for the periods 2000 – 2006 and 2006 – 2011 (Prentice et al., 2012)

Figure 6.8 (a) The annual D1 $\delta^{18}O$; (b) the annual mean temperatures from a composite of highland (~4,000 masl and higher) PTFI stations; (c) as in (b) but from a composite of midland (~2,000 - 3,000 masl) stations; (d) as in (b) but from a composite of lowland stations near the coast
Figure 6.9 (a) The annual D1 $\delta^{18}$O; (b) the annual average precipitation from a composite of highland (~4,000 masl and higher) PTFI stations; (c) as in (b) but from a composite of midland (~2,000 - 3,000 masl) stations; (d) as in (b) but from a composite of lowland stations near the coast.

In addition, the impact of ENSO can clearly be seen in the station records. Precipitation at each altitude level significantly decreased during the 1997-98 El Niño event and increased during the 1998-2000 La Niña event (Fig 6.9). Temperatures in the Papua highlands were warmer (cooler) during the 1997-98 El Niño (1998-2000 La Niña).
according to the NINO3.4 SSTs in the central and eastern Pacific. This may be explained by the effect of ENSO on the tropical troposphere (Sobel et al., 2002; Chiang and Sobel, 2002). During El Niño, warmer SSTs in the eastern Pacific generate high and deep convection and release abundant latent heat in the free atmosphere. The heat circulates the tropical troposphere and warms the tropospheric temperatures, usually for up to 6 months (Sobel et al., 2002; Sobel, 2014), as shown in Fig 6.6. This seems to cause the warming temperatures in the Papua highlands in the west Pacific, as depicted in Fig. 2.4. During La Niña, high and deep convections are strongly enhanced above the New Guinea Island, generating high cloudiness and reducing the highland surface temperatures.

However, temperatures in the Papua midland and lowland are relatively cooler during an El Niño event, as the SSTs in the WPWP are also cooler which suppresses convection and reduces heat transfer into the atmosphere around Papua. This may induce extreme cold weather and frost damage in the midland interior as cloudless skies and drier conditions dominate. The opposite tends to occur during La Niña, when the warmer SSTs in the WPWP enhance deep convection and transfer more heat into the atmosphere above Papua. Surface temperatures along the coastal regions and at midland also become warmer.
Figure 6.10 (a) The monthly freezing level (0 °C isotherm) (blue circles) altitude derived from radiosonde data at Biak station (Fig 3.1). The red line represents the 13-month running means of FLA after interpolation of missing data. The thick black line indicates the drilling site’s altitude at 4,884 masl. (b) As in (a), but for the wind speeds at 500-mb level. (c) As in (b), but for the wind direction.
6.3 Interpretation of the top 4-meters of the Ice Core Records

An interesting feature in both Papua ice core records occurs in the top ~4 meters (dated ~2008 to mid-2010), where a gradual stable isotopic enrichment and smoothing occurs. This interval in both cores also contains high dust and major ion concentrations (Fig 5.1; Fig 5.2). Moreover, the meteoric water lines (MWLs) of core tops show shallower slopes (~6.6) than the GMWL and negative intercept values (Fig 6.11). When the top four meters are excluded, the MWLs of the rest of the cores are closer to the GMWL.

The gradual enrichment of δ\(^{18}\)O in the top ~4 meters can be explained by the warming temperatures that occurred from ~2008 to mid-2010 (Fig 6.8b; Fig 6.10a). However, the smoothness of δ\(^{18}\)O indicates that post-depositional process occurred in this part of the core. Evaporation and sublimation processes in glaciers may cause shallower slopes in the MWL, however these processes are unlikely to occur in a humid region such as Papua with an annual mean relative humidity of ~85% in the highland stations (Permana, 2011). On the other hand, the refreezing process would also cause a shallower slope in the MWL and may result in negative intercept values (Zhou et al., 2008; 2014). Large portions of the surface of the glaciers near Puncak Jaya have likely been influenced by rain (Kincaid, 2007), which will cause surface melting as rainwater transfers heat to the glacier.
Figure 6.11 Meteoric water lines (MWLs) of Cores D1 and D1B for the top ~4 meters (top), the ~4 meters to the bottom cores (middle), and the whole cores (bottom)
Fig 6.10a indicates a gradual increase in the FLA from ~4,900 masl in 2008 to ~5,100 masl in 2010, which may have been associated with a moderate El Niño in 2009-10 and also atmospheric warming. During this period, the FLAs were higher than the altitude of the ice core drilling sites (black thick line in Fig 6.8a) which may cause a greater proportion of precipitation to fall as rain rather than snow. Therefore, a possible explanation for the anomalous stratigraphy in the top four meters may involve the refreezing process caused by mixing rainwater on the glacier surface. The refreezing of the ice will result in relatively more enriched isotopic values compared to deeper parts of the ice column where rainwater has not infiltrated. This is simply because rainwater is enriched in heavy isotopes as it is generated under warmer conditions. This process may also explain the gradual enrichment and smoothing of the isotopic profile in the top four meters of the ice cores.

On the other hand, the high concentrations of aerosol in the top four meters (Fig 5.2) may be associated with the moderate La Niña in 2007-08, followed by the moderate El Niño in 2009-10. During the moderate 2007-08 La Niña, stronger easterly wind speeds might have increased aerosol deposition on the glaciers, but not as much as during the 2009-10 El Niño when the climate was drier with more dust sources exposure (Fig 6.10b,c). In addition, the refreezing of mixing rainwater and surface melt water on the glacier surface may help to concentrate more dust and soluble aerosols in the ice layers. These high concentrations of aerosols may also suggest that the refreezing processes occurred in a relatively short time period so that the water percolation was minimum. It is likely that rainfall and surface melting occur during the day and then refreeze during the
night. This is supported by the diurnal variations of mean temperature and precipitation in the ALP station (4400 masl; ~9 km to the west of the drilling sites) which show a higher precipitation from noon to midnight, and a colder temperature from 1 to 6 AM local time (Fig 6.12). It is worth noting that the temperatures at the ice core drilling site are ~2.2 °C colder than at the ALP station due to higher elevation with the surface lapse rate of 4.6 °C/km for elevation higher than 2,500 masl (Permana, 2011).

Figure 6.12 The diurnal variations of mean temperature and precipitation at the ALP station (4400 masl; ~9 km to the west of the drilling sites). Temperatures at the drilling site are ~2.2 °C colder than at the ALP station.
6.4  Possible Identification of Freezing Level Altitudes from Dust and Chemistry Records

As discussed in Section 6.3, the concentrations of dust and soluble aerosol in the ice cores are not only influenced by an increase of dust sources during a drier condition and a stronger wind circulation, but also by the FLAs that may have reached the drilling site's altitude and play role in post-depositional processes such as mixing rainwater, surface melting, and refreezing which tend to concentrate aerosols on the glacier surface. Therefore, it might be possible to use the dust and chemistry records to partly determine whether or not the FLAs had exceeded the drilling site's altitude in the past.

6.4.1  Reconstruction of Freezing Level Altitudes from Radiosonde Data

Radiosonde data from Biak station show that the mean FLA was about ~4,930 masl in the period of April 1994 to May 2011, which was higher than the drilling site's altitude (Fig 6.10a). The differences between the maximum (in May) and the minimum (in February) values of the 500-mb temperatures and the FLAs in Biak indicates that a 1°C change in the 500-mb temperatures is associated with a ~300 meters change in the FLAs (Fig 6.13a,b). Moreover, the seasonal variation of the 500-mb temperatures in Biak shows a similar pattern with the temperature seasonality at the ALP station near the glaciers site (Fig 6.13b,c). This suggests that the FLA - 500-mb temperature relationship at Biak in this period, which is represented by a transfer function: FLA = 179.4*(500-mb temp) + 5849.6 (Fig 6.14), can be used to calculate the FLAs over the glaciers site. By applying this transfer function to the monthly 500-mb temperatures over the glaciers site
(136° E; 4° S) that were derived from the NCEP 20th century reanalysis in the period of 1920 to 2010, the monthly FLAs over the glaciers site were possible to be reconstructed (Fig 6.15).

From 1920 to 1972, the FLA rose at rate of ~18 m/decade (Fig 6.15). This is consistent with the results from Prentice et al. (2012) which suggest the FLA rate was ~20 m/decade for this period using moraines and current topographic data. After that, the FLA rate was a three-fold increase (~47 m/decade) from 1972 to 2010. This is likely due to the acceleration of atmospheric warming in this period. However, this underestimates the reported FLA rate of ~100 m/decade from 1972 to 2005 by Prentice et al. (2012).

Figure 6.13 (a) the seasonal freezing level altitude at Biak station; (b) the seasonal 500-mb mean temperature at Biak station from radiosonde in the period of April 1994 to May 2011; (c) the seasonal mean temperature at ALP station in the period of January 2000 to April 2009

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Figure 6.14 The relationship between the 500-mb temperatures and the FLAs at Biak station from April 1994 to May 2011. A regression line is in red.

Figure 6.15 The monthly freezing level altitudes (FLAs) over the glaciers site from January 1920 to December 2010 that were reconstructed from the 500-mb temperatures at 136 °E, 4 °S of the NCEP 20th century reanalysis. The thick black line indicates the drilling site’s altitude at 4,884 masl. The dashed lines indicate the slopes of the FLA in particular time periods.
6.4.2 Comparison of Dust and Chemistry Records with Reconstructed Freezing Level Altitudes

Fig 6.16 depicts the comparison between the annual dust, ammonium and $\delta^{18}$O records of Cores D1 and D1B, and the reconstructed FLAs at the glaciers site. The dust concentration firstly peaked between 1939 and 1942, coincided with peaks of ammonium and enrichment of $\delta^{18}$O associated with the 1939 - 42 El Niño events. These peaks can partly be explained by the increased of dust sources (i.e., from the forest fires) due to the drier El Niño condition (Section 5.3 and Section 5.4.1). However, these peaks also coincided with an increase of the maximum FLAs that were ~100 meters higher than the altitude of the drilling sites during this period (Fig 6.16d). This suggests that a portion of precipitation fell as rain rather than snow and led to the refreezing of mixing rainwater during certain months in this period, causing more concentrated aerosols in ice layers. Previously, there were two El Niño events in 1925-26 and 1930-31 which were associated with increases of the FLAs, but there were no dust peaks associated with those events. This could partly be explained by low variance and amplitude of ENSO activity (Fig 5.10c) and also the possibility that the FLA may have not exceeded the drilling site's altitude (Fig 6.16d).

From 1950s to the mid-1970s, the concentrations of aerosols were more variable with a higher background values. Increasing background values of aerosols may associate with the cool phase of PDO and a stronger wind speed which may transport more aerosols from distant sources on the east of the glaciers site (Fig 5.10d,e). While, the interannual variation of aerosols concentration resembles the variation of maximum
Figure 6.16 Comparisons between the annual concentrations of dust and ammonium, $\delta^{18}O$ of D1 and D1B cores, and the reconstruction of maximum, mean and minimum freezing level altitudes over the glaciers site from 1920 to 2010. The thick black dashed line indicates the drilling site's altitude at 4,884 masl.
FLAs which were relatively higher than the drilling site's altitude during this period, supporting the increased concentrations of aerosols in the ice layer.

After the mid-1970s, an increase of aerosols seems to be associated with a rise of annual mean FLAs which were closer to the drilling site's altitude. Since 1987, the annual maximum FLAs have been higher than the drilling site's altitude suggesting that a greater portion of glacier surface being affected by rain which accelerated ice loss in these glaciers. This is in agreement with the result of Kincaid (2007). Since 2001, the annual mean FLAs have been slightly higher than the drilling site's altitude suggesting that the annual temperatures on the glaciers have been greater than 0°C, supporting the melting process and that more precipitation falling as rain than snow. On the same period, the annual minimum FLAs have been lower than the drilling site's altitude suggesting that it is possible that snow accumulation still has occurred.

6.5 Possible Impacts of ENSO and Monsoon System on Papua Ice Core Records

Although the timescales of the D1 and D1B cores were initially constructed by δ¹⁸O reference-matching with the NINO3.4 SST, other proxy evidences have supported the timescale reconstruction. In Section 5.3 and Section 5.4.1, it was shown that the dust and chemistry records support the reconstruction of past ENSO variations in the tropical Pacific. In Section 6.1 and Section 6.2, comparison between ice core δ¹⁸O and meteorological data suggested that the temperature effect is more dominant than the amount effect on the isotopic ratios, with highland temperature plays an important role as the main influence on ice core δ¹⁸O. This section discusses the possible impacts of ENSO
variation and the monsoon system on Papua's climate and the possible implications for the Papua ice core records as given in Table 6.3.

Table 6.3 describes the general climate seasonality in Papua associated with the interannual ENSO variation. Under ENSO-normal conditions, precipitation δ¹⁸O is depleted during the austral summer and enriched during the austral winter, as discussed in Section 4.3.3 and Section 4.6. This seasonal δ¹⁸O variation has been recorded at different elevations in Papua (Section 4.3.3). Descriptions in Table 6.3 were determined based on comparison of rainfall isotopic composition and meteorological data during 2013 - 2015 (see Chapter 4), long-term PTFI meteorological data from different elevations and interpretation of the reconstructed ice core records. During a strong El Niño, the precipitation δ¹⁸O is more enriched due to suppressed convection and lower condensation levels. In the midland (~1,500 - 2,500 masl), extreme cold weather and frost damage have been reported to occur during the winter due to nocturnal radiation cooling (Allen, 1989; Allen et al., 1989, Allen and Burke, 1997). Drought conditions also increase the potential for forest fires and biomass burning which becomes a potential source of dust and soluble aerosols onto the glaciers. During the winter, the highland temperatures are warmer as the latent heat release from the eastern tropical Pacific circulates and warms the tropical troposphere, and more solar radiation is absorbed. The warmer air temperature increases the FLA and the ELA of the glaciers. This condition may cause the ice loss of the glaciers and the deposition of more δ¹⁸O-enriched precipitation as rain rather than snow.
Table 6.3 The possible impacts of ENSO variation and monsoon system on climate and possible implications for glaciers near Puncak Jaya in Papua, Indonesia

<table>
<thead>
<tr>
<th>Normal year</th>
<th>Asian Monsoon System</th>
<th>Australian Monsoon System</th>
</tr>
</thead>
<tbody>
<tr>
<td>- Warm SSTs in the WPWP</td>
<td>- Austral Summer (November to April)</td>
<td>- Austral Winter (May to October)</td>
</tr>
<tr>
<td>- Normal Walker circulation</td>
<td>- Northwest season; Westerly winds</td>
<td>- Southeast season; Easterly winds</td>
</tr>
<tr>
<td>- Low pressure</td>
<td>- Warm (close to the sun)</td>
<td>- Cool (far from the sun)</td>
</tr>
<tr>
<td>- High convection</td>
<td>- Wet season (generally)</td>
<td>- Dry (‘fairly wet’) season (generally)</td>
</tr>
<tr>
<td>- High cloud top and condensation level</td>
<td>- Low pressure; Higher cloud top due to strong convection</td>
<td>- High pressure; Lower cloud top due to less convection and the trade wind inversion layer</td>
</tr>
<tr>
<td>- High cloudiness</td>
<td>- Precipitation $^{18}$O depleted</td>
<td>- Precipitation $^{18}$O enriched</td>
</tr>
<tr>
<td>- High precipitation</td>
<td>- Wet condition</td>
<td></td>
</tr>
<tr>
<td>- Wet condition</td>
<td>- El Niño year</td>
<td></td>
</tr>
<tr>
<td>- Cooler SSTs in the WPWP</td>
<td>- Cooler surface temperatures near the coastal region due to cooler adjacent SSTs</td>
<td>- Cooler conditions near the coastal region due to cooler adjacent SSTs</td>
</tr>
<tr>
<td>- Weaken easterly winds (Walker circulation)</td>
<td>- Cooler condition in midland (~2,000 - 3,000 masl)</td>
<td>- Cooler condition in midland due to nocturnal radiation cooling which causing frost damage</td>
</tr>
<tr>
<td>- Higher pressure</td>
<td>- Cooler condition in highland (&gt; 4,000 masl)</td>
<td>- Warmer condition in highland, possibly due to the circulation of latent heat release in the central Pacific via tropical atmosphere</td>
</tr>
<tr>
<td>- Suppressed convection</td>
<td>- Normal to drier condition at all elevations</td>
<td>- Much drier condition at all elevations</td>
</tr>
<tr>
<td>- Lower cloud top and condensation level</td>
<td>- Increased potential of forest fires and biomass burning</td>
<td>- Increased potential of forest fires and biomass burning</td>
</tr>
<tr>
<td>- Lower cloudiness</td>
<td></td>
<td></td>
</tr>
<tr>
<td>- Lower precipitation</td>
<td></td>
<td></td>
</tr>
<tr>
<td>- Drier condition</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Continued.
Table 6.3 Continued from previous page.

<table>
<thead>
<tr>
<th>La Niña year</th>
<th>Possible implications to the glaciers</th>
<th>Possible implications to the glaciers</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>• Precipitation $^{18}$O enriched</td>
<td>• Precipitation $^{18}$O enriched</td>
</tr>
<tr>
<td></td>
<td>• Potential dust and soluble aerosols deposition as more dust sources exposure</td>
<td>• Potential dust and soluble aerosols deposition as more dust sources exposure</td>
</tr>
<tr>
<td></td>
<td>• Low accumulation</td>
<td>• Increase freezing level altitude</td>
</tr>
<tr>
<td></td>
<td></td>
<td>• Lower accumulation with most precipitation fall as rain rather than snow</td>
</tr>
<tr>
<td></td>
<td>• Warmer temperatures near the coastal region due to warmer adjacent SSTs</td>
<td>• Warmer conditions near the coastal region due to warmer adjacent SSTs</td>
</tr>
<tr>
<td></td>
<td>• Warmer conditions in midland and highland, possibly due to more latent heat released in the atmosphere</td>
<td>• Warmer conditions in midland</td>
</tr>
<tr>
<td></td>
<td>• Mostly much wetter conditions at all elevations</td>
<td>• Cooler condition in highland, possibly due to higher cloudiness which reflect most of solar radiation back to space</td>
</tr>
<tr>
<td></td>
<td>• Strengthen upward vertical motion due to strong convection</td>
<td>• Wetter conditions at all elevations</td>
</tr>
<tr>
<td></td>
<td>Possible implications to the glaciers</td>
<td>• Strengthen upward vertical motion due to strong convection</td>
</tr>
<tr>
<td></td>
<td>• Precipitation $^{18}$O depleted</td>
<td>Possible implications to the glaciers</td>
</tr>
<tr>
<td></td>
<td>• Potential stronger wind speed and upward vertical motions</td>
<td>• Precipitation $^{18}$O depleted</td>
</tr>
<tr>
<td></td>
<td>• High accumulation</td>
<td>• Potential stronger wind speed and upward vertical motions</td>
</tr>
<tr>
<td></td>
<td></td>
<td>• Decrease freezing level altitude</td>
</tr>
<tr>
<td></td>
<td></td>
<td>• Higher accumulation</td>
</tr>
<tr>
<td></td>
<td>• Warmer conditions near the coastal region due to warmer adjacent SSTs</td>
<td>• Warmer conditions near the coastal region due to warmer adjacent SSTs</td>
</tr>
<tr>
<td></td>
<td>• Warmer conditions in midland and highland, possibly due to more latent heat released in the atmosphere</td>
<td>• Warmer conditions in midland</td>
</tr>
<tr>
<td></td>
<td>• Mostly much wetter conditions at all elevations</td>
<td>• Cooler condition in highland, possibly due to higher cloudiness which reflect most of solar radiation back to space</td>
</tr>
<tr>
<td></td>
<td>• Strengthen upward vertical motion due to strong convection</td>
<td>• Wetter conditions at all elevations</td>
</tr>
<tr>
<td></td>
<td>Possible implications to the glaciers</td>
<td>• Strengthen upward vertical motion due to strong convection</td>
</tr>
<tr>
<td></td>
<td>• Precipitation $^{18}$O depleted</td>
<td>Possible implications to the glaciers</td>
</tr>
<tr>
<td></td>
<td>• Potential stronger wind speed and upward vertical motions</td>
<td>• Precipitation $^{18}$O depleted</td>
</tr>
<tr>
<td></td>
<td>• High accumulation</td>
<td>• Potential stronger wind speed and upward vertical motions</td>
</tr>
<tr>
<td></td>
<td></td>
<td>• Decrease freezing level altitude</td>
</tr>
<tr>
<td></td>
<td></td>
<td>• Higher accumulation</td>
</tr>
</tbody>
</table>
During La Niña events, warmer SSTs in the WPWP lead to the strengthening of the easterly winds and upward vertical motion, subsequently enhancing the convection and higher precipitation with more depleted $\delta^{18}$O. The convective activity transfers heat from the surface into the mid-altitudes and highlands through latent heat release during the austral summer. Potentially stronger wind speeds and upward vertical motion may transport dust and soluble aerosols onto the glaciers but not as much as during El Niño when more dust sources are exposed. While during the austral winter, enhanced convection generates high cloudiness and precipitation with higher condensation levels. A more extensive cloud cover reflects most of the solar radiation back to space, which then cool the highland temperatures at first. At the glacier site, the cooler air temperature decreases the FLA and possibly the ELA. Furthermore, higher precipitation and cooler temperatures allow for higher snow accumulation with more depleted $\delta^{18}$O.

These possible impacts of ENSO variation may also apply on decadal and interdecadal timescales. Fig 6.17 shows the comparison between the PDO index and the ice core records, including $\delta^{18}$O, net accumulation, microparticle dust and ammonium concentrations. Net accumulation records were derived from the conversion of annual ice thickness in water equivalent, but not corrected for ice-flow (dynamic thinning) due to the lack of some important parameters such as surface accumulation rate. During the PDO cool phase (La Niña-like) from 1947 to 1976, the ice core records show higher net accumulations and higher concentrations of dust and ammonium. The higher accumulation may be caused by strong convection driven by warmer SSTs in the western Pacific that produce high precipitation and possibly lower the FLA. At the same time, it
also generates stronger easterlies and upward vertical motion, which potentially increase the atmospheric turbidity. Ammonium is usually used as a proxy of biogenic activity from tropical forests. Low net accumulation during La Niña 1998/2000 was possibly due to post-depositional effect by strong wind scouring.

Figure 6.17 Comparisons between the 37-month running means of the PDO index with the 3-yr running means of δ¹⁸O, net accumulations (not corrected for ice-flow dynamic), dust and ammonium ion concentrations in Cores D1 and D1B from 1920 to 2009
6.6 Comparison with other Paleoclimate Records near Papua

As discussed in Section 2.5, most paleoclimate records near Papua were derived from lower elevation sites, such as high-resolution coral records. Most of the coral records from the equatorial western Pacific capture the ENSO variability over the last two centuries (Tudhope et al., 2001; Charles et al., 2003; Asami et al., 2005; Osborne et al., 2014). During El Niño phase, the cool, dry conditions experienced in the equatorial western Pacific result in the more positive coral δ\textsuperscript{18}O, while the warm, wet conditions during the La Niña phase result in the more negative coral δ\textsuperscript{18}O (Tudhope et al., 2001). Coral δ\textsuperscript{18}O is sensitive to SST and seawater δ\textsuperscript{18}O related to SSS variations, while coral Sr/Ca is primarily a recorder of SST (Grottoli and Eakin, 2007). SSTs proxy time series from the equatorial western Pacific have been known to exhibit negative correlations with the NINO 3.4 index, which represents SSTs in the central and eastern Pacific (e.g. Tudhope et al., 2001; Cahyarini et al., 2014).

Figure 6.18 shows the locations of coral records in the vicinity of Papua and the central Pacific from previous studies. The descriptions for each coral site are given in Table 6.4. All of these coral records were from Porites species in order to reduce the possible different interpretation of multi-species coral records. Based on its location, coral records are divided into 5 regions (Indonesia seas, West Pacific, Northwest Pacific, Southwest Pacific and Central Pacific) to assess the possible teleconnection with the ice core records. Time series comparisons of annual Core D1 δ\textsuperscript{18}O with coral records (δ\textsuperscript{18}O and Sr/Ca ratio) used in this study are given in Appendix D and the correlation coefficients for these relationships are summarized in Table 6.5.
As expected, in general, there are positive correlations between the ice core δ\textsuperscript{18}O record and some of coral δ\textsuperscript{18}O records in the northwest, west and southwest Pacific Ocean, and Indonesia seas. These correlations are more significant when the long-term trends were removed (detrended) from time series records, whereas the ice core δ\textsuperscript{18}O has significant negative correlation with coral δ\textsuperscript{18}O records in the central Pacific. These relationships suggest a linkage between the Papua precipitation and the tropical Pacific processes which involve the oceanic and atmospheric teleconnection associated with ENSO variations.
<table>
<thead>
<tr>
<th>Coral Site</th>
<th>Location (Lat; Lon)</th>
<th>Time Period</th>
<th>Reference</th>
<th>Proxy used in this study</th>
<th>Region</th>
</tr>
</thead>
<tbody>
<tr>
<td>A. Bali</td>
<td>8.25°S;115.33°E</td>
<td>1782-1990</td>
<td>Charles et al. (2003)</td>
<td>δ¹⁸O</td>
<td>Indonesia Seas</td>
</tr>
<tr>
<td>B. Timor</td>
<td>10.12°S;123.31°E</td>
<td>1914-2004</td>
<td>Cahyarini et al. (2014)</td>
<td>δ¹⁸O; Sr/Ca</td>
<td>Indonesia Seas</td>
</tr>
<tr>
<td>C. Bunaken</td>
<td>1.01°N;124.82°E</td>
<td>1860-1990</td>
<td>Charles et al. (2003)</td>
<td>δ¹⁸O</td>
<td>Indonesia Seas</td>
</tr>
<tr>
<td>D. Palau</td>
<td>7.28°N; 134.25°E</td>
<td>1793-2008</td>
<td>Osborne et al. (2014)</td>
<td>δ¹⁸O</td>
<td>Northwest Pacific</td>
</tr>
<tr>
<td>E. Guam</td>
<td>13.58°N;144.83°E</td>
<td>1790-2000</td>
<td>Asami et al. (2005)</td>
<td>δ¹⁸O</td>
<td>Northwest Pacific</td>
</tr>
<tr>
<td>F. Madang</td>
<td>5.22°S;145.52°E</td>
<td>1880-1993</td>
<td>Tudhope et al. (2001)</td>
<td>δ¹⁸O</td>
<td>West Pacific</td>
</tr>
<tr>
<td>G. Laing</td>
<td>4.15°S;144.88°E</td>
<td>1884-1993</td>
<td>Tudhope et al. (2001)</td>
<td>δ¹⁸O</td>
<td>West Pacific</td>
</tr>
<tr>
<td>H. Kavieng</td>
<td>2.50°S;150.50°E</td>
<td>1823-1997</td>
<td>Alibert and Kinsley (2008)</td>
<td>Sr/Ca</td>
<td>West Pacific</td>
</tr>
<tr>
<td>I. Rabaul</td>
<td>6.28°S;152.33°E</td>
<td>1867-1997</td>
<td>Quinn et al. (2006)</td>
<td>δ¹⁸O</td>
<td>West Pacific</td>
</tr>
<tr>
<td>J. Vanuatu</td>
<td>15.94°S; 166.04°E</td>
<td>1842-2007</td>
<td>Gorman et al. (2012)</td>
<td>δ¹⁸O</td>
<td>Southwest Pacific</td>
</tr>
<tr>
<td>K. Fiji</td>
<td>16.82°S; 179.23°E</td>
<td>1617-2001</td>
<td>Linsley et al. (2006)</td>
<td>δ¹⁸O</td>
<td>Southwest Pacific</td>
</tr>
<tr>
<td>L. Maiana</td>
<td>1.00°N; 173.00°E</td>
<td>1840-1994</td>
<td>Urban et al. (2000)</td>
<td>δ¹⁸O</td>
<td>Central Pacific</td>
</tr>
<tr>
<td>M. Palmyra</td>
<td>5.88°N; 162.08°W</td>
<td>1886-1998</td>
<td>Nurhati et al. (2011)</td>
<td>δ¹⁸O</td>
<td>Central Pacific</td>
</tr>
</tbody>
</table>
Table 6.5 The correlation coefficients between the annual Core D1 $\delta^{18}$O and the coral records used in this study (see Fig 6.18 for locations, Table 6.4 for descriptions and Appendix D for time series comparisons). Darker blue (red) cell indicates more significant positive (negative) correlation.

<table>
<thead>
<tr>
<th>Coral Proxy vs. Core D1 $\delta^{18}$O</th>
<th>Time Period</th>
<th>Correlation Coefficient</th>
<th>Correlation Coefficient (detrended)</th>
<th>Region</th>
</tr>
</thead>
<tbody>
<tr>
<td>A. Bali ($\delta^{18}$O)</td>
<td>1920-1990</td>
<td>$R = 0.10; p = 0.42$</td>
<td>$R = 0.27; p &lt; 0.05$</td>
<td>Indonesia Seas</td>
</tr>
<tr>
<td>B. Timor ($\delta^{18}$O)</td>
<td>1920-2004</td>
<td>$R = 0.16; p = 0.16$</td>
<td>$R = 0.29; p &lt; 0.05$</td>
<td>Indonesia Seas</td>
</tr>
<tr>
<td>Timor (Sr/Ca)</td>
<td>1920-2004</td>
<td>$R = -0.20; p &lt; 0.10$</td>
<td>$R = 0.04; p = 0.73$</td>
<td>Indonesia Seas</td>
</tr>
<tr>
<td>C. Bunaken ($\delta^{18}$O)</td>
<td>1920-1990</td>
<td>$R = 0.30; p &lt; 0.05$</td>
<td>$R = 0.37; p &lt; 0.01$</td>
<td>Indonesia Seas</td>
</tr>
<tr>
<td>D. Palau ($\delta^{18}$O)</td>
<td>1920-2008</td>
<td>$R = 0.18; p &lt; 0.10$</td>
<td>$R = 0.27; p &lt; 0.05$</td>
<td>Northwest Pacific</td>
</tr>
<tr>
<td>E. Guam ($\delta^{18}$O)</td>
<td>1920-2000</td>
<td>$R = 0.00; p = 0.99$</td>
<td>$R = 0.17; p = 0.13$</td>
<td>Northwest Pacific</td>
</tr>
<tr>
<td>F. Madang ($\delta^{18}$O)</td>
<td>1920-1993</td>
<td>$R = 0.22; p &lt; 0.10$</td>
<td>$R = 0.31; p &lt; 0.05$</td>
<td>West Pacific</td>
</tr>
<tr>
<td>G. Laing ($\delta^{18}$O)</td>
<td>1920-1993</td>
<td>$R = 0.23; p &lt; 0.05$</td>
<td>$R = 0.33; p &lt; 0.01$</td>
<td>West Pacific</td>
</tr>
<tr>
<td>H. Kavieng (Sr/Ca)</td>
<td>1920-1997</td>
<td>$R = -0.07; p = 0.56$</td>
<td>$R = -0.15; p = 0.19$</td>
<td>West Pacific</td>
</tr>
<tr>
<td>I. Rabaul ($\delta^{18}$O)</td>
<td>1920-1997</td>
<td>$R = -0.06; p = 0.61$</td>
<td>$R = -0.09; p = 0.44$</td>
<td>West Pacific</td>
</tr>
<tr>
<td>J. Vanuatu ($\delta^{18}$O)</td>
<td>1920-2007</td>
<td>$R = 0.20; p &lt; 0.10$</td>
<td>$R = 0.35; p &lt; 0.01$</td>
<td>Southwest Pacific</td>
</tr>
<tr>
<td>K. Fiji ($\delta^{18}$O)</td>
<td>1920-2001</td>
<td>$R = 0.26; p &lt; 0.05$</td>
<td>$R = 0.45; p &lt; 0.01$</td>
<td>Southwest Pacific</td>
</tr>
<tr>
<td>L. Maiana ($\delta^{18}$O)</td>
<td>1920-1994</td>
<td>$R = -0.42; p &lt; 0.01$</td>
<td>$R = -0.28; p &lt; 0.01$</td>
<td>Central Pacific</td>
</tr>
<tr>
<td>M. Palmyra ($\delta^{18}$O)</td>
<td>1920-1998</td>
<td>$R = -0.57; p &lt; 0.01$</td>
<td>$R = -0.43; p &lt; 0.01$</td>
<td>Central Pacific</td>
</tr>
</tbody>
</table>

During an El Niño event, the SSTs in the western Pacific are cooler than the average, which lead to suppressed convection and thus drier conditions. Drier conditions increase seawater evaporation and result in the increased of seawater salinity. At the same time, evaporational process removes light water molecules from sea surface into vapor.
phase and leaves the seawater $\delta^{18}O$ enriched, enriching coral $\delta^{18}O$ in the western Pacific. Due to cooler SSTs, temperatures at lowland over New Guinea are relatively cooler than the average. During winter, temperatures at the highland are warmer due to a stable air mass with low moisture content descending over the New Guinea highlands on the high pressure end of the reverse Walker circulation. These conditions are likely to limit the cloud formation at lower elevation. As a result, water vapor from seawater and terrestrial evaporation would likely condense at lower (warmer) mean condensation level (temperature) which lead to the enrichment of precipitation $\delta^{18}O$ and then be recorded in the Papua ice core records. Over the central and eastern Pacific, warmer SSTs enhance deep convection and generate high precipitation which freshen the seawater and cause the seawater $\delta^{18}O$ to decrease over this region, leading to the depletion of coral $\delta^{18}O$ in the central Pacific. Conversely, the opposite tends to occur during La Niña. This explains the positive (negative) correlation between Papua ice core $\delta^{18}O$ and coral $\delta^{18}O$ in the western Pacific (the central Pacific).

Little correlations between Core D1 $\delta^{18}O$ and coral $\delta^{18}O$ records in Bali and Timor (Indonesia seas) may be due to local influences and the impact of Indian Ocean climate associated with the Indian Ocean Dipole (IOD) that affect the SST and SSS over this region, in addition to ENSO impact (Charles et al., 2003; Cahyarini et al., 2014). Ice core D1 $\delta^{18}O$ has no significant correlation with coral $\delta^{18}O$ records in Guam which may be due to its location on the northern edge of the WPWP, while no correlation with coral records in Kavieng and Rabaul are possibly because of the nonlinear response of ENSO to rainfall changes in this region, which tend to be drier during both ENSO extremes (see
Fig 2.5; Smith et al., 2013) and may influence the local SST and SSS in this region. Positive correlations between Core D1 δ\(^{18}\)O and coral δ\(^{18}\)O records in the southwestern Pacific (e.g. Vanuatu and Fiji) are due to the influence of SSS variation on interannual timescales in this region associated with ENSO related-changes in the South Pacific Convergence Zone (SPCZ; see Fig 2.1; Gorman et al., 2012). During La Niña (El Niño), the SPCZ is displaced by about 1 - 3° west (east) of the mean position (Salinger et al., 2014), which gets close to (away from) the region of Vanuatu and Fiji Island causing a high (low) precipitation and a decrease (increase) in the seawater δ\(^{18}\)O, thus a depletion (enrichment) in the coral δ\(^{18}\)O over this region.
CHAPTER 7

PAPUA GLACIERS RETREAT

This chapter discusses the possible factors that control the retreat of glaciers near Puncak Jaya based on the Papua ice core records. The recent glacier retreat has been monitored by annual aerial photography since the 2010 ice core drilling project. Since January 2015, our PTFI colleagues have committed to monitor the glacier changes by taking monthly aerial photographs. This chapter discusses the Papua glacier recession in the larger context of the current state of tropical glacier retreat from the recent published literatures. In addition, the measurement of a 30-meter stake that was placed in one of ice core boreholes in 2010 by the OSU BPCRC team was successfully done in November 2015 provided recent net balance rates of the glaciers over ~5 years which allow assessment and projection of the future for these glaciers.

7.1 Past and Recent Retreat of Papua Glaciers

As discussed in Section 2.2.2, the retreat of glaciers near Puncak Jaya has occurred since ~1850 AD to the present day (Peterson et al., 1973; Allison, 1974; Hope et al., 1976; Allison and Kruss, 1977; Peterson and Peterson, 1994; Van Ufford and Sedgwick, 1998; Klein and Kincaid, 2006; Kincaid, 2007; Prentice and Hope, 2007; Prentice and Glidden, 2010; Prentice et al., 2011; 2012). Total glacier area has decreased
from about 19 km$^2$ in ~1850 (Peterson and Peterson, 1994; Van Ufford and Sedgwick, 1998) to only about 1.8 km$^2$ in 2005 (Kincaid, 2007). A recent study estimates a more extensive glacier area of ~30 km$^2$ in ~1850 based on moraines and current topographic data and only 1 km$^2$ remained in 2011 (Prentice et al., 2012). Much of the history of this retreat should be recorded in the ice core records, as the ice core timescale covers the period from 1920 to 2010.

As shown in Fig 6.1, the annual isotopic values of Core D1 show an increasing linear trend with a slope of 0.012‰ per year, which is associated with the rising air temperature in the Tropics and particularly over the Papua region during the 20th century. It suggests that air temperature is likely the main driver that controls the glacier retreat over this time period. Other contributing factors include increased radiation absorption resulting from reduced albedo, and increased mixing of rainwater on the surface glacier as more precipitation falls as rain rather than snow. However, these processes are in fact basically the consequences of the atmospheric warming that accelerates the glacier retreat.

During the El Niño 1972/73, the glacier retreat was enhanced due to higher solar radiation absorption which led to glacier melting, rather than evaporation or sublimation, due to relatively low regional wind speed and relatively humid air (Allison, 1976). From 1972 to 1987, average monthly atmospheric temperature in the region has increased by 0.24 °C, while from 1988 to 2005 precipitation increased along with the rising of freezing level altitude, which may have led to a greater amount of rain than snow precipitated onto the glacier surface and accelerated glacier retreat (Kincaid, 2007).
Radionsonde data from Biak (the closest location to the glacier site; see Fig 3.1) suggest that the average freezing level (0°C isotherm) altitude was higher than the glacier site's altitude during El Niño events in 1997-98, 2002-03 and 2009-10 (Fig 6.10a). Again, this causes more precipitation to fall as rain than snow (see Section 6.3 and 6.4). Rainwater may induce surface glacier melting and runoff by transferring heat to the underlying ice layer. A greater amount of rain and meltwater could infiltrate into the crevasses, as observed in the Northwall Firn during the 2010 ice core expedition (Fig 2.8). As rain and melt water infiltrate into the crevasses, they would find a way to the glacier bed and act as a lubricant of basal sliding leading to increase ice flow rates (Tedstone et al., 2013; Shannon et al., 2013). The basal sliding very much depends on the temperature of the area, the slope of the glacier, the bed roughness, the amount of meltwater from the glacier, and the glacier's size.

The early aerial photographs of these glaciers, taken in 1936 by J. J. Dozy (see Fig 3a and Fig 5a in Peterson and Peterson (1994)), indicated that the glaciers were still very wide and thick. Meanwhile, the D1 core timescale demonstrates that the 1936 AD layer is at depth of 31.13 m or only 0.97 m from the bedrock. This suggests that most of ice before 1936 were gone by 2010, possibly due to intensive basal melting. This is supported by the measurement of borehole temperatures at the ice core drilling sites which was about ~0 °C (Thompson, L.G., personal communication) allowing the basal melting to occur. Basal melting is enhanced by infiltrated rain and meltwater which are driven by atmospheric warming, more solar radiation absorption and the increase of the freezing level altitude.
Table 7.1 Ice surface areas (km$^2$) for glaciers near Puncak Jaya from ~1850 - 2011 (compilation of previous studies).

<table>
<thead>
<tr>
<th>Year</th>
<th>Total ice area of glaciers near Puncak Jaya (km$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>~1850</td>
<td>19.3$^{(1,3)}$; 30$^{(8)}$</td>
</tr>
<tr>
<td>1936</td>
<td>13$^{(1,4)}$</td>
</tr>
<tr>
<td>1942</td>
<td>9.9$^{(1)}$; 11$^{(7)}$</td>
</tr>
<tr>
<td>1972</td>
<td>6.9$^{(1)}$; 7.3$^{(2)}$</td>
</tr>
<tr>
<td>1974</td>
<td>6.4$^{(2)}$; 6.6$^{(2)}$</td>
</tr>
<tr>
<td>1987</td>
<td>3$^{(3)}$; 5.09$^{(5,6)}$</td>
</tr>
<tr>
<td>1993</td>
<td>~4$^{(3)}$</td>
</tr>
<tr>
<td>2000</td>
<td>2.25$^{(b)}$; 2.33$^{(3)}$; 2.4$^{(1)}$</td>
</tr>
<tr>
<td>2002</td>
<td>2.10$^{(6)}$; 2.15$^{(5)}$</td>
</tr>
<tr>
<td>2003</td>
<td>1.89$^{(6)}$; 2.07$^{(6)}$</td>
</tr>
<tr>
<td>2004</td>
<td>1.91$^{(6)}$</td>
</tr>
<tr>
<td>2005</td>
<td>1.72$^{(6)}$</td>
</tr>
<tr>
<td>2011</td>
<td>1$^{(8)}$</td>
</tr>
</tbody>
</table>


Table 7.1 and Fig 7.1 show compilation of changes in the total ice area for glaciers near Puncak Jaya Papua since ~1850 to 2011 from previous studies. The linear trend of the retreat for these glaciers suggests that the ice area has decreased at ~0.15 km$^2$ per year since ~1850 which implies that these glaciers will likely disappear in ~2017-2018. However, this assessment is based on a considerable uncertainty of the initial ice extent back in ~1850. In ~1850, the total ice area for these glaciers was estimated at ~19 km$^2$ based on the extrapolation of the retreat rates for the period 1936-1974 and the numerical modeling (Allison and Peterson, 1976), but a recent study suggests that it was about ~30 km$^2$ based on moraines and current topographic data (Prentice et al., 2012).
Moreover, the reliability of the ice extent data are considerably lower before 2000 since the estimated glacier areas were calculated based on topographic maps, aerial photographs and limited field survey data. However, since 2000, the total ice areas were measured based on high resolution satellite images (e.g. IKONOS, ASTER and LANDSAT) with less uncertainties and more data reliability (Klein and Kincaid, 2006; Kincaid, 2007).

Figure 7.1 Ice surface area changes for glaciers near Puncak Jaya, Papua, Indonesia from ~1850 to 2011 (compilation of previous studies; see Table 7.1). Red circles are the mean total ice areas and error bars indicate ± 1 standard deviation for associated years. A dashed line represents the linear regression of the glacier retreat.
A direct observation made in June 2010 illustrates the alarming rate of ice retreat. After ~3 weeks camping at Saddle Camp on the Northwall Firn during the ice core drilling campaign, ice around a tent left in place had melted a staggering 30 cm, which suggests that the ice thickness had diminished at a rate of ~10 cm/week (which translates to the ice thinning rate of ~5.2 m/yr (Fig 7.2)). Assuming that this condition is consistent at least since 2010, the glaciers near Puncak Jaya are projected to disappear by ~2017-2018. This is consistent with an assessment of the retreat rate of ice surface area.

![Figure 7.2 A 30 cm thick table of ice around a tent within ~3 weeks stay on the Northwall Firn in June 2010. Photograph was taken by Paolo Gabrielli](image)

Since the ice core drilling program in 2010, efforts in monitoring the ice surface area changes of these glaciers have been undertaken through annual aerial photography.
Since the beginning of 2015, aerial photography has been more frequent through the efforts of our colleagues at PTFI. A series of aerial photographs of the Northwall Firm ice field since 2010 illustrate the retreat of these ice fields in the last five years (Fig 7.3 to Fig 7.10). In 2015, the Northwall Firm was still snow-covered during the austral summer (Fig 7.7), however the snow melted during the austral winter (Fig 7.9). Unfortunately, there is no measurement of the ice area loss for these glaciers from 2010 to 2015 due to inconsistency of angle-capture setting among aerial photographs and no scaling measurement. More extensive work need to be established to perform future mapping for these aerial photographs.

Figure 7.3 An aerial photograph of the Northwall Firn ice field in June 2010. It was taken by Endang Budianto (PTFI Environmental Division)
Figure 7.4 An aerial photograph of the Northwall Firn ice field in 2011. It was taken by Paul Warren (PTFI)

Figure 7.5 An aerial photograph of the Northwall Firn ice field in 2013. It was taken by Yohanes Kaize (PTFI Environmental Division)
Figure 7.6 An aerial photograph of the Northwall Firn ice field in March 2014. It was taken by Yohanes Kaize (PTFI Environmental Division)

Figure 7.7 An aerial photograph of the Northwall Firn ice field in January 2015. It was taken by Yohanes Kaize (PTFI Environmental Division)
Figure 7.8 An aerial photograph of the Northwall Firn ice field in March 2015. It was taken by Yohanes Kaize (PTFI Environmental Division)

Figure 7.9 An aerial photograph of the Northwall Firn ice field in July 2015. It was taken by Yohanes Kaize (PTFI Environmental Division)
One of the intriguing features of these photographs is the separation of the two ice masses that connect the Saddle Camp and the drilling sites in the Northwall Firn ice field (Fig 7.3), which was clearly seen in 2015 (Fig 7.9 and Fig 7.10). The rate of separation likely has been accelerated by the impact of the strong 2015/16 El Niño. In fact, the separation of these two ice masses allows an estimate of the recession of the Northwall Firn south ice front. Fig 7.11 illustrates the changes in the area for the Meren Glacier system which includes the east Northwall Firn and the Meren Glaciers from 1936 to 2006 (Prentice and Glidden, 2010). The separation of the two ice masses containing drilling sites and the Saddle Camp indicates that the south ice front of the East Northwall Firn had receded at least ~510 m from 2006 to 2015 (a red line in Fig 7.11) or a ~33%
decrease in length of the Meren Glacier system fronts since 1936, suggesting a retreat rate of 51 m/year. This rate is about 3.6 times faster than the retreat rate calculated from 1936 to 2006 that receded about ~1 km (~14 m/year). This acceleration rate of retreat is most likely due to the atmospheric warming and its positive feedback.

Figure 7.11 Map of the Meren Glacier System (which includes the East Northwall Firn and the Meren Glaciers) between 1936 and 2006. The inset map shows the 1936 extent on the 2000 orthophotograph. The drilling sites and the saddle camp are marked by a red circle and triangle, respectively. A red line represents the closest distance of ice fronts separating the two sites; modified from Figure 3 of Prentice and Glidden (2010)
7.2 Papua Glacier Recession in Context of the Current State of Tropical Glacier Retreat

In a global context, the tropical glacier recession during the 20th century and into the 21st century has been well documented (e.g. as shown in Fig 7.12 from Thompson et al., (2011)). In the tropical Andes, glaciers have been retreating at an increasing rate since the late 1970's (Rabatel et al., 2013) and have resulted in complex impact to water resources availability and human vulnerability due to the loss of glacier meltwater in Cordillera Blanca (Bury et al., 2011; Baraer et al., 2012), Cordillera Huaytapallana (López-Moreno et al., 2014) in Peru. Mean temperature trend in the tropical Andes has increased at the significant rate of 0.10°C per decade in the last 70 years while precipitation did not show a significant trend since the middle of the 20th century, suggesting atmospheric warming is the main factor causing the current glacier recession with a large proportion of this warming is transmitted to a glacier through a phase change of precipitation from snow to rain (Rabatel et al., 2013). In addition, on interannual and decadal timescales, the variability of tropical Pacific SST associated with the ENSO and the PDO is the main factor controlling the mass balance of glaciers in the tropical Andes (Rabatel et al., 2013; Veettil et al., 2014; 2015; López-Moreno et al., 2014), consistent with the strong positive correlations between ice core δ18O records from Quelccaya Ice Cap, Peru and the NINO4 index (Thompson et al., 2011; 2013).
Figure 7.12 Rate of ice loss percentage per year from four tropical glaciers: (a) Qori Kalis, Peru, (b) Naimona’nyi, western Himalaya, (c) Kilimanjaro, Tanzania, and (d) the glaciers near Puncak Jaya, Papua, Indonesia; Figure 7 from Thompson et al. (2011)

During El Niño events, glaciers experience negative mass balance (e.g. Wagnon et al. (2011) in Bolivia and Francou et al. (2004) in Ecuador), recession of glaciated areas (e.g. Morizawa et al. (2013) in Bolivia) and increase in snowline altitudes (Veettil et al., 2014) due to a warmer air temperature and a drier condition. The warming results in more precipitation to fall as rain, while a drier condition results in reduced cloud cover and more absorption of solar radiation at the ice surface. Both of these conditions induce glacier ablation. The opposite conditions tend to occur during La Niña events during which glaciated areas advance (Morizawa et al., 2013), mass balance increases and
remains positive, and the snowline altitude decreases for a while (Veettil *et al.*, 2014; 2015). Although the glaciated areas are modulated by ENSO events, the magnitude of ENSO impact is observed to be controlled by the phase changes of the PDO (Veettil *et al.*, 2014; 2015). This evidence shows that tropical Andean glaciers respond to changes in both air temperature and atmospheric moisture content related to the ENSO and the PDO on interannual and decadal timescales, respectively. The higher frequency of El Niño events since the late 1970s together with an atmospheric warming may explain much of the recent dramatic shrinkage of glaciers in the tropical Andes (Rabatel *et al.*, 2013).

Due to their locations at the Pacific basin, glaciers near Puncak Jaya, Papua and glaciers in the tropical Andes share a broadly similar response and impact to changes in both air temperature and atmospheric moisture content associated to the tropical Pacific SST variation (Bradley *et al.*, 2003; Diaz *et al.*, 2003) which are closely linked to the ENSO and the PDO phases on interannual and decadal timescales, respectively. This impact is particularly apparent in the high mountain regions where these glaciers are located. This is in contrast with the condition at the sea level where the SST variations in West Pacific (Papua) and in East Pacific (the Andes) are generally inversely correlated. On longer timescales, both of these tropical glaciers recorded an atmospheric warming as indicated by a consistent enrichment of ice core δ¹⁸O over the last century.

In tropical Africa, atmospheric warming has been reported as a primary driver of the recent tropical glacier recession (Thompson *et al.*, 2002; 2009; 2011a). The evidence comes from ice core data from the Northern Ice Field on Mt. Kilimanjaro, Tanzania with a bottom age of ~11.7 kyr and which survived the most severe drought (4,200 yr BP) in
the Holocene (Thompson et al., 2002). This suggests that atmospheric warming, not
drying, is the main controlling factor of the recent ice loss on Mt. Kilimanjaro
(Thompson et al., 2002; 2011a). In addition, a δ\(^{18}\)O record from diatoms in Lake Challa
on the eastern flank of Mt. Kilimanjaro, which recorded the moisture balance
(precipitation/evaporation), was found to be negatively correlated with the δ\(^{18}\)O ice core
record from Mt. Kilimanjaro, suggesting that the moisture balance is not the primary
driver on the long term trend in the ice core δ\(^{18}\)O (Barker et al., 2011). Recent glacial
recession in the Rwenzori Mountains of East Africa is also likely due to a rising air
temperature (Taylor et al., 2006). On the other hand, modeling studies propose that the
atmospheric drying is the primary driver of glacier recession in East Africa during the
20th century (Mölg et al., 2003; 2009; Kaser et al., 2004). Atmospheric drying leads to a
decrease in precipitation, cloudiness and glacier albedo, and consequently increased
absorption of solar radiation, leading to ablation via evaporation and sublimation.
However, the meteoric water lines (MWLs) of all ice cores from Mt. Kilimanjaro in 2000
lies very close to the global meteoric water line (GMWL) which indicates that neither
evaporation nor sublimation play a major role in ice loss today or in the past on
Kilimanjaro (Thompson et al., 2011a).

Recent studies have reported dramatically ice loss of glaciers in tropical Africa
(Prinz et al., 2011; Thompson et al., 2011a; Cullen et al., 2013). Lewis Glacier in Mt.
Kenya had lost of 79% of total ice area and 90% of total ice volume over 76 years from
1934 to 2010 (Prinz et al., 2011). From another study, the recent mapping of glaciers on
Mt. Kilimanjaro suggests that the ice area had retreated from 11.4 km\(^2\) in 1912 to 1.76
km$^2$ in 2011, which represent a total loss of ~85% of ice area over the last 100 years (Cullen et al., 2013). In a broader context, the percentage of the total loss of glacier area in tropical Africa is in agreement with the total loss of ice area in Papua highland which was ~95% over 70-75 years from 1936/1942 to 2011 (Prentice et al., 2012).

Glacier recession has also been reported in the Tibetan Plateau and the Himalaya by previous studies (Yao et al., 2007; 2012; Pan et al., 2012; Pandey and Venkataraman, 2013; Thakuri et al., 2014; Racoviteanu et al., 2014). Most of glaciers in these regions retreated severely because of atmospheric warming with only 10% of glaciers advancing from the 1980s to the 1990s. The retreats were more extensive with only 5% of glaciers advancing in the Third Pole Region (Tibetan Plateau and Himalayas) from the 1990s to 2005 (Yao et al., 2007). The glacier shrinkages generally decrease from the monsoon-dominated regions of the Himalayas to the continental interior of Tibetan Plateau which are dominated by the subtropical westerlies (Yao et al., 2012). In addition to rising temperature, the systematic increase of freezing level altitudes (FLAs) over glaciated areas in the High Asia have been evidenced from radiosonde and reanalysis data since the second half of the 20th century and likely to cause a decrease in glacier mass balances and an increase in the equilibrium line altitudes (ELAs) in these regions (Zhang and Guo, 2011; Wang et al., 2014). This is in agreement with the rising of FLAs and ELAs over Papua glacier region (Prentice et al., 2010; 2012), as shown in Fig 6.15.

In addition, ice core $\delta^{18}$O records from glaciers in the Tibetan Plateau and the Himalaya generally show an enrichment trend since 1900 associated with the climatic warming (e.g. Guliya, Puruogangri and Dasuopu) (Thompson et al., 2006; Yao et al.,
Ice core records from Dasuopu glacier (28°N) which is located in the monsoon-dominated central Himalayas, recorded the south Asian summer monsoon intensity in the past as its net snow accumulations are dominated by summer monsoon precipitation (Thompson et al., 2000a; Davis et al., 2005). Recent study suggests that Dasuopu ice core \( \delta^{18}O \) record is strongly driven by long-term air and sea surface temperature in regions upwind of the ice core site over decadal and multi-decadal timescales, while local/regional precipitation dominates on interannual timescales associated with the ENSO strength variation (Philippoff, 2014). Considering that most of tropical ice core \( \delta^{18}O \) records are modulated by the ENSO variation, they may explain the climate teleconnection through a linkage among tropical precipitations in Papua, the tropical Andes and the Himalayas, and the tropical Pacific processes via precipitation stable isotopes.

7.3 **An Assessment of Papua Glaciers Retreat based on Stake Measurement**

During the ice core drilling project in June 2010, a stake was placed in one of the ice core boreholes on the Northwall Firn. The stake consisted of 15 PVC pipes with length of ~2 meters each and a rope was put through their hole made to connect them while an 8 cm knot was tied at the base of stake to keep the stake in place. Because the pipes extension was not enough to reach the surface, a rope was used to connect the PVC pipe to an iron stake that was hammered in the ice surface to hold the pipe stake with 149 cm of rope length below the 2010 ice surface. The total depth of the stake is 31.49 meters. The detail diagram of the stake is given in Fig 7.13.
Figure 7.13 A sketch of a stake that was placed in a borehole on the Northwall Firn

Since 2010, many attempts have been made to locate and measure the stake. One was conducted by collaboration teams of BPCRC, BMKG and PTFI on August 9 - 19, 2014, but unfortunately, it was unsuccessful due to bad weather conditions and a limited time of expedition. After that, nearly monthly efforts were conducted to locate the stake by PTFI environmental division when the weather allowed, but none of which were successful until November 1 - 7, 2015.
On November 4, 2015, the joint team between BMKG and PTFI found and measured the 2010 stake. They measured a total of 3.77 meters of 2 PVC pipe segments exposed at the ice surface with 2.07 m of the first segment and 1.70 m of the second segment. This indicated along with the 1.49 m of rope that the ice surface has been thinned by ~5.26 meters since 2010, which translated into a rate of ~1.05 m loss of ice thickness per year between 2010 and 2015. This rate of ice loss indicates that ice may disappear within the next ~25 yrs. This assessment was slower than a previous assessment based on the total loss of surface ice area as discussed in Section 7.1. Fig 7.14 depicts the location of the stake on the Northwall Firn ice field in 2010 and 2015, while Fig 7.15 shows the process of the stake measurement during the November 2015 expedition.
Figure 7.14 (Top) the stake location that was placed in the ice core borehole June 2010 (photograph was taken by David Christensen and Greg Chimura). (Bottom) the stake location on November 2015 (photograph was taken by Yohanes Kaize)
Figure 7.15 A stake measurement on November 4, 2015 at 7 AM local time on the Northwall Firn ice field. The measurement indicates that the ice field thinned by ~5.26 meters since 2010. The team members in the field were Yohanes Kaize (PTFI), Rumlus D (PTFI), M. Najib Habibie (BMKG) and Andrew (Helicopter Pilot). Photographs were taken by Yohanes Kaize.
CHAPTER 8
CONCLUSIONS

8.1 Summary

Glaciers near Puncak Jaya in Papua, Indonesia (137°E, 4°S) are situated at 4,884 masl, the highest elevation between the Himalayas and the Andes, in the heart of the west Pacific warm pool which serves as the heat engine for the global climate system. During the Papua ice core drilling expedition from May to June 2010, three ice cores were recovered from the East Northwall Firn. Two ice cores to bedrock were drilled from the west dome of the ice field (Sumantri peak), measuring 32.13 m (D1) and 31.25 m (D1B) in length, and a third core (D2) was drilled from the east dome (Soekarno peak/Ngga Pulu) measuring 26.19 m in length.

In order to support the interpretation of stable isotopes from the Papua ice cores, 1362 rain samples were collected daily between January 2013 and February 2014 (ENSO-normal period) and between December 2014 and September 2015 (El Niño period) at selected stations along the altitudinal transect (9 - 3,945 masl) on the southern slope of the central mountain ranges in Papua. These rainfall isotopes data, along with the instrumental meteorological measurements, contribute to an improved understanding of the factors that control the temporal variability of the isotopic composition of precipitation in this region.

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Rainfall isotope analyses indicate the nature of the altitude effect, with a mean isotopic lapse rate for $\delta^{18}$O ($\delta D$) of $-2.4\%/km$ ($-18.2\%/km$). The results suggest that regional convective activity, rather than precipitation amount and local surface temperature, is a more important factor controlling rainfall $\delta^{18}$O variability on daily to interannual timescales. There is a significant weak positive correlation between rainfall $\delta^{18}$O and OLR values (low OLR is associated with enhanced convective activity and vice versa), but no correlation with precipitation amount is apparent on daily timescales. Air mass back trajectories were simulated to discern the moisture sources and transport pathways. The findings also suggest that the spatio-temporal integrative effects of convective activity on rainfall $\delta^{18}$O are more important than the rainout processes along the transport paths to the collection sites. On an intraseasonal timescale, the $\delta^{18}$O variability in the region resembles the large scale MJO cycle, with major $\delta^{18}$O depletion events associated with the active (wet) phases of the MJO. Furthermore, seasonal rainfall $\delta^{18}$O is generally characterized by summer $\delta^{18}$O depletion and winter $\delta^{18}$O enrichment following the seasonal pattern of regional convective activity, but not precipitation. The regional convective activity is also still the primary driver of monthly rainfall $\delta^{18}$O variation within the ~2-year collection period. In addition, rainfall $\delta^{18}$O ($\delta D$) values at all stations are generally more enriched by 1.6‰ - 2‰ (11‰ - 15‰), and $d$ values are lower during El Niño periods than during ENSO-normal periods.

During the ENSO-normal period in 2013, the moisture sources, the transport pathways, moisture convergence and raindrop (secondary) evaporation appear to have had no significant seasonal effect on rainfall $\delta^{18}$O. This leads to the conclusion that on
seasonal timescales the convection effect is likely associated with the temperature (at mean condensation level) effect. Mean condensation level is represented by the altitude of latent heat release in the troposphere. Therefore, the more depleted rainfall $\delta^{18}$O during the summer is associated with enhanced deep convection such that precipitation is generated at higher mean condensation levels where temperatures are cooler. During the winter, local wind circulation facilitates more shallow convection which is limited by the TWI layer and results in greater precipitation amounts from lower mean condensation levels where temperatures are warmer and this eventually leads to more enriched rainfall $\delta^{18}$O. This observation also applies to rainfall isotope data at the GNIP/IAEA stations in Jayapura and Madang.

The $\delta^{18}$O and $\delta$D profiles of the two longer ice cores (D1 and D1B) show similar variations, confirming the reproducibility of the two records with significant $\delta^{18}$O ($\delta$D) variability of 5 to 6‰ (45 to 50‰). The mean $\delta^{18}$O ($\delta$D) value of both cores is $-16.24%$ ($-116.62%$), which is comparable with mean rainfall $\delta^{18}$O ($\delta$D) of $-16.63%$ (116.07‰) over the ~2-year collection period at the highland GRS site. Significant depletion of $^{18}$O and D occur at depths of 26, 17, 12 and 8 meters in the ice core records. The high-resolution (~2.8 cm sample length) Core D1 record of stable isotopes, dust and major ions illustrates prominent aerosol events from 20 to 29 m depth and in the top eight meters. This upper interval also contains gradual isotopic enrichment, with considerable smoothing in the top four meters, suggesting significant post-depositional alteration. However, the high variability of the stable isotopes and major ions through the rest of the cores suggest that past climate and environmental records are preserved in the ice.
Development of the time series for the ice core records has been very challenging. The lack of organic material in the ice precluded independent radiocarbon dating. However, tritium analysis of Core D1 yielded one absolute time marker which allowed the placement of 1964 CE at a depth of 23.4 m. By using the matching between the long-term time series of rainfall $\delta^{18}O$ from the GNIP/IAEA stations in Papua and NINO3 SST, reference-matching between the latter and Core D1 $\delta^{18}O$ was possible. Assuming that the top of the ice core represents the time that it was drilled (i.e., time is not immediately removed from the surface), this suggests that the Core D1 record covers ~90 years (back to 1920 CE) and Core D1B covers ~77 years (back to 1933 CE). The ice core timescale reconstruction was supported by dating of the dust and major ion profiles (described below), which was done independently from the $\delta^{18}O$ reference-matching.

The ice core dust records identify the ENSO activity in the past on decadal to interdecadal timescales, with high activity often associated with higher dust levels. Increased dust in 1950 coincided with the PDO shift from warm to cool which was associated with stronger easterly winds over Papua. On interannual timescales, El Niño (La Niña) events are characterized by high (low) dust concentrations.

Chemistry records generally correlate with atmospheric circulation on decadal to interdecadal timescales, particularly which associated with the PDO phases. During the cool phase of PDO (La Nina-like) in 1947-1976, the concentrations of major ions were high, possibly due to stronger wind speed and upward vertical motion which increased the tropospheric turbidity. In contrast, during the PDO warm phase, the easterly wind speeds weaken and aerosol transport to the glaciers is reduced. High concentrations of
ammonium and potassium in the ice cores, which are often associated with biogenic emissions, coincide with the PDO cool phases when wind velocities were stronger and supported aerosol transport to the glaciers. However, on interannual timescales, peaks of ammonium and potassium are likely associated with biomass burning such as forest fires that are often associated with El Niño-linked drought. In addition, the sulfate ion record may explain potential volcanic activity in the past, particularly for volcanoes that are located to the east of the glaciers.

The D1 δ¹⁸O time series shows an increasing trend from 1920 to 2010 with a slope of 0.012 ‰ per year which is significantly correlated with the mean tropical air temperature and regional temperature anomalies at the surface and the 550-mb level (where the glaciers are located). Interestingly, the correlations between the annual D1 δ¹⁸O and the 550-mb air temperature anomalies are significantly positive along the tropical bands between 15°S and 15°N during this period, suggesting that the D1 δ¹⁸O record may serve as a good proxy for the tropical upper-level air temperatures. A significantly positive correlation exists between NINO3.4 SSTs and the 550-mb air temperatures over the Papua glacier region (with maximum correlation when NINO3.4 SST leads by 6-months). This suggests a linkage between the equatorial Pacific SSTs and the tropical atmospheric temperatures and potentially explains the impact of ENSO variation on the Papua glaciers.

Over a shorter time period (1997 - 2010) on a regional scale, the annual D1 δ¹⁸O has a significant positive correlation with convective activity (represented by OLR data). In addition, there is a positive correlation between the annual D1 δ¹⁸O and annual mean
temperature both at the 550-mb level and at lowland stations, with increasing trends of 0.014‰ per year, 0.013 °C per year, and 0.017 °C per year, respectively, suggesting atmospheric warming over this period. On the other hand, an insignificant negative correlation exists between the annual D1 δ¹⁸O and the precipitation rate; however, a positive trend in the precipitation rate during this period (0.019 - 0.031 mm/day per year) refutes the argument for an amount effect on longer timescales.

Over the same period but on smaller time intervals, the annual D1 δ¹⁸O has a significant positive correlation with the annual mean temperatures in highland stations (higher than 4,000 masl), but has insignificant positive to negative correlations with annual mean temperatures in the lowland (near coastal region) and midland (2,000 - 3,000 masl) stations. Moreover, an insignificant negative correlation exists with lowland and midland stations, and no correlation exists in the highlands between the annual D1 δ¹⁸O and annual precipitation. Positive correlation between the annual D1 δ¹⁸O and the highland temperatures is supported by the Freezing Level Altitude (FLA; 0°C isotherm) data derived from the closest radiosonde station in Biak (~345 km to the northwest of the glaciers site), which indicate that the increase (decrease) of the FLA during El Niño (La Niña) events are associated with warmer (cooler) conditions at the glacier site.

The top ~4 meters of the Papua ice cores (dated ~2008 to mid-2010) are characterized by tremendously high dust and soluble aerosol concentrations and a smooth gradual stable isotopic enrichment showing a MWL with a slope of ~6.6 and a negative intercept value, indicating that strong post-depositional processes affected this part of the core. As inferred by radiosonde data, during period 2008 - 2010 the FLA was higher than
the glacier site altitude, which may have caused more precipitation to fall as rain rather than snow on the ice surface. Since evaporation and sublimation are unlikely to occur in this humid region, a possible explanation for the anomalous stratigraphy in the top ~4 meters may involve the refreezing process of mixed rainwater and meltwater on the surface. In addition, the refreezing process may help to concentrate more dust and soluble aerosols in the ice layers.

Reconstruction of the FLA over the glacier site from 1920 to 2010, based on radiosonde data from Biak and NCEP 20th century reanalysis data, suggests that the FLA rose at rate of ~18 m/decade from 1920 to 1972 and ~47 m/decade from 1972 to 2010. The results also indicate that an increase of 1°C in the 500-mb air temperature is associated with an FLA rise of ~300 meters.

Comparisons of dust and chemistry records of Papua ice cores and the FLA reconstruction suggest that peaks of dust and ammonium often are associated with El Niño events associated with the increased of maximum FLAs that have exceeded the altitude of the ice core site. This suggests that the high concentrations of aerosols in the ice cores are not only influenced by increased dust sources during drier conditions and a stronger wind circulation, but also partly by post-depositional processes such as refreezing of mixed rainwater and meltwater associated with high FLAs.

Using a combination of rainfall isotopic variations, instrumental data during 2013 - 2015, long term meteorological data from PTFI, and reconstructed ice core records, a description of the possible impacts of ENSO and the monsoon system on the Papua glaciers can be described. During a strong El Niño event, cooler summer SSTs in the
west Pacific influence lower temperatures at all elevations, which result in weakened easterly winds, suppressed convection, lower mean condensation levels and drier conditions in Papua. This increases the potential of forest fires and biomass burnings. In the midlands, nocturnal radiation cooling often causes frost damage. During winter, temperatures in the highlands are warmer, possibly due to the circulation of latent heat release from the central Pacific via the tropical atmosphere which increases the FLA in the region. Implications for the glaciers include precipitation that is enriched in $^{18}\text{O}$, increases in aerosol deposition on the ice surface from more exposed sources, and low accumulation with most precipitation falling as rain rather than snow.

During a strong La Niña event, warmer SSTs in the west Pacific lead to the strengthening of easterly winds and upward vertical motion, subsequently enhanced convection and high precipitation with higher mean condensation levels. Temperatures are warmer at all elevations as the convection transfers energy from the surface to the highland through latent heat release during the summer. During winter, enhanced convection generates more extensive cloud cover which reflects most solar radiation back to space and subsequently lowers the highland temperatures and decreases the FLA in the region. Implications for the glaciers include higher precipitation and cooler temperatures in the highlands allow for potentially higher snow accumulation with more depleted $\delta^{18}\text{O}$.

Positive correlations between Papua ice core $\delta^{18}\text{O}$ records and other paleoclimate records, particularly nearby coral $\delta^{18}\text{O}$ records from the northwest, west and southwest Pacific Ocean and Indonesia seas, suggest linkages between the Papua precipitation and the tropical Pacific processes which involve the oceanic and atmospheric teleconnection
associated with ENSO variations. In contrast, the ice core $\delta^{18}O$ has significant negative correlation with coral $\delta^{18}O$ records in the central Pacific. The positive correlations are more significant when the long-term trends were removed from the time series.

Past and recent retreat of glaciers in the Papua highlands have been studied by measuring the total surface ice area using topographic maps, aerial photographs, field survey data and high resolution satellite images. Compilation of previous studies indicates that the total ice area near Puncak Jaya has decreased at a rate of ~0.15 km$^2$ per year since ~1850, which implies that these glaciers will likely disappear by ~2017-18. Personal observations made during the 2010 ice core drilling campaign emphasize the fate of these glaciers. After ~3 weeks of living and working at Saddle Camp on the Northwall Firn, ice around the shelter had melted a staggering 30 cm, suggesting that the ice thickness was diminishing at a rate of ~10 cm/week or ~5.2 m/yr. If this condition has persisted since 2010, this supports the ground-based and satellite projections that the glaciers near Puncak Jaya will disappear by ~2017-18. A recent measurement of the stake that was placed in the ice core borehole in 2010 suggests that the ice thickness has reduced by ~5.26 meters between 2010 and 2015 with a thinning rate of ~1.05 m/year. This rate suggests that glaciers in Papua will completely disappear by ~2040 which provides the upper bound of the date for the glaciers disappearance.

Observations made in 2015 show that the two ice masses containing the drilling sites and the Saddle Camp had separated, indicating that the East Northwall Firn south ice front had receded at least ~510 m between 2006 to 2015, suggesting a retreat rate of 51 m/year. This rate is about 3.6 times greater than the retreat rate of the same ice front
that receded about ~1 km between 1936 to 2006 (~14 m/year). The accelerating rate of retreat is most likely due to atmospheric warming, resulting in more snow falling as rain across all elevations on these glaciers.

### 8.2 Suggestions for Future Work

Future research is warranted in order to generate more comprehensive data sets and to provide a better understanding of the climate system in this region, particularly the controlling factor of precipitation stable isotopic variation in the Tropics. To accomplish this it is highly recommended that the rain sample collection in the Papua region be continued for a longer time period, particularly through a strong La Niña event, as it would improve the interpretation of rainfall stable isotopic variability in this tropical region. This would allow a comparison of rainfall isotopic ratios between ENSO-normal, El Niño and La Niña events.

Continued and frequent measurements of the summit accumulation stake will allow determination of the seasons of surface thinning and how it varies from year to year. These data, along with aerial photographs of the glaciers, would allow the documentation of the accelerating retreat of these glaciers.

In a global context, a comparison among tropical ice core records, including those from Papua, South America, the Tibetan Plateau and tropical Africa, needs to be conducted to obtain a broader view, as well as a better understanding, of the causes and mechanisms of tropical glacier retreat and global impacts of phenomena such as ENSO and monsoon variability. In addition, such a compilation of tropical ice core records will
aid in the reconstruction of past and current tropical climate change, and eventually aid
the assessment of future climate variability based on the understanding of how climate
changed in the past through the use of global and regional climate models.
REFERENCES


Reeves, J. M., Barrows, T. T., Cohen, T. J., Kiem, A. S., Bostock, H. C., Fitzsimmons, K. E., Jansen, J. D., Kemp, J., Krause, C., Petherick, L., and Phipps, S. J. (2013). Climate variability over the last 35,000 years recorded in marine and terrestrial archives in the Australian region: An OZ-INTIMATE compilation. Quaternary Science Reviews, 74(0), 21-34.


Torrence, C., and Webster, P. J. (1999). Interdecadal changes in the ENSO - monsoon system. *Journal of Climate, 12*(8), 2679-2690.


Appendix A:

Possible climatic impact of ENSO extremes on the Asia-Pacific Region
POSSIBLE CLIMATIC IMPACT OF AN LA-NIÑA EVENT ON THE ASIA-PACIFIC REGION

Legend
Change in Precipitation
- Less Precipitation
- More Precipitation
Change in Temperature
- Cooler Temperatures
- Warmer Temperatures

Countries covered by the Regional Office for the Asia-Pacific

Map Data Source: UNOCHA
Map Information: 19 September 2010
Web Resources: http://data.unocha.org/la-nina

Notes:
- During the Boreal winter months, La-Niña conditions are most likely to develop in the South Pacific and North America.
- During the Boreal summer months, La-Niña conditions are most likely to develop in the tropical Pacific and the northeastern United States.
Appendix B:

The seasonal variation and monthly time series of temperature and precipitation of BMKG stations used in this study

(See Fig 3.1 for locations and Table 3.1 for descriptions)
Seasonal Variation

Sta No 2. FAK-FAK (130 masl)

Sta No 3. MANOKWARI (3 masl)

Sta No 6. BIAK (11 masl)

Sta No 7. SERUI (3 masl)

Sta No 8. NABIRE (3 masl)

Sta No 9. ENAROTALI (1770 masl)
Sta No 10. TIMIKA (56 masl)

Sta No 11. SARMI (3 masl)

Sta No 13. SENTANI (99 masl)

Sta No 14. JAYAPURA (3 masl)

Sta No 15. WAMENA (1660 masl)

Sta No 16. TANAH MERAH (16 masl)

Sta No 17. MERAUKE (3 masl)
Monthly Time Series
Sta No 2. FAK-FAK (130 masl)
Sta No 3. MANOKWARI (3 masl)

**Temperature Graphs:**
- Tmin: \( y = -0.033x + 97.33 \)
- Tmax: \( y = 0.011x + 4.46 \)
- Tmean: \( y = 0.050x - 76.51 \)

**Precipitation Graph:**
- Monthly precipitation

From 2002 to 2020.
Sta No 6. BIAK (11 masl)
Sta No 7. SERUI (3 masl)
Sta No 8. NABIRE (3 masl)

monthly - NABIRE

Tmin
Tmax
Tmean

\[ y = 0.025x - 18.96 \]

\[ y = 0.016x - 5.41 \]

\[ y = 0.021x - 19.36 \]

monthly - NABIRE

precipitation

mm

0 100 200 300 400 500 600 700 800 900 1000

Sta No 9. ENAROTALI (1770 masl)
Sta No 10. TIMIKA (56 masl)
Sta No 11. SARM (3 masl)
Sta No 13. SENTANI (99 masl)
Sta No 14. JAYAPURA (3 masl)
Sta No 16. TANAH MERAH (16 masl)
Appendix C:

The seasonal variation and monthly time series of temperature and precipitation of PTFI AWS network

(See Fig 3.2 for locations and Table 3.2 for descriptions)
Seasonal Variation

PORT (9 masl)

M21 (27 masl)

TMK (37 masl)

KK (67 masl)

M50 (617 masl)

TPR (1900 masl)
M66 (2350 masl)

RCMP (2410 masl)

M74 (2750 masl)

GRS (3945 masl)

DISP (4109 masl)

ALP (4400 masl)
Monthly Time Series
PORT (9 masl)
M21 (27 masl)
TMK (37 masl)

- Tmin
- Tmax
- Tmean

$y = 0.007x + 18.54$

$y = -0.000x + 26.83$

$y = -0.013x + 48.91$

precipitation

mm

0 100 200 300 400 500 600 700 800 900

KK (67 masl)

![Graphs showing temperature and precipitation trends over time.](image)

- Tmin: $y = -0.041x + 114.07$
- Tmax: $y = -0.085x + 154.88$
- Tmean: $y = -0.062x + 145.43$

- Precipitation: Over the years, the precipitation levels vary significantly, with peaks in 2004, 2008, 2011, and 2014.
M50 (617 masl)

![Graph showing temperature and precipitation trends](image)

- Tmin: $y = -0.055x + 138.48$
- Tmax: $y = -0.112x + 246.82$
- Tmean: $y = -0.109x + 236.85$

- Precipitation: Variations over time from 2000 to 2025
M66 (2350 masl)
RCMP (2410 masl)
M74 (2750 masl)
GRS (3945 masl)

\[ y = -0.088x + 186.53 \]
\[ y = -0.050x + 105.66 \]
\[ y = -0.016x + 33.89 \]
ALP (4400 masl)

\[ y = 0.508x + 1207.96 \]

\[ y = 0.220x + 436.52 \]

\[ y = 0.152x + 303.70 \]
Appendix D:

Time series comparisons of annual Core D1 δ¹⁸O and coral records in the Papua vicinity and the central Pacific

(See Fig 6.18 for locations and Table 6.4 for descriptions)
Region: Indonesia Seas

Bunaken

R = 0.30; p = 0.01
R = 0.37; p = 0.00 (detrended)

Bali

R = 0.10; p = 0.42
R = 0.27; p = 0.02 (detrended)

Timor

R = 0.16; p = 0.16
R = 0.29; p = 0.01 (detrended)

Timor

R = -0.20; p = 0.06
R = 0.04; p = 0.73 (detrended)
Region: West Pacific

**Madang**

$R = 0.22; \ p = 0.06$

$R = 0.31; \ p = 0.01$ (detrended)

**Laing**

$R = 0.23; \ p = 0.04$

$R = 0.33; \ p = 0.00$ (detrended)

**Rabaul**

$R = -0.06; \ p = 0.61$

$R = -0.09; \ p = 0.44$ (detrended)

**Kavieng**

$R = -0.07; \ p = 0.56$

$R = -0.15; \ p = 0.19$ (detrended)
Region: Central Pacific

Maiana

R = -0.42; p = 0.00
R = -0.28; p = 0.01 (detrended)

Palmyra

R = -0.57; p = 0.00
R = -0.43; p = 0.00 (detrended)