Last Deglaciation Rainfall Changes

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Three drier periods (lower rainfall) (i.e., before ~17, ~1513.5, and 7–3 ka BP) and three wetter periods (higher rainfall) (i.e., ~17–15, ~13.5–7, and after ~3 ka BP) were detected on Southern Indonesia (off southwest Sumba) based on geochemical element (terrigenous input) proxies (ln Ti/Ca and K/Ca). During the Last Deglaciation, AISM rainfall responded to high latitude climatic events related to the latitudinal shifts of the austral summer ITCZ. Sea level rise, solar activity, and orbitally-induced insolation were most likely the primary driver of AISM rainfall changes during the Holocene, but the driving mechanisms behind the latitudinal shifts of the austral summer ITCZ during this period are not yet understood.

Keywords: paleoclimate ; Australian-Indonesian monsoon ; elemental ratio ; Southern Indonesia

1. Introduction

Despite its importance to the livelihood of people in the densely populated Southern Indonesia region, Australian-Indonesian monsoon (AIM) rainfall is still poorly understood [1][2]. A study on AIM past changes is significant to generate robust analogs as the basis to predict and model AIM rainfall future changes [3][4]. Previous studies from both marine and non-marine proxies in the AIM region (Southern Indonesia and Northern Australia) suggested millennial – multi-millennial scale changes of AIM during the Last Glacial Maximum (LGM)—Holocene [1][5][6][7][8][9][10]. Based on modern conditions as analog, drier (wetter), or lower (higher) rainfall conditions characterizes the Australian-Indonesian winter (summer) monsoon intensification [2][11]. In general, drier condition was inferred at Last Glacial while wetter condition characterized the Holocene [1][7][8][11][12][13]. Throughout the Last Deglaciation, wetter and drier periods on a millennial-scale which corresponded to the high latitude climatic events (i.e., Heinrich Stadial 1, Antarctic Cold Reversal, Bølling-Allerød Interstadial, and Younger Dryas) were inferred [1][8][9][10][14]. This indicates the importance of atmospheric-oceanic interhemispheric teleconnection on past AIM changes [1][4][15]. Wetter and drier periods on millennial scales were also inferred throughout the Holocene [1][2][8][10]. The Early Holocene (~11–7 ka BP) was characterized by wetter conditions [6][7][16] while Mid and Late Holocene were marked by the contrast condition between Southern Indonesia (drier Mid Holocene and wetter Late Holocene) [1][5][7][16] and Northern Australia (wetter Mid Holocene and drier Late Holocene) [13][17][18][19].

Orbital-induced insolation, solar forcing, glacial-interglacial changes in climatic and oceanographic conditions (i.e., surface temperature and eustatic sea level), and the high latitude climatic events (caused by the gradual melting of northern high latitude ice sheets) were most likely the driving mechanisms of AIM variability during the Last Glacial—Holocene which resulted in the changes of AIM rainfall [1][2][7][8][11]. The orbital forcing variability affects the insolation changes on a multi-millennial scale [15][20][21][22]. In the case of AIM, the rising (decreasing) in southern hemisphere (SH) insolation results in a stronger (weaker) Australian-Indonesian summer monsoon (AISM) [4]. Variability of solar forcing induces the earth's surface temperature, which affects the quantity of atmospheric water vapor from seawater evaporation [23]. The rise and fall of eustatic sea levels are related to the changes in polar ice volume, which are induced by the changes in surface temperature [4][24]. Gradual melting of the northern high latitude ice sheets was responsible for the changes of Atlantic Meridional Ocean Circulation (AMOC), which resulted in the high latitude climatic events throughout the Last Deglaciation [15]. Although the past AMOC changes were induced by events in the North Atlantic, their effects could have extended to the SH through the "bipolar seesaw mechanism" [25][26]. This mechanism can be depicted by the co-occurrence of Bølling-Allerød Interstadial (B-A) (warm event) in the northern hemisphere (NH) and Antarctic Cold Reversal (ACR) (cold event) in SH ~15–13.5-kilo annum (ka) Before Present (BP) [15].

Southern Indonesia is an ideal region to investigate the past AIM changes as the contrast rainfall between the AIWM (lower) and AISM (higher) monsoon is detected here, which indicates AIM as the principal driver of rainfall [27]. While most of the previous studies in this region inferred a similar trend in AIM rainfall since the LGM [1][2][6][7][10][16], a discrepancy has been inferred between marine records on the sea around Sumba Island [5][28]. A marine record on off northwest Sumba inferred drier (wetter) condition on Mid (Late) Holocene [2] while the opposite condition was detected on the southwestern Savu Sea [28] (Figure 1) which corresponds to Mid–Late Holocene AIM rainfall in Northern Australia [13][17][18][19].

This research used the logarithmic values of elemental ratios (ln Ti/Ca and ln K/Ca) which are suggested to reflect the river runoff (terrigenous input) [1][5][8][10]. These proxies are widely applied to infer the past changes in AIM rainfall linked to the strengthening and the weakening of AISM [1][5][8][10][29]. The elemental data was obtained from X-Ray Fluorescence (XRF). XRF's main advantage lies in the direct acquisition of elemental data without complicated sample preparation, which will shorten the analysis time, even if a higher resolution is desired [30][31]. The elemental data obtained from XRF analysis are semi-quantitative and require conversion before its application to quantitative analysis [30]. The conversion of XRF-determined elemental data, which involves mass-balance and flux calculations, tends to be biased due to elemental interactions, the effect of specimen inhomogeneities, the variety of water content, and the lack of control on geometric measurements [30][32][33]. The application of elemental ratios (presented in logarithmic values) instead of single element data is suggested and proven efficient by many authors in minimizing the semi-quantitative factors of XRF analysis [32][33][34][35].

The palynological proxies (pollen and spores), which indicated climatic-induced vegetation changes [36][37][38][39][40], were also employed. Rainfall, as one of the paleoclimate parameters, can be reflected too by the abundance of pollen and spores (e.g., Poaceae and Pteridophyta) recorded in marine sediments [36][38][39]. For Southern Indonesia, the inferred rainfall is most likely AIM rainfall [28].

2. Palynological Proxies

During the Last Deglaciation, the rainfall changes were most likely connected to the high latitude climate events, i.e., early stage of Deglaciation (ED) (before ~17 ka BP), HS1 (~17–15 ka BP), ACR (~15–13.5 ka BP), and Younger Dryas (~13.5–11 ka BP) (Figure 1). Wetter periods coincided with the NH cold events (i.e., HS1 and YD) while drier periods indicated NH warm events (ED) and southern hemisphere (SH) cold events (ACR). Higher rainfall during HS1 and YD might be induced by the southward shift of the austral summer ITCZ while lower rainfall during ED and ACR were linked to the northward shift of the austral summer ITCZ [41][42] (Figure 2). The NH cooling (during HS1 and YD) enhanced the boreal winter cold surges, which pushed the ITCZ southward [41]. This mechanism also resulted in a southward shift of the boreal summer ITCZ, indicated by the lower EASM rainfall [43][44]. The lower rainfall during ACR could be explained by its co-occurrence with B-A [45] hence the northward shift of austral summer ITCZ was most likely linked to the NH warming (during ED and B-A) [41][46]. The Last Deglaciation rainfall records of ST08 were similar to d^{18}O records from Bali Gown Cave (Northwestern Australia) [47] (Figure 1), indicating the southern boundary of the austral summer ITCZ during wetter periods (HS1 and YD) [41] (Figure 2). On the other hand, D^{18}O (d^{18}O*Gs.ruber*−d^{18}O*G.bulloides*) records (which indicated AIWM changes) from off south Java [41]and terrigenous input proxies (ln Ti/Ca) from off southwest Java [48] showed the opposite rainfall changes during the ACR (Figure 1). This indicated that the records from Java region responded to B-A instead, so they hinted a considerable influence of NH cross-equatorial moisture transport and the southern boundary of the austral summer ITCZ [48] (Figure 2).

Figure 1. Drier (yellow highlight) and wetter (green highlight) periods inferred from terrigenous input proxies (**a**,**b**)[49]. Other paleoclimate records are presented for comparison: (**c**). Terrigenous input proxy (ln Ti/Ca) record of GeoB10043-3 [48], (**d**). Terrigenous input proxy (ln Ti/Ca) record of GeoB10065-7 [50], (**e**). d^{18}O record of stalagmites from Bali Gown cave (Northwestern Australia) [47], (**f**). d^{18}O *Globigerinides* (*Gs.*) *ruber*–d^{18}O *Globigerina* (*G.*) *bulloides* (D^{18}O) record of GeoB10053-7 [51], (**g**). C_{30} *n*-alkanoic fatty acids d^{13}C (d$^{13}C_{FA}$) record of GeoB10069-3 [52], (**h**). d^{18}O record of Antarctic (EPICA Dronning Maud Land/EDML) ice core [53], (**i**). d^{18}O record of Greenland (GISP2) ice core [54], (**j**). Reconstructed relative sea level from d^{18}O of Red sea benthic foraminifera [55] [56], (**k**). 10-years averaged reconstructed sunspot numbers [57], and (**l**). 20° S Dec. (austral summer) insolation (red) and 30° N Jun. (boreal summer) (blue) [58]. Data are plotted against the mean ages obtained from the age model. Black curves indicate smoothed values (exponential smoothing, df: damping factor).

Figure 2. The southern limit of the austral summer ITCZ during the drier (lower rainfall) periods (red dashed line) and during the wetter (higher) rainfall periods (blue dashed line) as suggested by this study and other previous studies [41][42][49][59][60]. Numbers show the location of ST08 ((**a**), this study) and other studies used for comparison i.e., GeoB10043-3 (**b**) [48], GeoB10053-7 (**c**) [51], GeoB10065-7 (**d**) [50], Liang Luar cave (Flores) (**e**) [60], GeoB10069-3 (**f**) [52], and Bali Gown cave (Northwestern Australia) (**g**) [46].

The YD wetter period continued until Early Holocene (EH). The abrupt rainfall increase in EH, which was inferred in off southwest Java [48], not detected on ST08. This could be linked to the relatively constant terrigenous input in off southwest Sumba due to the considerable distance from the recently drowned-Sunda Shelf, as opposed to off southwest Java [48]. $\delta^{18}O$ records on off south Java, which changes closely followed boreal summer insolation, increased during YD–EH [41] (Figure 1). This indicated the simultaneous increase of AISM and AIWM, but the effect of the strengthening AIWM and lower austral summer insolation (which should result in AISM weakening) was most likely distressed by the enhanced moisture supply related to abrupt sea-level rise [48][51][61](Figure 1).

During Mid-Holocene, the rainfall records of ST08 were similar to the ln Ti/Ca records from off northwest Sumba, which inferred drier (wetter) Mid (Late) Holocene [50] (Figure 1). Lower rainfall during the Mid Holocene (MH) (~7–3 ka BP) was most likely linked to the decrease in solar activity (hence lower sunspot numbers) [42][60], which suppressed the effect of increasing austral summer insolation [50][57]. During the Late Holocene (LH) (after ~3 ka BP), an increase in austral summer insolation and solar activity resulted in the enhancement of rainfall [50][58][57]. The southern boundary of the austral summer ITCZ during MH was most likely located around its position during ED and ACR [1,19,20] and shifted southward during LH to around its position during HS1, YD, and EH [41][50][59] (Figure 2), but their relation to solar activity and orbitally-induced austral summer insolation is still not understood [61]. The carbon isotope composition of the C_{30} *n*-alkanoic fatty acids ($d^{13}C_{FA}$) records from the southwestern Savu Sea [52] showed contradictive rainfall records (Figure 1). This contradiction might be related to the differences in climate signals recorded on terrigenous input and $d^{13}C_{FA}$ proxies. $d^{13}C_{FA}$ reflects the dry season (AIWM) water stress connected to the amount of rainfall during AIWM (AIWM rainfall) [52]. We suggest a joint analysis of terrigenous input and $d^{13}C_{FA}$ proxies from the same site in future studies to reconstruct the past changes of both AISM and AIWM rainfall, so more robust AIM rainfall records are produced.

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