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Invited review

Carbon storage and release in Indonesian peatlands since the last deglaciation

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ABSTRACT

Peatlands have been recognised as globally important carbon sinks over long timescales that produced a global, net-climatic cooling effect over the Holocene. However, little is known about the role of tropical peatlands in the global carbon cycle. We therefore determine the past rates of carbon storage and release in the Indonesian peatlands of Kalimantan and Sumatra – the largest global concentration of tropical peatlands – since 20 ka (kiloannum before present). Using a novel GIS (geographic information system) approach we provide a spatially-explicit reconstruction of peatland expansion in a series of paleogeographic maps.

Sea-level change is identified as the principal driver for peatland formation and expansion in western Indonesia as it controls both atmospheric moisture supply and the hydrological gradient on the islands. Initiation of inland peatlands in Kalimantan was coupled to periods of rapid deglacial sea-level rise with rates of over 10 mm yr^{-1} whereas coastal peatlands could only form after 7 ka when the rate of sea-level rise had slowed to 2.4 mm yr^{-1}. Falling sea levels after 5 ka led to rapid peatland expansion in coastal lowlands and a doubling of the total peatland area in western Indonesia to 131,500 km^2 between 2.3 ka and 0 ka. As a result of slow peatland expansion from 15 to 6 ka and rapid expansion afterwards the rate of annual carbon storage of all western Indonesian peatlands remained <1 Tg C yr^{-1} until 6 ka and then increased to 7.2 Tg C yr^{-1} by 0 ka. Associated with this rise in carbon storage was an exponential growth of the peat carbon pool from 0.01 Pg C by 15 ka to 23.2 Pg C at present, of which 70% is stored in coastal peatlands. In inland Kalimantan peatlands, falling sea levels together with increased El Niño activity induced an annual carbon release of 0.15 Tg C yr^{-1} from aerobic peat decay since 2 ka. Cumulative carbon losses from anaerobic decomposition do not seem to limit peat bog growth in the tropical peatlands of Indonesia. Carbon losses from Holocene peat fires are only known from the Kutai basin since 4.4 ka with an associated release of 0.1–3.6 Tg C per fire event, which never surpassed the contemporaneous annual C storage. The peatlands of western Indonesia were thus a persistent carbon sink since 15 ka but this sink was of global importance only over the past 2000 years when it likely contributed to a slower growth in atmospheric CO_2 concentrations. Currently, annual losses of carbon from peat drainage and fires are on average 28 times higher than the pre-disturbance rate of uptake implying that this carbon reservoir has recently switched from being a net carbon sink to a significant source of atmospheric carbon and is currently in danger of eradication.

1. Introduction

Tropical peatlands are an integral part of the Earth's terrestrial biosphere and have been an important component of the global carbon cycle since the Middle Paleozoic as indicated by continental coal deposits (Berner, 2003; Greb et al., 2006). Coal reservoirs are generally derived from peat deposits produced by swamp forests, in which woody material only partially decomposes in a waterlogged setting (Hedges et al., 1985). The equatorial peatlands of Indonesia, Malaysia, and Brunei (Southeast Asia, Fig. 1) are generally considered to represent important modern analogues for past coal forming environments (Cobb and Cecil, 1993).

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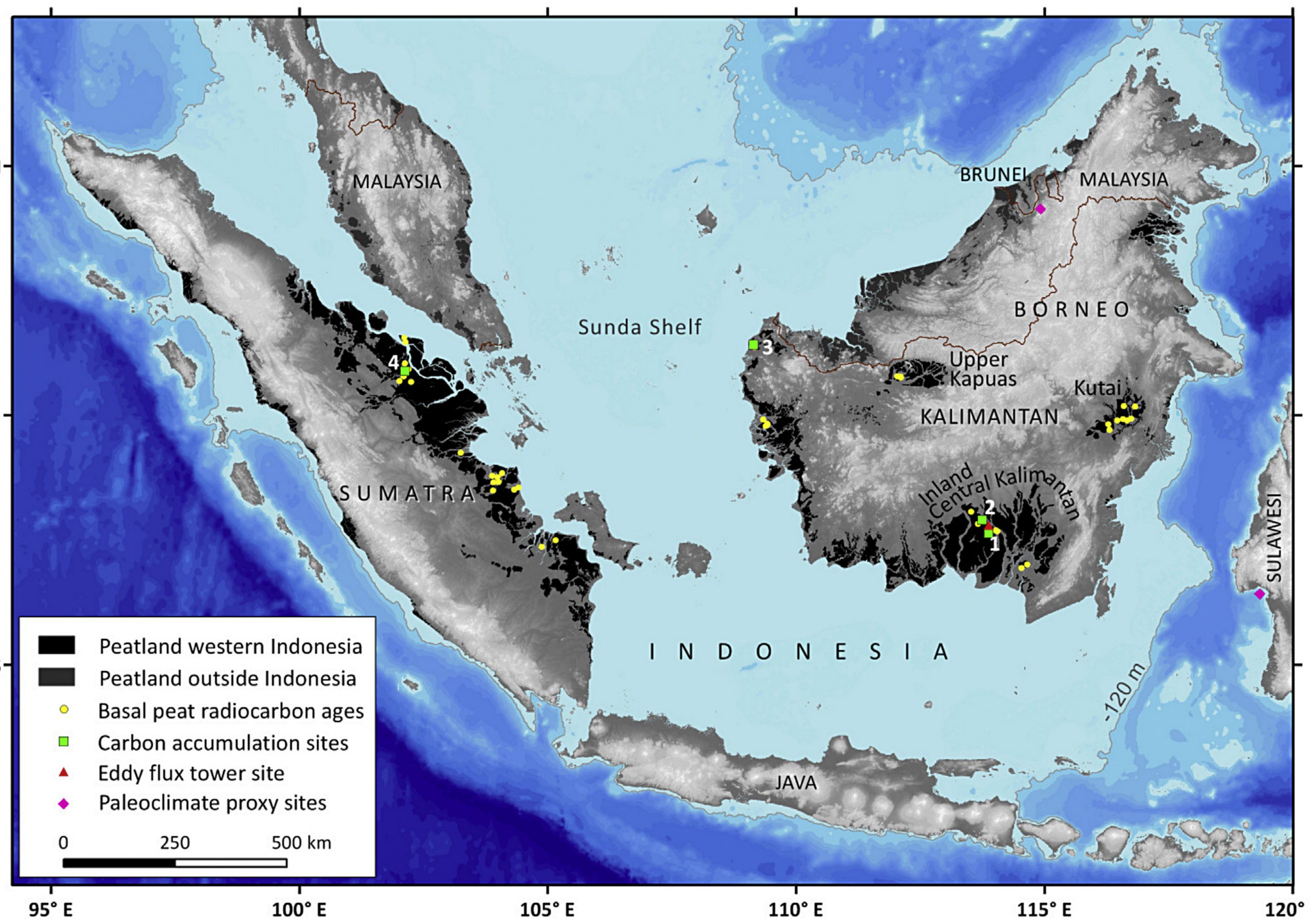


Fig. 1. Peatland distribution in the Sunda-region. Black areas represent the lowland peatlands of western Indonesia, dark grey areas peatlands in Malaysia, Brunei and Thailand. Locations of basal peat dates are marked by yellow dots. Locations of carbon accumulation records from western Indonesia are marked by green squares and numbered: 1) Sebangau, 2) Palangka Raya, 3) Teluk Keramat, 4) Siak Kanan. The eddy flux tower site of Hirano et al. (2007, 2012) is marked by a red triangle and the paleoclimate proxy sites of Partin et al. (2007) in northern Borneo and Tierney et al. (2012) off Sulawesi by pink diamonds.

Peatlands are known from equatorial Southeast Asia since the Paleocene, with the first ombrotrophic peat formations developing in the Late Oligocene (Morley, 2012, 2013). Domed peatlands were widespread in the Miocene and pollen records of their coal deposits reflect floristic assemblages strikingly similar to modern peat swamps of Southeast Asia (Anderson and Muller, 1975; Moore and Hilbert, 1992; Demchuk and Moore, 1993; Morley, 2012, 2013). Modern peat swamp forests are thus an ancient ecosystem that has been an element of Southeast Asia's vegetation for 15–20 million years (Morley, 2012, 2013). During the Cenozoic, Southeast Asian peat swamp forests have effectively transferred carbon from the atmosphere into terrestrial peat deposits, which in certain tectonic settings were eventually buried and preserved within the geologic rock reservoir. The modern peatlands of Indonesia therefore provide a unique, but rapidly disappearing opportunity to study the influence of peat swamp forests, as analogues of ancient coal forming ecosystems, on the global carbon cycle over millennial timescales.

Approximately half of all known tropical peatlands and 84% of those in Southeast Asia are located in Indonesia (Page et al., 2011). Indonesian peatlands are relatively young geologic deposits that mostly originated during the past 14,000 years (Dommain et al., 2011), but spatially and temporally explicit information on their origin and development is rare. Currently, estimates are only available for 1) the current carbon pool that is stored in Indonesian peatlands, 2) the age of initiation for selected peatlands and 3) the past rates of peat accumulation for an even more limited number of sites. A more detailed spatial and chronological context to these estimates is needed to understand how the carbon fluxes and the build-up of the carbon pool in Indonesian peatlands were affected by changes in sea level, climate, and paleo-drainage networks.

Sea level has been invoked as an important driver for the formation of Southeast Asian peatlands during the Cenozoic (Dommain et al., 2011; Morley, 2012). Dommain et al. (2011) suggested that both the rise and subsequent fall in sea level during the Holocene induced peatland initiation across western Indonesia, Malaysia, and Brunei. Moreover, sea-level changes over the Sunda Shelf seem to be particularly important in regulating the regional moisture availability on glacial–interglacial timescales (DiNezio and Tierney, 2013). However, a more quantitative approach is needed to determine the influence of sea-level changes on the spatial expansion of peatlands in Indonesia and to identify potential thresholds that trigger their initiation.

The carbon balance of tropical peatlands also seems to be strongly influenced by rainfall seasonality, including variability in the El Niño-Southern Oscillation (ENSO), as shown by modern carbon flux studies (Hirano et al., 2007, 2012). However, no general understanding exists on how rainfall and other climate factors influenced the long-term carbon balance of tropical peatlands. Because Indonesian peatlands are generally dome-shaped landforms that solely rely on atmospheric sources of water (Dommain et al., 2010) their peat layers should preserve a sensitive record of the carbon cycling responses to past climatic changes.

Peatlands are believed to be important in the global carbon cycle as significant long-term sinks of atmospheric CO_2 (Frolking and Roulet, 2007; Kleinen et al., 2010; Yu, 2011). Whereas the age of peatland initiation across western Indonesia is fairly well known, reliable estimates of past areal peatland extent are thus far lacking. Without reliable data of peatland expansion the role of tropical peatlands in regulating atmospheric concentrations of CO_2 since the last deglaciation cannot be accurately assessed.

In this paper we explore the importance of deglacial sea-level changes for peatland development by specifying spatio-temporal patterns of peatland initiation and by quantifying their expansion in relation to past sea levels. We illustrate these relationships in a series of paleogeographic maps in millennial time slices. Furthermore, the paper addresses the importance of past changes in temperature, rainfall supply, and seasonality, including variations in ENSO, for long-term peatland carbon accumulation. In order to clarify the role of Indonesian peatlands in the global carbon cycle during the Holocene we quantify rates of carbon storage and the changing size of the Indonesian carbon pool over time. We specifically relate changes in carbon storage of Indonesian peatlands to past variations in atmospheric CO_2 concentration. As a last point, we compare ongoing carbon losses from peat fires and from peat decomposition with natural carbon storage in order to address the question whether Indonesian peatlands still function as a carbon sink and thus as a potential coal-forming environment.

2. Study region

This study focuses on the lowland peatlands of western Indonesia, which occur on the island of Sumatra and the Indonesian part of the island of Borneo known as Kalimantan. Both islands are part of the continental Sunda Shelf and formed a coherent land mass with mainland Asia during the Last Glacial Maximum (26.5–19 ka; kiloannum = 1000 calibrated years before present, where present is 1950) when sea levels were around 120 m lower than today (Fig. 1).

The Indonesian archipelago lies in the heart of the Indo-Pacific Warm Pool (IPWP), a global center of atmospheric deep convection (Yan et al., 1992). Due to year-round high sea-surface temperatures (SST) of over 28 °C convective activity is high, resulting in annual rainfall of between 2 and 4 m (Aldrian and Susanto, 2003). The climate is governed by the cross-equatorial Austral-Asian monsoon system, the Intertropical Convergence Zone (ITCZ), and the Walker circulation. The seasonal cycle in insolation affects the direction and strength of monsoon circulation and the position of the Intertropical Convergence Zone during the year. Interannual variability in precipitation over western Indonesia is significantly correlated to the El Niño-Southern Oscillation (ENSO; Hendon, 2003) and particularly in Sumatra also strongly influenced by the Indian Ocean Dipole (IOD) (Saji et al., 1999). El Niño events and/or positive IOD modes cause prolonged episodes of extreme rainfall deficit and mark the major episodes of human induced peat and forest fires today (van der Werf et al., 2008; Field et al., 2009). Indonesia has a diurnal climate (*Tageszeitenklima* sensu Troll, 1943), in which daily temperature fluctuations exceed annual fluctuations considerably. Mean monthly temperatures are high and vary between 26 and 27 °C.

The vast majority of peatlands in Indonesia are restricted to lowlands with elevations of less than 50 m above sea level (a.s.l.). Peat covered land forms in these lowlands are generally coastal alluvial plains and prograding river deltas of Holocene age. Such land forms have their greatest concentration along the east coast of Sumatra and the south and west coast of Kalimantan (Fig. 1). Besides in this coastal region, lowlands containing substantial areas of peatland also occur in inland Kalimantan. The Upper Kapuas basin with an elevation of 25–50 m a.s.l. in West Kalimantan (Anshari et al., 2001, 2004; ca 180 km inland) and the Kutai basin with elevations of less than 5–25 m a.s.l. in East Kalimantan (Hope et al., 2005; ca 120 km inland; Fig. 1) are both characterized by a mosaic of floodplain lakes and minero- and ombrotrophic peatlands. In southern Kalimantan between 30 and 80 km from the coast the coastal alluvial plain grades into a sandy Pleistocene podzol plain that is draped with domed peatlands up to 230 km inland. This peat covered podzol plain with an altitude of 5–50 m a.s.l. is the peatland region of inland Central Kalimantan (Dommain et al., 2011, Fig. 1). Sieffermann et al. (1987, 1988) called the central elevated interfluves of this region “high peat” and the adjacent peat filled river valleys “basin peat”.

Based on this geographic pattern of peatland distribution we distinguished five peatland regions of western Indonesia: 1) coastal peatlands of Sumatra, 2) coastal peatlands of Kalimantan, 3) inland peatlands of Central Kalimantan, 4) inland peatlands of the Kutai basin, and 5) inland peatlands of the Upper Kapuas basin. Peatlands of the two coastal regions have similar basal ages and nearly identical long-term rates of peat accumulation (Dommain et al., 2011). The inland peatland regions differ markedly with respect to age, mode of origin, underlying substrate and long-term rates of peat accumulation from the coastal peatland regions (Anshari et al., 2001; Dommain et al., 2011).

The lowland peatlands of Indonesia are covered with peat swamp forest that on ombrotrophic peat domes is arranged into concentric forest communities (Anderson, 1983; Bruenig, 1990). Indonesian peat swamp forests are floristically very diverse in comparison to northern peatlands. For example, Simbolon and Mirmanto (2000) report 310 plant species from Central Kalimantan peat swamp forests and Brady (1997a) lists 144 tree species from peat swamp forests of Sumatra. The aboveground woody biomass of Bornean peat swamp forests ranges between ca 260 and 400 t ha^{-1} (Verwer and van der Meer, 2010).

3. Materials and methods

3.1. Late-Quaternary sea level and paleo-drainage reconstructions of the Sunda Shelf

We reconstructed changes in land-sea distribution on the Sunda Shelf in a series of maps at 1000 year intervals from the end of the Last Glacial Maximum (21 ka) to the present (0 ka). This reconstruction is based on the most recent sea-level curve for the Sunda Shelf by Hanebuth et al. (2011; shown in Fig. 7) and the bathymetric and topographic ETOPO 1 Global Relief Model (http://www.ngdc.noaa.gov/mgg/global/global.html). ETOPO 1 has a horizontal resolution of one arc minute (ca 1.85 km at the equator) and a vertical resolution of one metre (Amante and Eakins, 2009). Erosion and especially sedimentation will have changed the sea floor topography of the Sunda Shelf since the Late Pleistocene, but measured sedimentation rates on the open Sunda Shelf are generally quite low (~10 cm ka^{-1}, Hanebuth and Stattegger, 2003; Hanebuth et al., 2011). Moreover, post-glacial crustal movement is minimal for the Sundaland core (Tjia, 1996). Therefore assigning past sea levels to the modern bathymetric contours provides a reasonable approximation of the Lateglacial and Holocene extent of exposed land surfaces (Voris, 2000; Sathiamurthy and Voris, 2006; Hanebuth et al., 2011; Hall, 2012). Based on the ETOPO1 digital relief model we modelled the paleo-river network and paleo-drainage basins of the exposed shelf and adjacent uplands using the hydrology tool box of ArcMap 10.1 (ESRI, 2011). The modern land surface is visualized with Shuttle Radar Topography

Mission (SRTM) data with 90 m horizontal and 1 m vertical resolution.

3.2. Reconstructing areal peatland expansion

We reconstructed the origin and expansion of peatlands for the modern land area of Sumatra and Kalimantan (i.e. Indonesian Borneo) since 20 ka at intervals of 1000 years. These reconstructions are based on the GIS versions of the Wetlands International peat atlas for Sumatra (Wahyunto et al., 2003) and Kalimantan (Wahyunto et al., 2004). These two GIS atlases contain polygons ($n = 2669$) mapping the distribution, extent (area) and depth of peatland as well as peat type and physico-chemical peat properties for each district of Sumatra and Kalimantan (Wahyunto and Suryadiputra, 2008). We discarded 47 polygons covering <10 km^2 located at high altitudes. The atlas for Sumatra distinguishes four classes of peat depth, viz. 50–100 cm, 100–200 cm, 200–400 cm and >400 cm, whereas the atlas for Kalimantan distinguishes five classes (viz. 50–100, 100–200, 200–400, 400–800 and 800–1200 cm) (Wahyunto and Suryadiputra, 2008). Each polygon represents a spatially distinct area that falls into a single peat depth class.

Using ArcMap 10.1, the extent and depth of the peatland polygons of the atlases were updated with digitized versions of maps published by Sieffermann et al. (1988), Brady (1997a), Hope et al. (2005), Wösten et al. (2006) and Jaenicke et al. (2008) and checked against recent Landsat imagery. In addition, we used field data collected by Wetlands International – Indonesia Programme and R. Dommain in Central Kalimantan and Sumatra in 2008 and 2010.

The basal age of the mapped polygons was determined in two ways. First, available basal radiocarbon dates for Indonesian peatlands (Dommain et al., 2011; with additional dates from: Diemont and Pons, 1992; Brady, 1997a; Hope et al., 2005) were georeferenced and added to the peatland map. In ArcMap 10.1 we then assigned the calibrated basal radiocarbon ages ($n = 54$, Fig. 1) to respective peatland polygons (covering 15% of the total peatland area). Second, the basal age of the remaining, undated polygons was calculated by dividing the peat depth with the mean long-term rate of peat accumulation published in Dommain et al. (2011). Applied rates of peat accumulation take into account the following regional differences between the existing lowland peatland types: 1) 1.77 mm yr^{-1} for coastal Sumatra and Borneo (Kalimantan and Malaysian Sarawak), 2) 0.54 mm yr^{-1} for inland Central Kalimantan, 3) 1.89 mm yr^{-1} for the Kutai basin, and 4) 0.46 mm yr^{-1} for the Upper Kapuas basin. The latter rate was determined by averaging the peat accumulation rates from the dated peat cores of Anshari et al. (2001, 2012). In other words, we applied regionally specific transfer functions to derive a basal age for each undated polygon.

Given that the Wetlands International peat atlas provides only depth classes for each polygon, we tested which specific depth (minimum, mean, maximum of the class) yields the most likely age. To this end we applied the transfer functions to the dated polygons and compared the resulting age estimates with the actual radiocarbon dates. This comparison showed that the maximum depth of the depth classes provided the best age estimates for coastal peatlands, the Kutai basin and the Upper Kapuas basin. The mean depth provided the best age estimate for inland Central Kalimantan. Some polygons from Central Kalimantan that lack basal radiocarbon ages are apparently truncated as indicated by (old) near-surface radiocarbon dates. To estimate their basal age we applied the regional transfer function and added the result to the available near surface age. Within each peatland region, all undated polygons with the same peat depth class are assigned the same basal age and the reconstructed peatland expansion is inevitably stepwise.

3.3. Reconstructing carbon storage

3.3.1. General considerations

Dommain et al. (2011) estimated mean rates of Holocene carbon accumulation in coastal Sumatra and Borneo to be 77 g C m^{-2} yr^{-1}, in contrast to 31.3 g C m^{-2} yr^{-1} in inland Central Kalimantan. To estimate carbon accumulation rates (CAR) for the Kutai basin for which data on dry bulk density and carbon content for peat are still not available (G. Hope, pers. com.), we applied a carbon density of 0.064 g C cm^{-3} based on previously analysed cores from inland Bornean peatlands in Central Kalimantan (Dommain et al., 2011) and Danau Sentarum/Upper Kapuas (Warren et al., 2012). We then multiplied this carbon density with previously published rates of peat accumulation from Kutai (Hope et al., 2005; Dommain et al., 2011). For the Upper Kapuas basin we determined the regional mean CAR by averaging cores A and B of Anshari (2010) (with ^{14}C dates from Anshari et al., 2012) and core HN3 of Anshari et al. (2001). A mean carbon density of 0.066 g C cm^{-3}, taken from Warren et al. (2012), was used for the latter core.

A mean rate of carbon accumulation was determined for each millennium for each of the five peatland regions by averaging annual rates. Determining carbon accumulation rates for the most recent millennium is not trivial, because the surface of a peat profile (where the age is 0 ka) is not necessarily equivalent to the ground surface, but is instead located at some depth beneath the floor of the peat swamp forest. The base of the trees in these tropical swamp forests forms an interlocking continuum with the roots that can extend over a metre below the apparent land surface depending upon the depth of the water table (Furukawa, 1988, 1994). In order to compare CAR for the most recent millennium with that of the preceding millennia we disregarded the upper living root mat.

For coastal Sumatran peatlands, Brady (1997a, b) determined the thickness of the living root mat by radiocarbon dating and consistently found modern radiocarbon ages from 10 to 20 cm depth and pre-1950 radiocarbon ages for depths below 30 cm. In order to avoid overestimating recent carbon accumulation we assumed that the modern peat surface (i.e. 0 ka) is located at a depth of 30 cm for coastal and Kutai basin sites. In inland Central Kalimantan the root mat-peat interface is substantially deeper owing to the generally lower water table in these peatlands (e.g. Moore et al., 1996; Hirano et al., 2012). As a result peatlands from inland Central Kalimantan have modern calibrated radiocarbon ages down to a depth of ~1 m below the surface (Weiss et al., 2002; Page et al., 2004; Wüst et al., 2008). For three inland Central Kalimantan peat profiles we rejected the modern, uppermost radiocarbon dates and extrapolated the peat accumulation rate derived from the two preceding (non-modern) radiocarbon dates to establish the 0 ka depth level. This level was located at 79, 96 and 105 cm below the ground surface respectively.

3.3.2. Carbon accumulation in individual peatlands

Detailed records of organic matter and carbon accumulation rates (CAR) from western Indonesian peatlands have been reported for four sites by Diemont and Supardi (1987a), Neuzil (1997), and Page et al. (2004); summarized in Dommain et al. (2011) (Fig. 1). We re-calculated annual CAR from these records after re-calibrating all radiocarbon dates with the IntCal 09 calibration curve (Reimer et al., 2009) in Calib 6.0.1 and by using the weighted averages of the probability distribution functions as point age estimates (Telford et al., 2004). The annual CAR of each profile was summed to determine cumulative carbon mass accumulation (in Mg C ha^{-1}, Mg = megagram = 10^6 g, equal to one metric tonne).

3.3.3. Carbon storage

We define carbon storage as the rate of annual carbon accumulation integrated over area. Rates of carbon storage were determined in units of Tg C yr^{-1} (Tg = teragram = 10^{12} g, equal to one megatonne) for the three inland peatland regions, coastal Sumatra and coastal Kalimantan, and all of western Indonesia. First, carbon storage was calculated for single peatlands for which records of CAR ($n = 6$) or rates of peat accumulation along with mean values of carbon density ($n = 26$) were available (covering 13% of the total peatland area). Second, for all remaining peatlands we used 1000 year mean values of CAR for each region (Table 1). Multiplying past rates of carbon accumulation with contemporaneous extent of peat cover yielded past rates of carbon storage per site. For each peatland region we derived annual rates of carbon storage as the sum of all site specific values.

Summing up the total annual rates of carbon storage since the first peatland formed yields the current carbon pool. The size of the peat carbon pools (in Pg C = petagram = 10^{15} g, equal to one gigatonne) is reported for each peatland region and for all of western Indonesia for the end of each millennium since 15 ka (Table 1, Fig. 4).

3.4. Reconstructing carbon release

During the Late Holocene, peatlands in inland Borneo apparently lost surface peat by both peat fires and climatically-driven peat decomposition (Hope et al., 2005; Dommain et al., 2011). Reconstructing past rates of carbon release from these sources is challenging because the chronology and magnitude of degradation are difficult to resolve from stratigraphic profiles. A number of peat cores ($n = 8$) from undrained sites in inland Central Kalimantan are truncated with near surface (0–120 cm) ages of between ~7.5-to-5 ka (Dommain et al., 2011). Truncated profiles with surface ages of 5.2 and 6.5 ka have also been reported from the Nung peat domes of the Upper Kapuas basin ($n = 2$, Anshari et al., 2012). Sieffermann et al. (1988) proposed that the truncated peatlands of Central Kalimantan continued to accumulate peat until about 2 ka and have since that time constantly lost surface peat. Following this hypothesis we estimated additional carbon accumulation from the truncation age of each core until 2 ka by applying the contemporaneous mean CAR from Central Kalimantan or the Upper Kapuas basin, respectively. After 2 ka all of this additional carbon is assumed to be gradually released at a constant rate. All truncated Central Kalimantan profiles are located within the "high peat" area of Sieffermann et al. (1987, 1988). We therefore assumed that all polygons with truncated profiles of the high peat area of Central Kalimantan and from the Upper Kapuas basin have released carbon since 2 ka.

Truncated, burnt peat profiles have been reported from the Kutai basin that are now overlain by clayey lake or floodplain sediments (Diemont and Pons, 1992; Hope et al., 2005). Based on recent observations (Wösten et al., 2006; van Eijk et al., 2009), we conservatively assumed that 1 m of peat was lost from these truncated profiles due to recurrent fires to create permanently flooded conditions. We then estimated the time required for one metre of peat to accumulate based on the peat accumulation rates calculated for these peatlands prior to truncation. This time interval

Table 1
Changes in peatland area, carbon accumulation rates, rates of regional carbon storage and release, and of sizes in carbon pools in western Indonesia.

Peatland region		Age (ka)															
		0	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Inland Central Kalimantan	Area (1000 km^2)	17.88	17.88	15.27	13.15	12.70	12.43	10.45	10.11	9.02	8.81	7.44	6.28	3.75	2.79	2.79	0.81
	1000 yr mean CAR (g C m^{-2} yr^{-1})	18.15	16.94	21.97	19.78	18.78	24.42	28.82	42.25	53.17	57.68	42.81	35.39	28.83	36.86	20.23	1.19
	C storage (Tg C yr^{-1})	0.25	0.23	0.33	0.25	0.23	0.27	0.30	0.44	0.50	0.64	0.35	0.23	0.13	0.17	0.18	0.00
	C release (Tg C yr^{-1})	0.14	0.14														
	C pool (Pg C)	3.83	3.73	3.67	3.35	3.10	2.87	2.62	2.32	1.93	1.46	0.93	0.58	0.39	0.27	0.09	0.00
Upper Kapuas	Area (1000 km^2)	4.62	4.62	4.62	3.02	3.02	2.15	2.15	2.15	2.15	1.86	1.86	1.86	1.86	1.86	1.39	1.25
	1000 yr mean CAR (g C m^{-2} yr^{-1})	22.58	14.11	6.75	1.61	21.20	27.08	19.78	29.25	29.19	29.30	31.62	31.29	30.75	26.72	19.04	0.78
	C storage (Tg C yr^{-1})	0.09	0.06	0.03	0.00	0.07	0.07	0.05	0.07	0.07	0.07	0.07	0.07	0.07	0.06	0.03	0.00
	C release (Tg C yr^{-1})	0.01	0.01														
	C pool (Pg C)	0.84	0.76	0.71	0.69	0.68	0.63	0.56	0.51	0.44	0.37	0.30	0.23	0.16	0.09	0.04	0.01
Kutai	Area (1000 km^2)	4.55	3.35	3.35	3.35	3.35	3.14	2.96	1.11	0.41							
	1000 yr mean CAR (g C m^{-2} yr^{-1})	64.03	58.32	87.02	83.66	87.72	111.41	177.89	213.89	145.02							
	C storage (Tg C yr^{-1})	0.25	0.16	0.29	0.27	0.29	0.34	0.41	0.30	0.06							
	C release (Tg C yr^{-1})		0.01		0.01	0.003											
	C pool (Pg C)	2.01	1.80	1.64	1.35	1.09	0.79	0.46	0.16	0.02							
Coastal Kalimantan	Area (1000 km^2)	33.46	22.70	13.60	5.97	5.97	0.50	0.50									
	1000 yr mean CAR (g C m^{-2} yr^{-1})	64.38	68.30	66.21	79.47	91.44	67.40	84.50									
	C storage (Tg C yr^{-1})	2.16	1.55	0.90	0.48	0.56	0.03	0.04									
	C pool (Pg C)	4.28	2.42	1.39	0.85	0.37	0.07	0.03									
Coastal Sumatra	Area (1000 km^2)	71.03	53.32	31.74	22.06	21.39	9.40	0.69	0.09								
	1000 yr mean CAR (g C m^{-2} yr^{-1})	64.38	68.30	66.21	79.47	91.44	67.40	84.50	86.93								
	C storage (Tg C yr^{-1})	4.47	3.59	2.13	1.97	2.43	0.91	0.06	0.01								
	C pool (Pg C)	12.14	8.29	5.95	4.29	2.34	0.43	0.04	0.01								
Western Indonesia (sum)	Area (1000 km^2)	131.54	101.87	68.58	47.55	46.43	27.62	16.74	13.45	11.59	10.67	9.30	8.14	5.61	4.65	4.18	2.06
	mean CAR (g C m^{-2} yr^{-1})[a]	54.84	55.19	53.94	63.00	77.13	58.67	51.94	60.69	54.84	66.11	44.96	37.08	35.26	49.42	49.20	0.94
	C storage (Tg C yr^{-1})	7.21	5.61	3.68	2.99	3.58	1.62	0.87	0.82	0.64	0.71	0.42	0.30	0.20	0.23	0.21	0.00
	C release (Tg C yr^{-1})	0.15	0.16		0.01	0.003											
	C pool (Pg C)	23.10	17.00	13.35	10.52	7.57	4.79	3.72	3.00	2.39	1.83	1.23	0.80	0.55	0.36	0.13	0.01

[a] Represents area-weighted means at 1 ka boundaries (not intervals). Note that coastal Kalimantan and Sumatra have the same mean CAR values.

was added to the age of the truncation surface to date the occurrence of the first fire. The fire related carbon losses are deducted from the carbon storage value for the millennium in which they occur. The amount of regional carbon release from fires was estimated based on the area of GIS peatland and lake polygons that match with the location of truncated cores.

4. Results

4.1. Peatland expansion over time

Peat formation in western Indonesia was slow and gradual during the deglacial period and Early Holocene but then increased rapidly after 6 ka. Between 19.9 ka and 11.7 ka about 5% of the total present peatland area (5600 km^2) had formed at a rate of approximately 700 km^2 ka^{-1}. Between 11.7 ka and 5.6 ka peatland expansion was more than twice as fast at about 1900 km^2 ka^{-1}. By 5.6 ka there were about 17,400 km^2 of peatlands representing 13% of the present day area. Subsequently, the peatland area increased rapidly to cover over 46,000 km^2 (35%) by 4.3 ka and continued to increase at a rate of about 28,000 km^2 per millennium from 3-to-0 ka.

This reconstruction shows that only 50% of the total modern peatland area existed at 2.3 ka and this area then doubled over the next 2000 years to produce a total area of 131,540 km^2. More than half of this peatland area is located in Sumatra, whereas Kalimantan has a total of 60,510 km^2, of which about half is found in coastal lowlands and ca 27,000 km^2 in inland areas. Today, coastal peatlands dominate the peatland area of western Indonesia covering 104,480 km^2 (80%).

With the beginning of the last deglaciation at 19.9 ka the first western Indonesian peatlands formed in inland Borneo in the Upper Kapuas basin. Peatlands expanded in this region to nearly 2000 km^2 by 13.7 ka, which corresponds to 40% of the Upper Kapuas' present peatland area and represented 40% of the total peatland cover in western Indonesia at that time. There was little further expansion in these peatlands through the Early Holocene, but by 2 ka they had increased to their modern-day area of 4620 km^2, which represents only 3.5% of the total western Indonesian peatland area (Table 1, Figs. 1, 2).

In inland Central Kalimantan peatland formation started at 18.5 ka and peatlands covered 2800 km^2 by 14.5 ka, corresponding to approximately 65% of the contemporaneous total peatland area. By 8 ka half of the present peatland area (9000 km^2) in this region had already formed, representing about 80% of the overall peatland area at that time. Afterwards addition of new peatland area was rather slow, but between 3 and 1 ka peatlands expanded over an additional area of 4700 km^2 to a total of 17,885 km^2. Inland Central Kalimantan is thus clearly the largest inland peatland region in Kalimantan and after the coastal peatlands of Sumatra and Kalimantan, respectively, the third largest peatland region in western Indonesia, representing 13.6% of the total peatland area (Table 1, Figs. 1, 2).

In the Kutai basin peat started forming at 8.3 ka and peatlands expanded rapidly until ca 6 ka when almost 3000 km^2 ($\sim$65%) of Kutai peatlands had formed. After a period of stagnation, the peatlands again expanded after 0.5 ka to reach a total area of 4555 km^2. This area is nearly identical in size to that of the Upper Kapuas peatland region and corresponds to 3.5% of the total western Indonesia peatland area (Table 1, Figs. 1, 2).

In comparison to the inland peatland areas the coastal peatlands of Kalimantan and Sumatra formed substantially later, but expanded to cover a much larger area. In Kalimantan the first coastal peatlands formed at 6.7 ka. Areal expansion was small until 4.5 ka when the peatland area increased to cover over 5700 km^2,

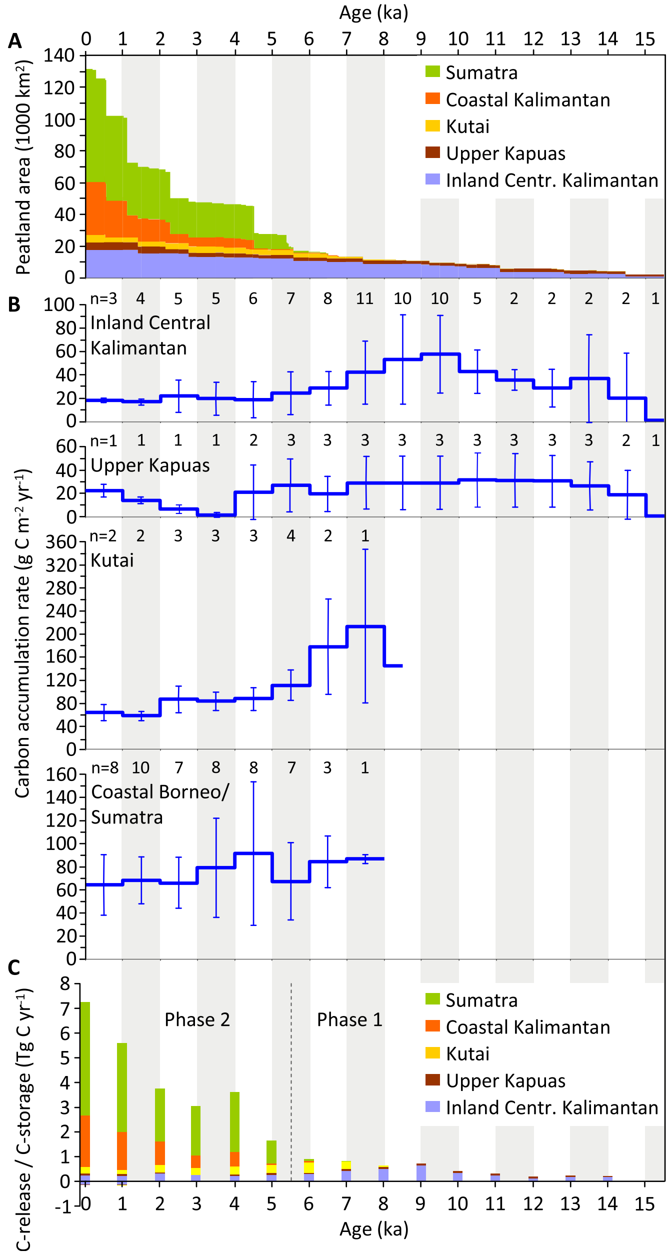


Fig. 2. Reconstructions for the five major peatland regions of western Indonesia for the past 15,000 years of (A) changing peatland area, (B) mean carbon accumulation rates in 1000 year intervals (with standard deviation and number of cores), and (C) rates of carbon storage and release. Carbon storage and release are expressed as annual rates at each millennial boundary.

surpassing the peatland areas of Kutai and Upper Kapuas. Afterwards coastal peatlands of Kalimantan only increased marginally until 2.3-to-0.5 ka when an additional 27,000 km^2 of peatland formed producing a total area of 33,455 km^2 of coastal peatland in Kalimantan. Coastal Kalimantan is the second largest peatland region in western Indonesia today, representing 25% of the total peatland area (Table 1, Figs. 1, 2).

In Sumatra coastal peatland began forming at 7.7 ka but areal expansion remained minimal until 5.4 ka when peatlands formed and spread over 7800 km^2. With the formation of 12,000 km^2 peatland between 4.7 ka and 4.5 ka the overall peat cover grew to over 21,000 km^2, which corresponds to 30% of the current peatland area in Sumatra and to nearly 50% of the western Indonesian peatland area at the time. However, the largest increase in peatland area occurred after 1.1 ka when an additional 38,000 km^2 formed producing a total peatland area of 71,025 km^2 in the coastal lowlands of Sumatra. These coastal lowlands currently comprise the largest peatland region in western Indonesia representing more than double the area of coastal peatlands on Kalimantan and 54% of the total peatland cover of western Indonesia (Table 1, Figs. 1 and 2).

4.2. Rates of carbon accumulation in individual peatlands

Records of carbon accumulation are available for two inland peatlands in Central Kalimantan (Sebangau and Palangka Raya) and for two coastal peatlands (Siak Kanan in Sumatra and Teluk Keramat in Borneo; Fig. 1). At Sebangau (Fig. 3a) carbon accumulation rates were highest between 9.5 and 8.5 ka with maximum rates of 121 g C m^{-2} yr^{-1} at 9.1 ka and of 130 g C m^{-2} yr^{-1} at 8.6 ka. After 7.7 ka CAR remained low between about 30 and 5 g C m^{-2} yr^{-1} producing a convex carbon mass accumulation curve (quadratic polynomial fit, $r^2 = 0.953$). The total accumulated carbon mass after 12,800 years amounts to 2950 Mg C ha^{-1}.

The Palangka Raya peatland (Fig. 3a) is a special case because it is truncated and has a current near-surface age of about 4.5 ka at −75 cm. Carbon accumulation rates were highest between 7.6 ka and 7 ka (91 g C m^{-2} yr^{-1}) but afterwards varied between 55 and 65 g C m^{-2} yr^{-1}. The carbon mass accumulation curve is linear until 4.5 ka ($r^2 = 0.984$) or slightly concave (quadratic polynomial fit, $r^2 = 0.995$). However, the completion of the carbon mass accumulation curve to 0 ka changes the curve to a convex shape (scenario (a) in Fig. 3a). This peatland currently stores a carbon mass of 4370 Mg C ha^{-1}.

The coastal peatland Siak Kanan (Sumatra, Fig. 3b) shows generally high rates of carbon accumulation, which fluctuate around a mean of ca 70 g C m^{-2} yr^{-1}. These long-term rates include centennial-scale peaks in carbon accumulation of 204 g C m^{-2} yr^{-1} between 4.1 and 4 ka, 113 g C m^{-2} yr^{-1} between 3.1 and 3 ka, and about 110 g C m^{-2} yr^{-1} between 1.3 and 0.8 ka. The resulting carbon accumulation curve is linear ($r^2 = 0.997$) to slightly concave ($r^2 = 0.998$) producing a total carbon mass of 3800 Mg C ha^{-1} after about 5100 years of peat growth.

Accumulation at the coastal site Teluk Keramat (Kalimantan, Fig. 3b) also fluctuates around an average of about 70 g C m^{-2} yr^{-1}, but with much less amplitude. A maximum CAR of 98 g C m^{-2} yr^{-1} occurs between 0.5 and ~0.2 ka. The corresponding carbon mass accumulation curve is linear ($r^2 = 0.995$) to slightly concave

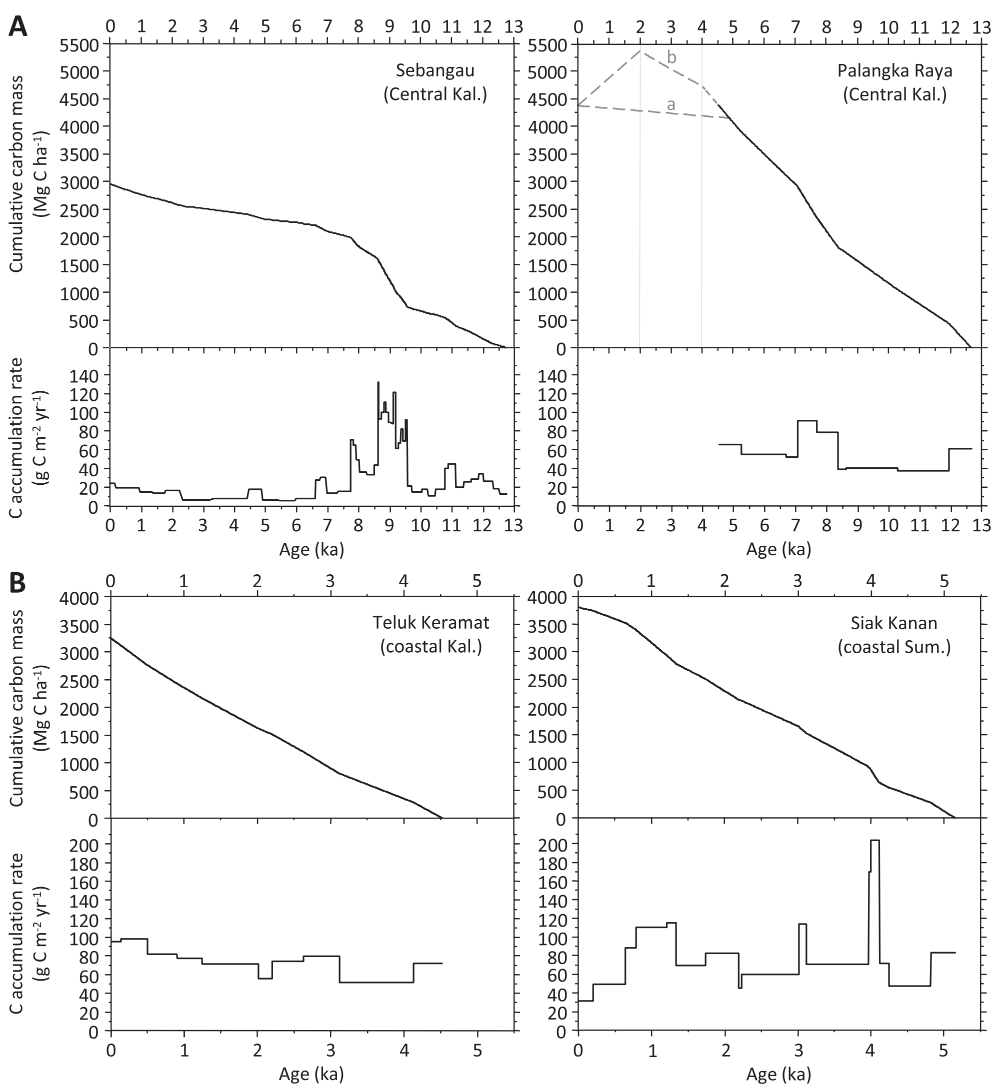


Fig. 3. Carbon accumulation rates and cumulative carbon mass accumulation over time for (A) two peatlands from inland Central Kalimantan and (B) two coastal peatlands at the bottom. Note the differences in scale. The data for Sebangau are from Page et al. (2004), for Palangka Raya and Teluk Keramat from Neuzil (1997) and for Siak Kanan from Diemont and Supardi (1987a) and Supardi et al. (1993). Location of the sites is marked in Fig. 1.

(quadratic polynomial, $r^2 = 0.999$). The total accumulated carbon mass of this 4500 year old peatland is 3250 Mg C ha^{-1}. In comparison to the two inland peatlands from Central Kalimantan carbon accumulation in the two (substantially younger) coastal peatlands is three to four times higher (Fig. 3).

4.3. Regional carbon storage and release and carbon pools

Average regional rates of carbon accumulation, annual rates of carbon storage and release, and the size of the regional and total carbon pool(s) in western Indonesia are summarized in Table 1 and in Figs. 2 and 4. The rate of annual carbon storage over all western Indonesian peatlands increased from 0.002 Tg C yr^{-1} by 15 ka to 7.2 Tg C yr^{-1} by 0 ka. This increase in carbon storage led to the build up of a regional peatland carbon pool that currently contains 23.2 Pg C (Fig. 4). The western Indonesian peat carbon pool initially accumulated slowly with only 2% (0.55 Pg C) of the current pool present prior to the Holocene and 10% (2.4 Pg C) by 8 ka limited solely to the inland peatlands of Kalimantan. Afterwards the total carbon pool accumulated more rapidly reaching 21% (4.8 Pg C) by 5 ka and 50% (12 Pg C) by about 2.5 ka. The carbon pool then doubled in size over the past 2000 years. Overall, the reconstructed growth of the peat carbon pool at the millennial timescale was nearly exponential over the Holocene with a factor of 1.4 ± 0.1 (Fig. 4). The evolution of this peat carbon pool in western Indonesia is the product of the individual carbon storage histories of each peatland region as detailed below.

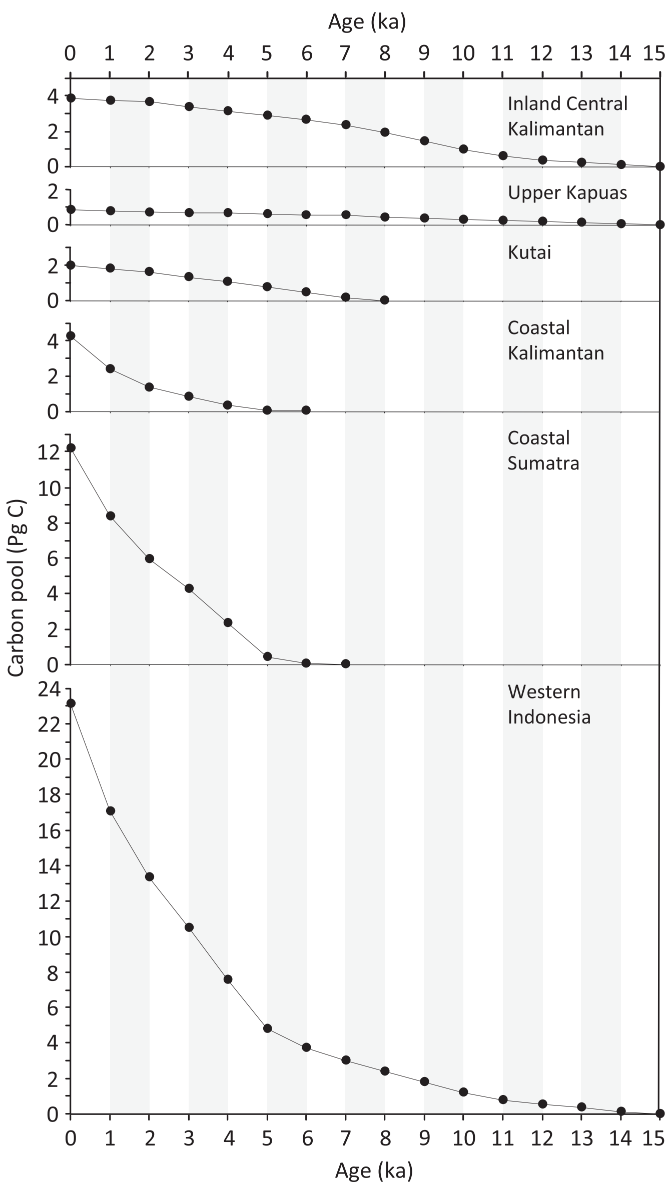


Fig. 4. Peatland carbon pools of western Indonesia over the past 15,000 years.

4.3.1. Upper Kapuas Basin

The rate of carbon storage across the entire Upper Kapuas basin remained below 0.1 Tg C yr^{-1} over the entire study period. The maximum value of 0.09 Tg C yr^{-1} at 0 ka represented only 1% of the contemporaneous rate of carbon storage of all western Indonesian peatlands. Since 2 ka carbon was also released in some of the Upper Kapuas peatlands through aerobic peat decay from an estimated 610 km^2 large area at a rate of 0.01 Tg C yr^{-1} (Fig. 2). The low regional carbon sink combined with Late Holocene carbon losses resulted in a carbon pool of only 0.84 Pg C. Representing 4% of the western Indonesian carbon pool the Upper Kapuas peatlands are the smallest among the regional carbon pools (Fig. 4, Table 1).

4.3.2. Inland Central Kalimantan

Inland Central Kalimantan was the dominant carbon sink in western Indonesia from 14 to 7 ka and contained the largest peat carbon pool at this time (Table 1, Fig. 4). The rate of annual carbon storage in this region was below 0.2 Tg C yr^{-1} prior to the Holocene and then increased to a Holocene maximum of 0.64 Tg C yr^{-1} at 9 ka or 90% of the contemporaneous peat carbon storage in western Indonesia. Regional carbon storage declined to remain between 0.3 and 0.2 Tg C yr^{-1} from 6 ka to 0 ka.

During the past two millennia carbon was released from peatlands of inland Central Kalimantan by aerobic peat decomposition over an estimated 4320 km^2 at a rate of 0.14 Tg C yr^{-1} (Table 1, Fig. 2). The growth of the regional carbon pool conforms to a sigmoid curve. The total mass of 3.8 Pg C by 0 ka makes up 16% of the total western Indonesian peat carbon pool (Table 1, Fig. 4).

4.3.3. Kutai Basin

In the Kutai basin peat carbon storage began at 8 ka with a rate of 0.06 Tg C yr^{-1} and increased to a Holocene maximum of 0.4 Tg C yr^{-1} at 6 ka when this region dominated peat carbon storage in western Indonesia (47% of total storage). The rate subsequently decreased to 0.25 Tg C yr^{-1} by 0 ka, representing only 3% of the total carbon storage in western Indonesia. Carbon release from fires after 5 ka accounted for annualized losses between 0.003 and 0.01 Tg C yr^{-1} (Table 1). Approximately 10% or 440 km^2 of the Kutai peatlands were affected by fire-related carbon losses and truncation of their peat profile prior to modern land use. However, over the last 8000 years a peatland carbon pool of 2 Pg C developed in the Kutai basin, representing 9% of the entire western Indonesian pool (Table 1, Fig. 4).

4.3.4. Coastal peatlands of Kalimantan and Sumatra

Currently, coastal peatlands of Sumatra and Kalimantan together store 16.5 Pg C or 71% of the total peatland carbon pool of the study region. Prior to anthropogenic degradation coastal peatlands also dominated the western Indonesian peat carbon sink with 6.6 Tg C yr^{-1} or 92%.

The regional rate of carbon storage in coastal Kalimantan was still insignificant at 6 and 5 ka, but at 4 ka it was already comparable to the carbon storage of all inland regions of Kalimantan. Carbon storage rose to a final, maximum rate of 2.2 Tg C yr^{-1} at 0 ka (30% of the overall carbon sink; Table 1, Fig. 2). Associated with this rising rate in annual carbon storage is the rapid growth of the region's carbon pool to a total of 4.3 Pg C by 0 ka (Table 1, Fig. 4). This reservoir represents 18.6% of the total peat carbon pool making it the second largest (Fig. 4).

In Sumatra the rate of carbon storage became significant by 5 ka with a value of 0.9 Tg C yr^{-1} surpassing the collective carbon storage of all other regions (Fig. 2). This rate increased to 2.4 Tg C yr^{-1} by 4 ka followed by lower rates of around 2 Tg C yr^{-1} at 3 to 2 ka. Over the last two millennia the regional rate of carbon storage increased greatly to 4.5 Tg C yr^{-1} at 0 ka (Table 1). Until recently, Sumatra's peatlands were the most important regional carbon sink contributing 62% of annual carbon storage in the peatlands of western Indonesia. The carbon pool of Sumatra reached a total of 12.2 Pg C by 0 ka, which accounts for 53% of the peat carbon pool of western Indonesia (Fig. 4).

5. Discussion

5.1. Uncertainties in the reconstruction of peatland area changes

Our reconstruction of peatland expansion reveals two important findings: first, the spatial expansion of inland peatlands in Indonesian Borneo was substantial during the last deglaciation between 20 and 11.7 ka when about 20% of the inland peatlands in Kalimantan formed. This expansion and the initiation of new peatlands continued throughout the Holocene. Second, the expansion of coastal peatlands was greatest during the past two millennia. This latter finding supersedes the earlier assessment of Dommain et al. (2011) that was based on a compilation of oldest basal dates and suggested a maximum in coastal peatland initiation and expansion between 7 and 4 ka. The present study emphasizes the importance of young, shallow peat deposits, which are rarely dated, as radiocarbon dating tends to focus on deeper deposits that cover longer time intervals and store more paleo-data.

Although basal dates may provide a reasonable proxy for peatland initiation and also to some extent local peatland expansion, they are not well suited for estimating the total regional extent of peatlands over time. In order to assess the potential error we compare our reconstruction of peatland expansion with the more common approach of using the cumulative number of oldest basal dates (Fig. 5). The cumulative number of oldest basal dates indicates that peatland expansion would have occurred much earlier and that the peatland area was substantially larger throughout the Holocene. For example, at 6 ka basal dates suggest that almost 50% of the total current peatland area was present, whereas our approach indicates only 13% at 6 ka and 50% not until about 3500 years later (Fig. 5).

Our reconstruction for inland Central Kalimantan places the onset of peatland formation mainly after 15 ka with rapid expansion during the Early Holocene. Some of the inferred old ages (18.5 ka–17 ka) of inland Central Kalimantan sites could be overestimates because the assumed (mean) peat accumulation rate of 0.54 mm yr^{-1} seems too low for very deep peat deposits (>1000 cm). However, the total area of this older peat is very small (~2000 km^2). Peatland initiation in Kutai took place after 8.3 ka in agreement with the oldest basal dates from this region, but coastal peatlands did not form before 8 ka, although they afterwards continuously expanded to the present, consistent with available basal radiocarbon dates.

The peatland atlases identify 36,700 km^2 of shallow coastal peat (0–1 m), which represents more than 25% of the total peatland area of western Indonesia. We assigned young basal ages (<2 ka) to these shallow coastal peatlands based on the high rate of peat accumulation derived for coastal settings. Available radiocarbon dates support this assumption: five sites along a 300 km long stretch of the Sumatran coastal plain (up to 30 km from the current

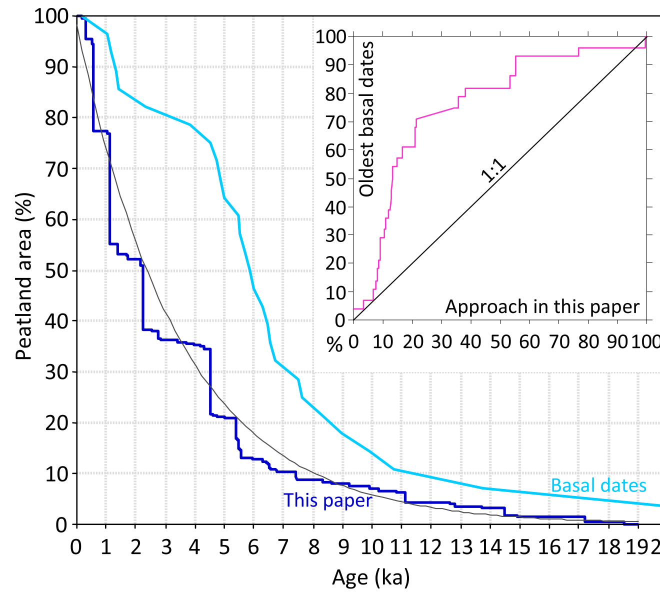
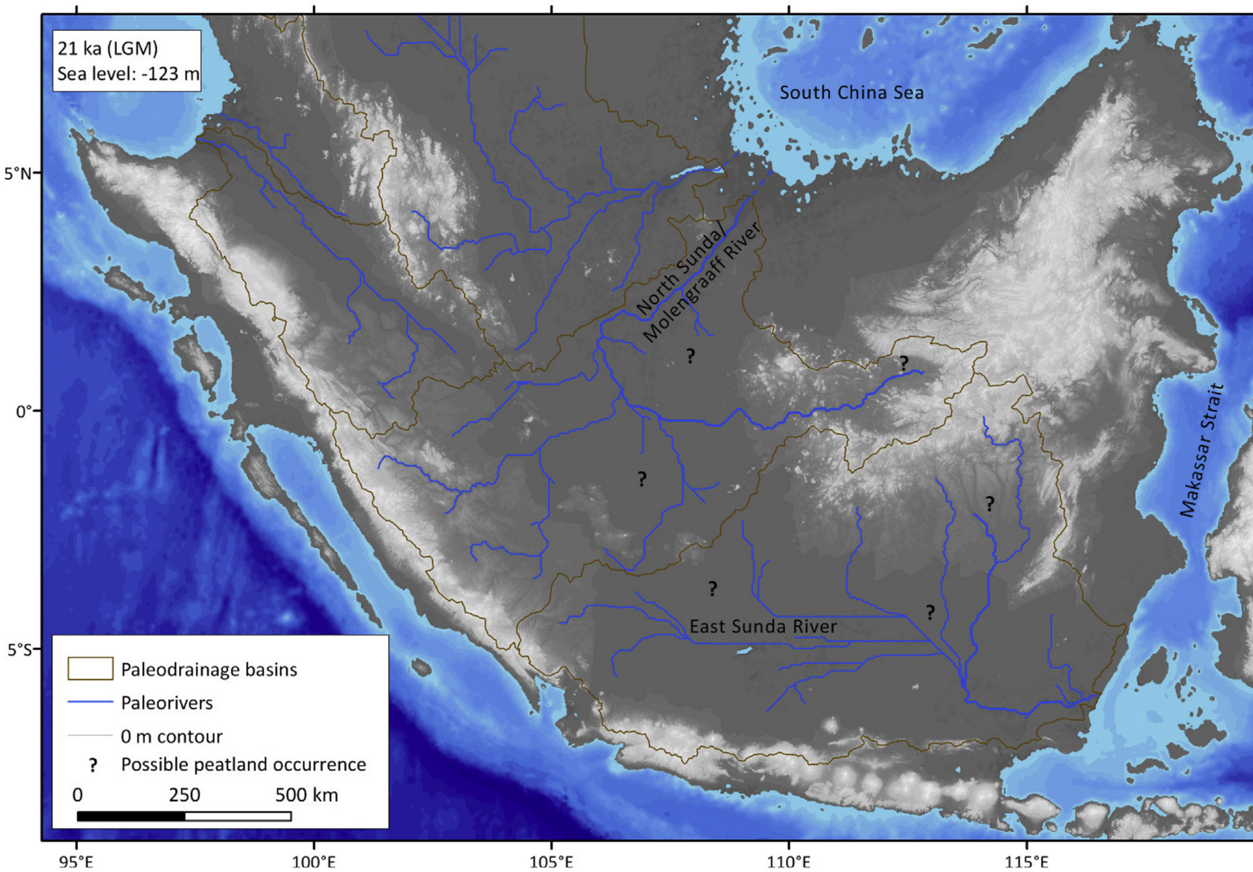


Fig. 5. Comparison of two methods for reconstructing peatland area: the dark blue line (fitted with an exponential curve) represents reconstructed increase in peatland area using the transfer function method of this paper, the light blue line using the cumulative number of oldest basal dates. Peatland area increases much earlier in the basal dates approach. The difference between both methods is shown in the inset graph.

coastline) have calibrated basal radiocarbon dates younger than 2 ka as have two sites in coastal Kalimantan (8–50 km from the coast; Fig. 6). Similarly young peatlands are also found in the coastal areas of northwest Borneo (Wilford, 1960; Tie and Esterle, 1992; Staub and Esterle, 1994; R. Dommain, unpublished data). The seven youngest dated sites, moreover, show fast rates of vertical peat growth. Their mean peat accumulation rate of 2.2 mm yr^{-1} (1.3–3 mm yr^{-1}) is higher than the average value of 1.77 mm yr^{-1} applied to derive basal dates for undated coastal polygons. We thus conclude that shallow coastal peatlands most likely developed rapidly and are unlikely to be of (much) earlier origin.

This conclusion is supported by the 27,000 km^2 of Sumatra and 19,000 km^2 of Kalimantan that are presently located at an elevation that is only ≤5 m a.s.l. according to SRTM elevation data. This land must have been submerged during the highstand in sea level between 5 and 4 ka (Fig. 6). At least 33% (9000 km^2) of this low-lying land is currently covered with peat in Sumatra and 26% (~5000 km^2) in Kalimantan. The total area of submerged land was probably even larger because 1) the SRTM elevation data do not represent the actual ground surface in vegetated terrain (i.e. peat swamp forest; Hofton et al., 2006; Jaenicke et al., 2008) and 2) vertical peat growth has in the meantime raised many peatland surfaces more than five metres above present sea level. Their topographic situation thus implies that a large fraction of the coastal peatlands could only have formed after sea level had fallen and made these extensive coastal areas available for plant colonization.

The continued spread of peatlands is not only related to the initiation of new peatlands, but also to the lateral expansion of existing (domed) peatlands (e.g. Korhola et al., 2010). Fig. 6 shows lateral expansion of interfluvial peat domes in inland Central Kalimantan and of the large peatlands of central eastern Sumatra between 2 and 1 ka. Radiocarbon evidence for continuous lateral peatland expansion since 2 ka exists for the Batang Hari peatland in Sumatra (Supiandi, 1988) and for Teluk Keramat and Sebangau in Kalimantan (Neuzil, 1997; Page et al., 1999, Fig. 6). These multiple



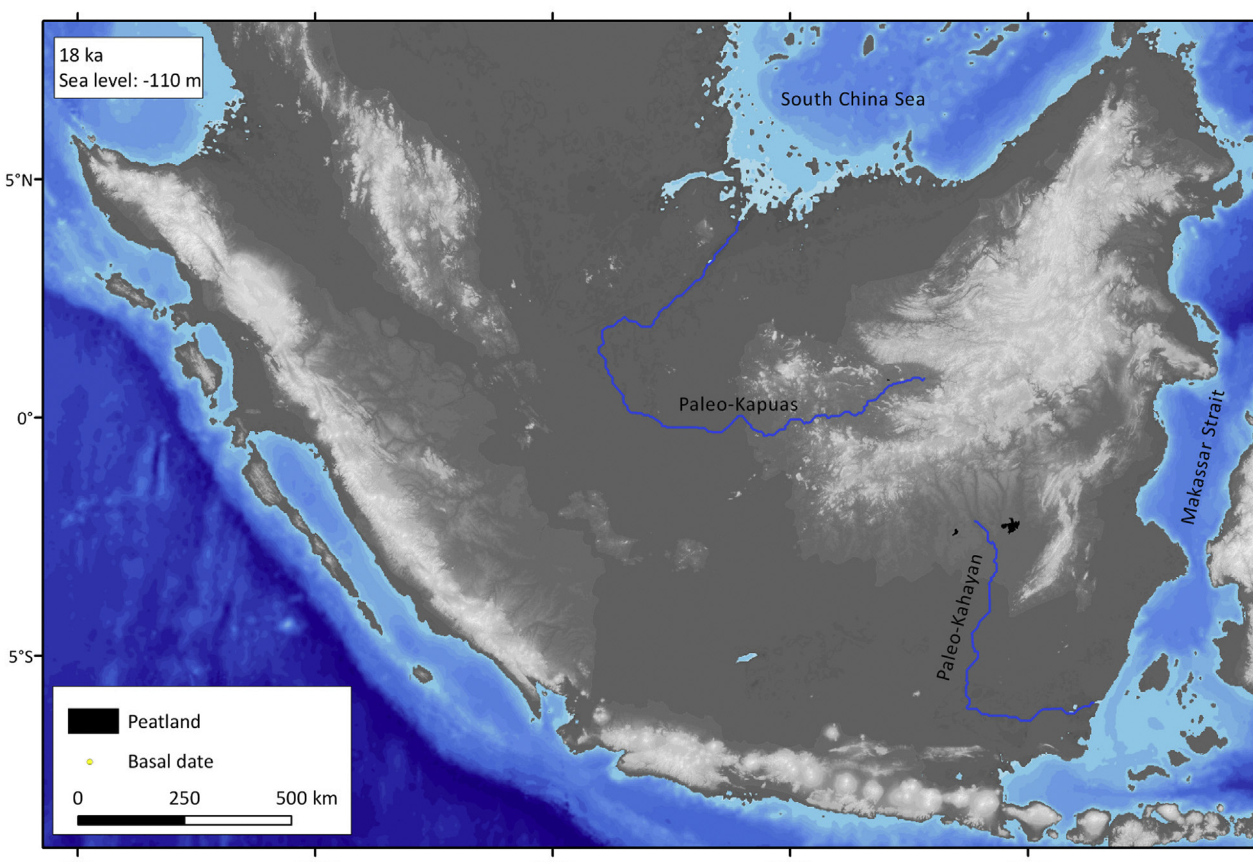
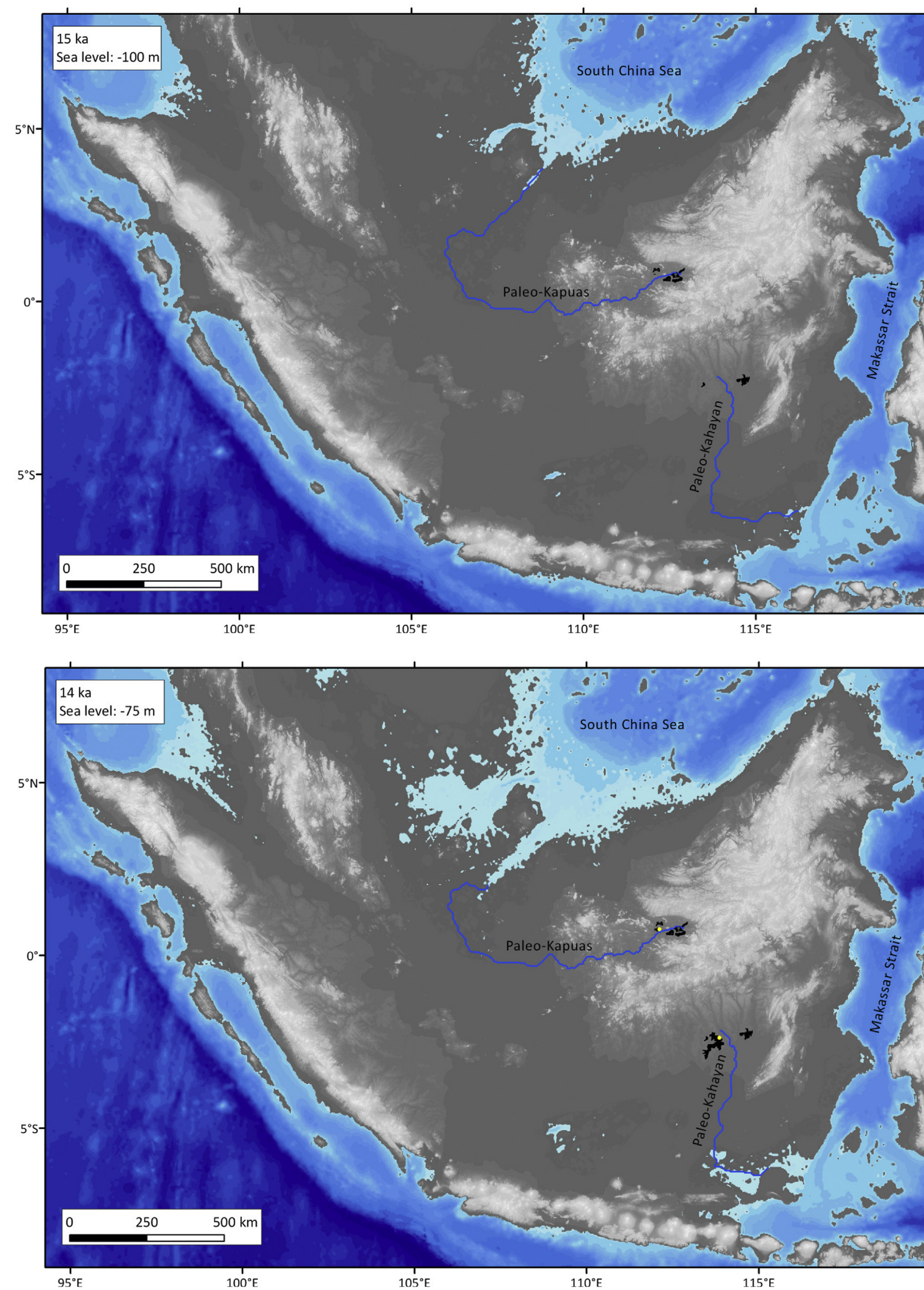


Fig. 6. Paleogeographic maps of peatland expansion (black areas) in western Indonesia, sea-level changes over the Sunda Shelf, and changes in the paleo-drainage network since 21 ka. Yellow dots indicate location of basal peat dates available for the respective millennia. Sea-level changes are based on the sea-level curve of Hanebuth et al. (2011), shown in Fig. 7. Elevation profiles of the Paleo-Kahayan-East Sunda River and the Paleo-Kapuas-North Sunda (Molengraaff) River are shown in Fig. 8.



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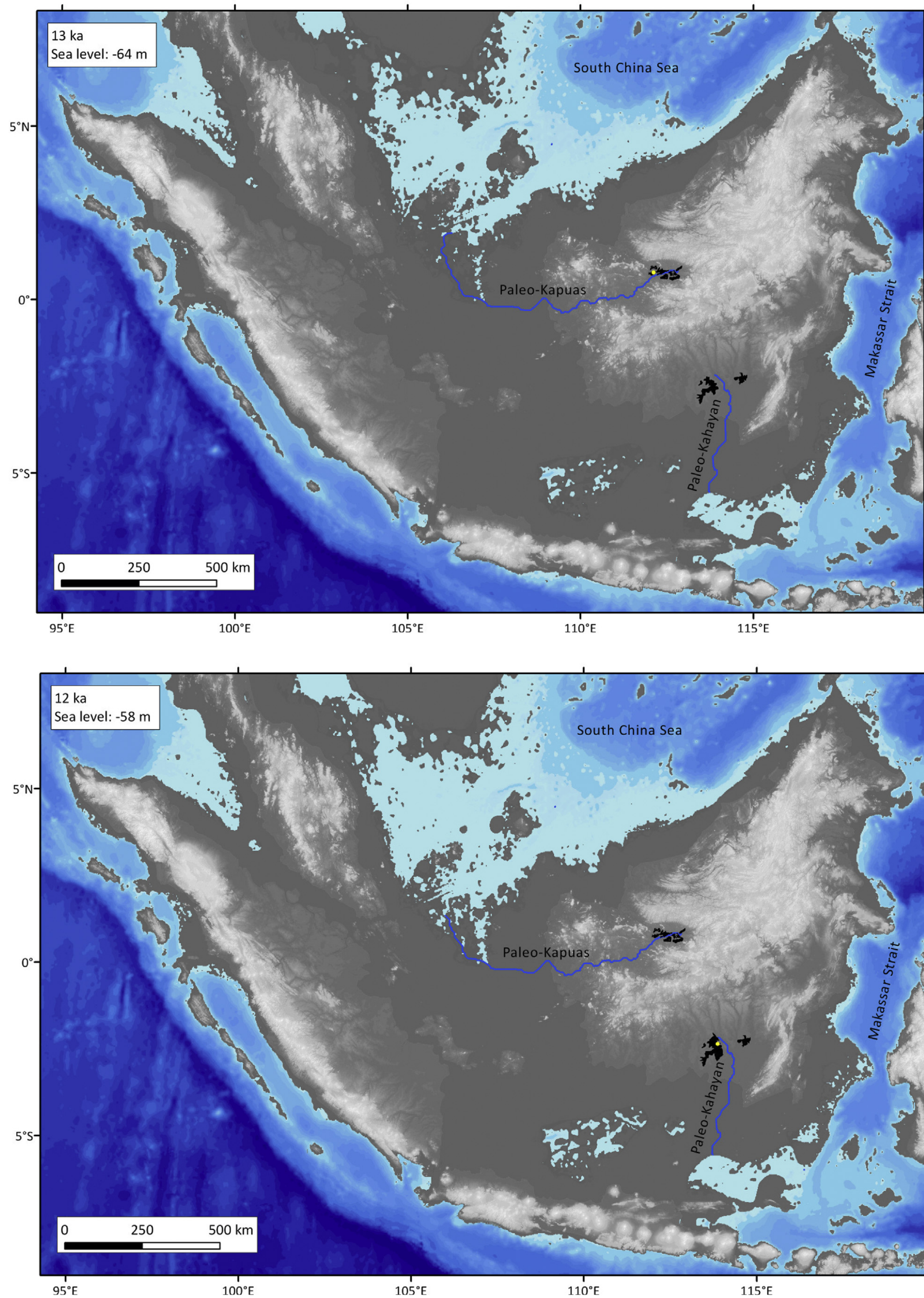

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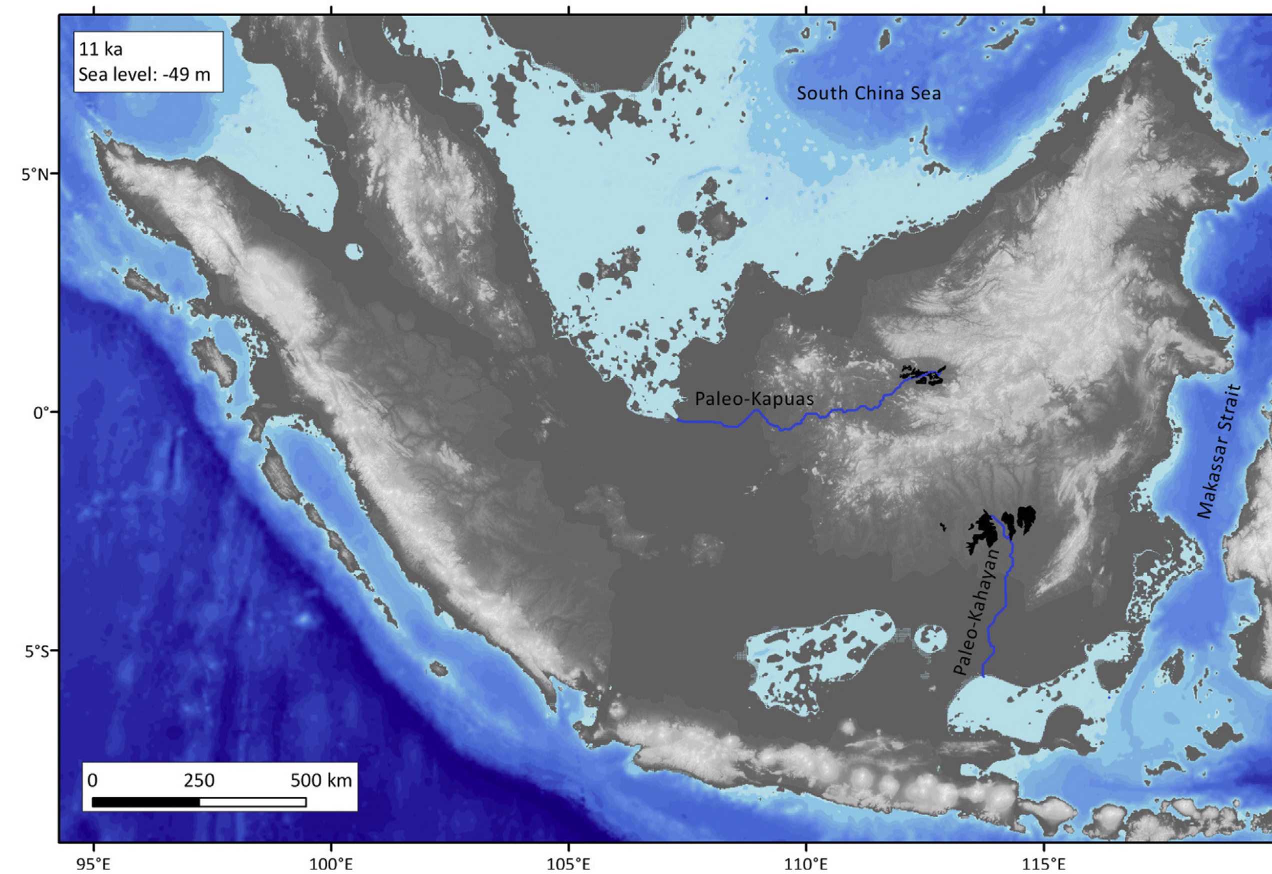

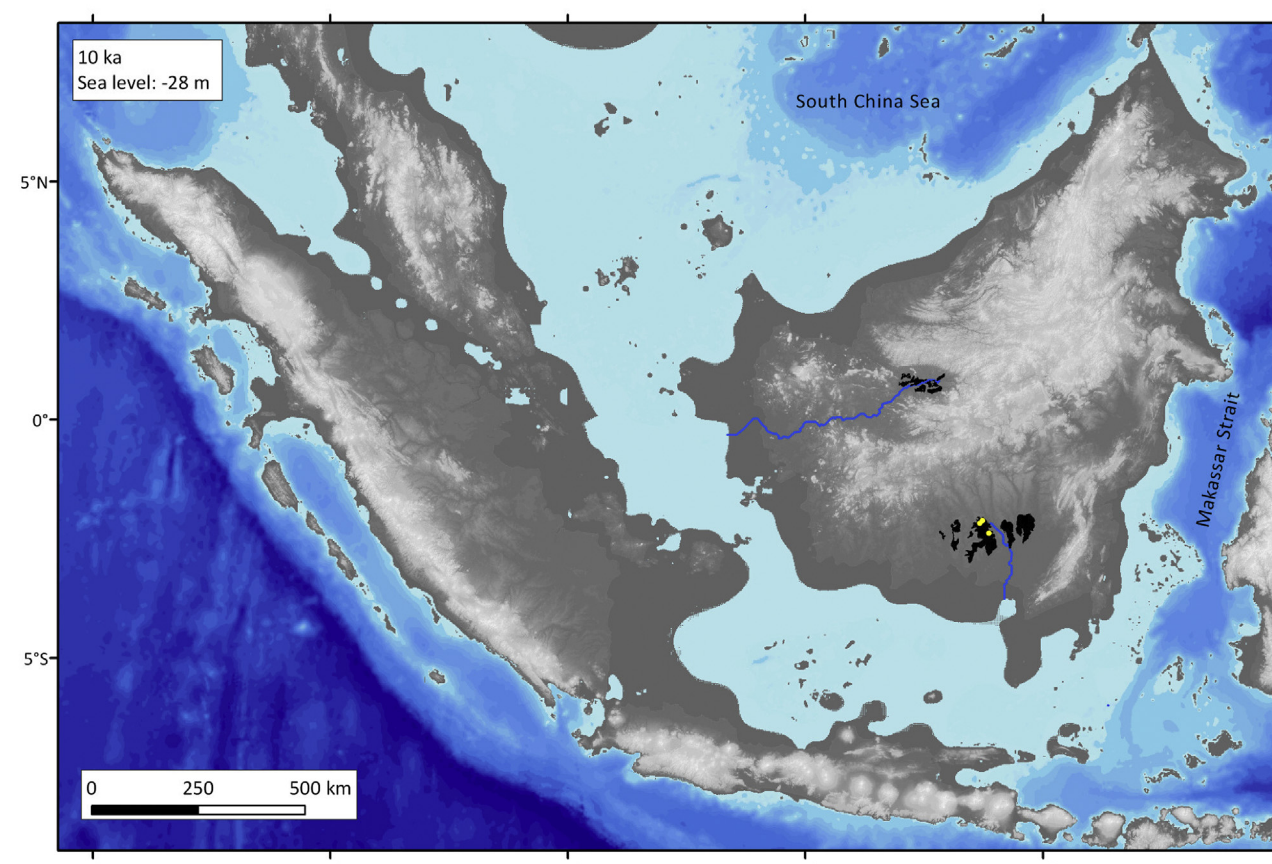

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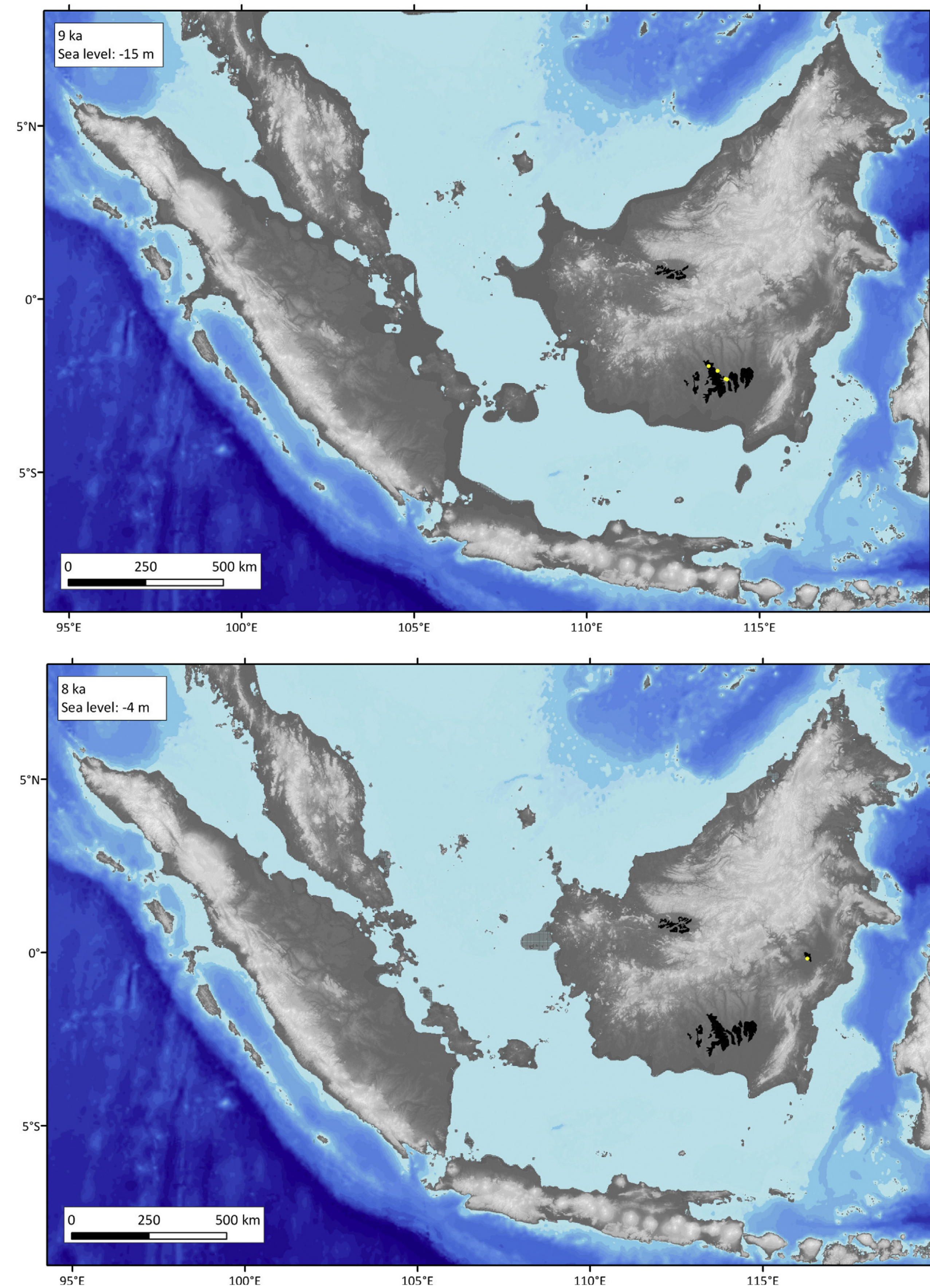

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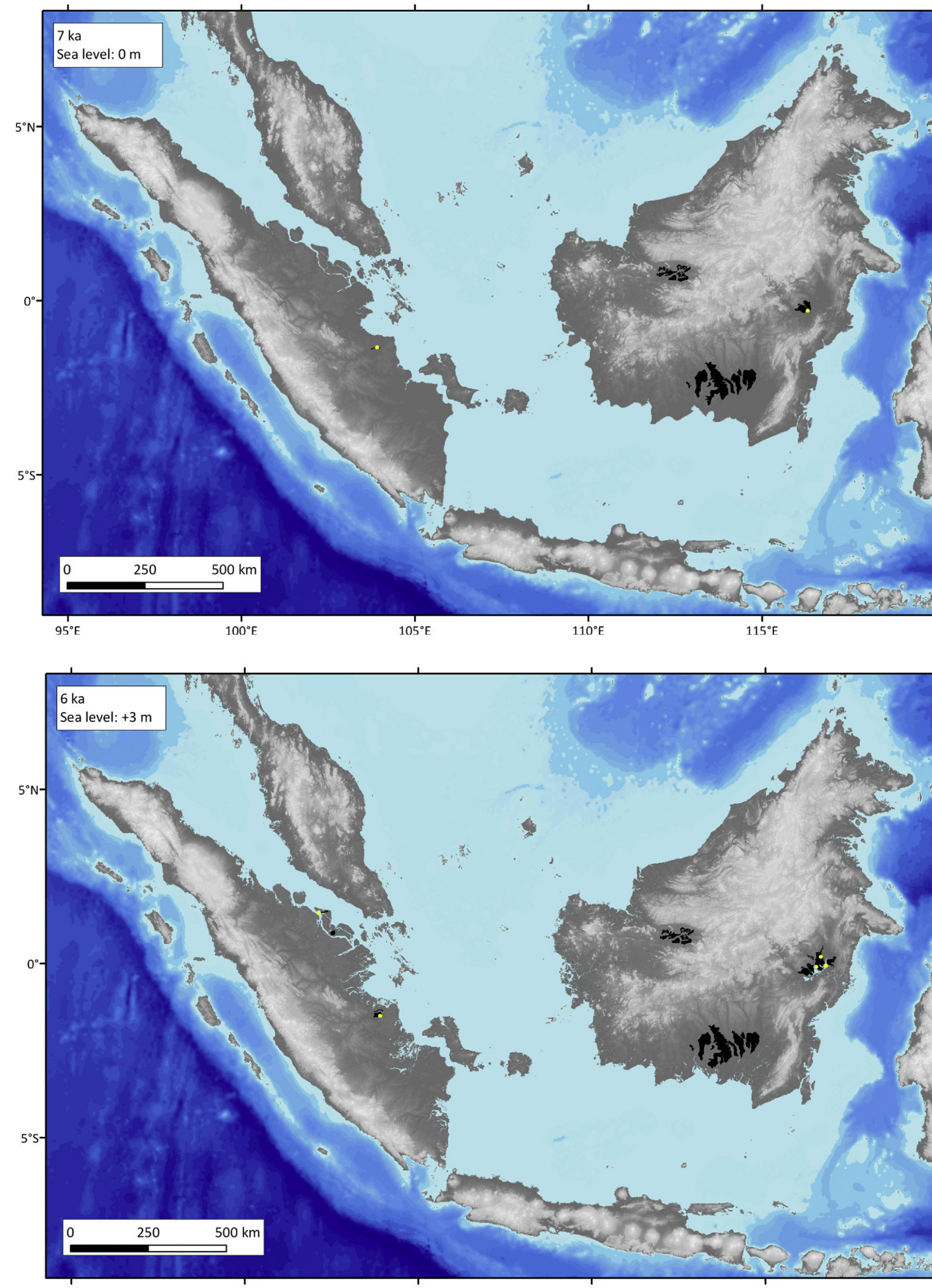

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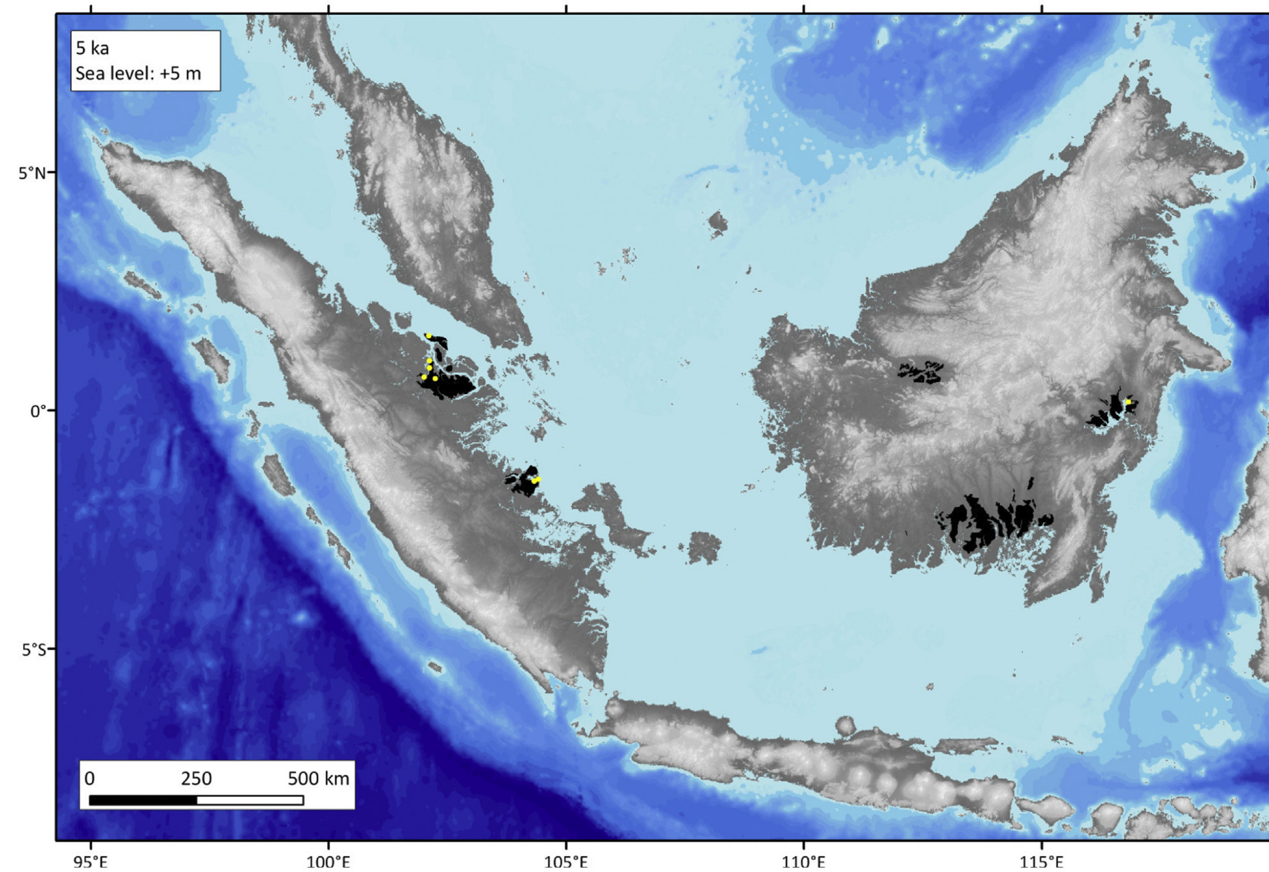

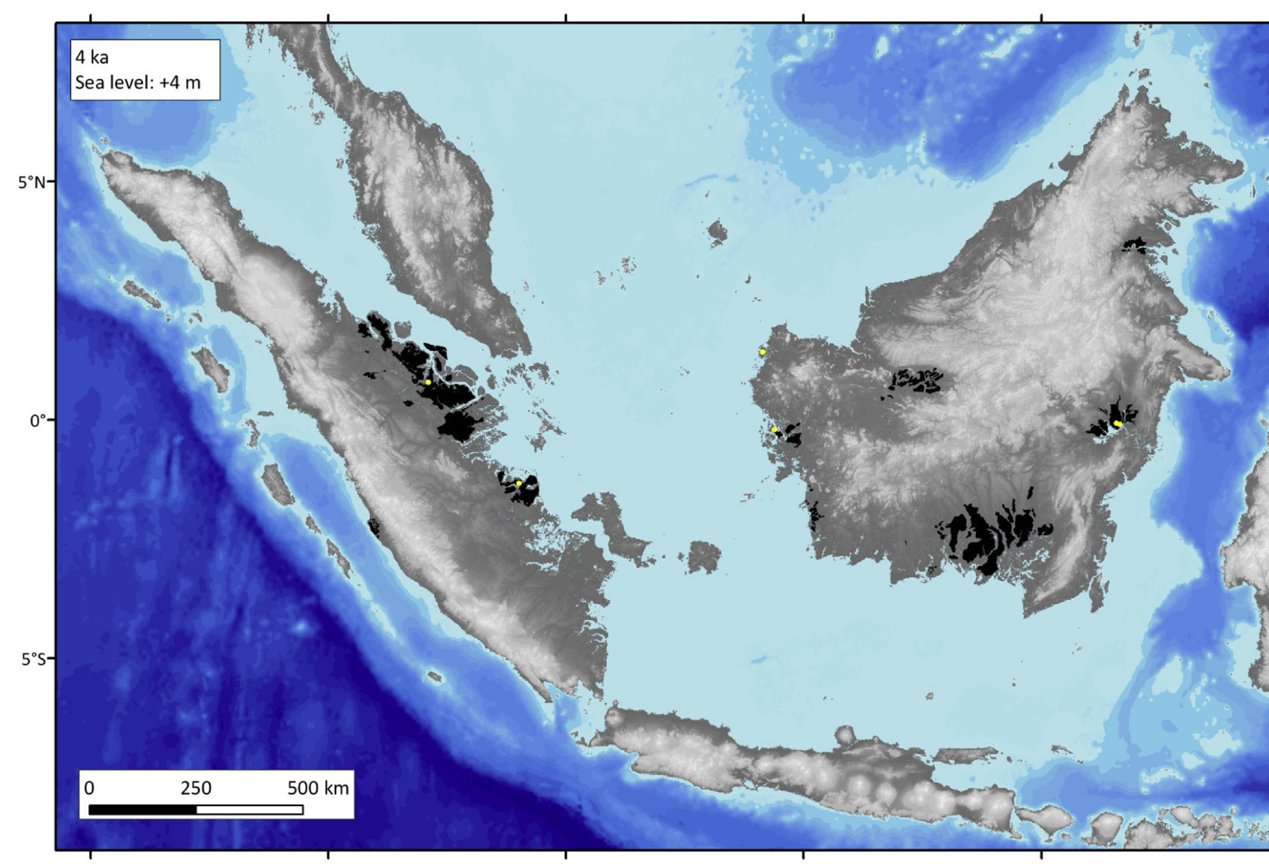

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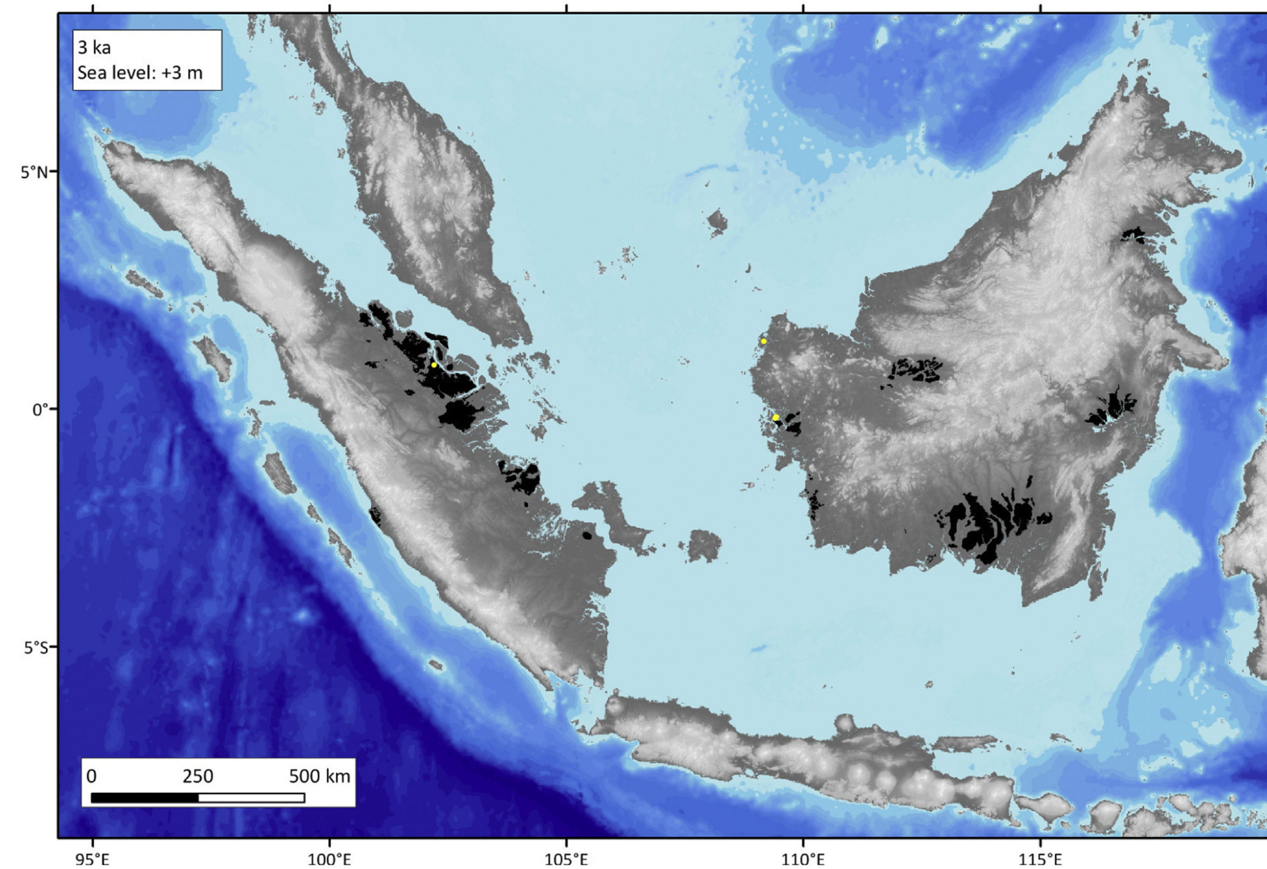

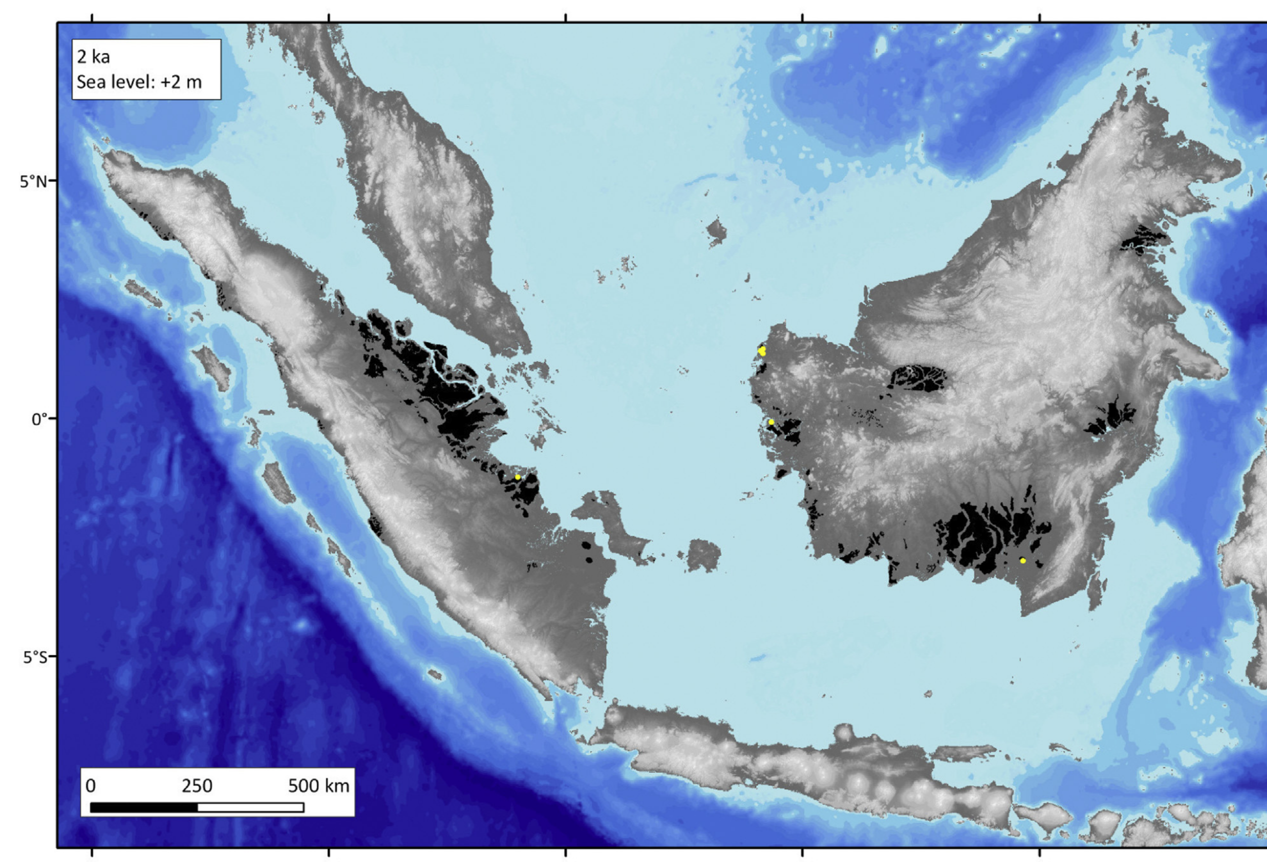

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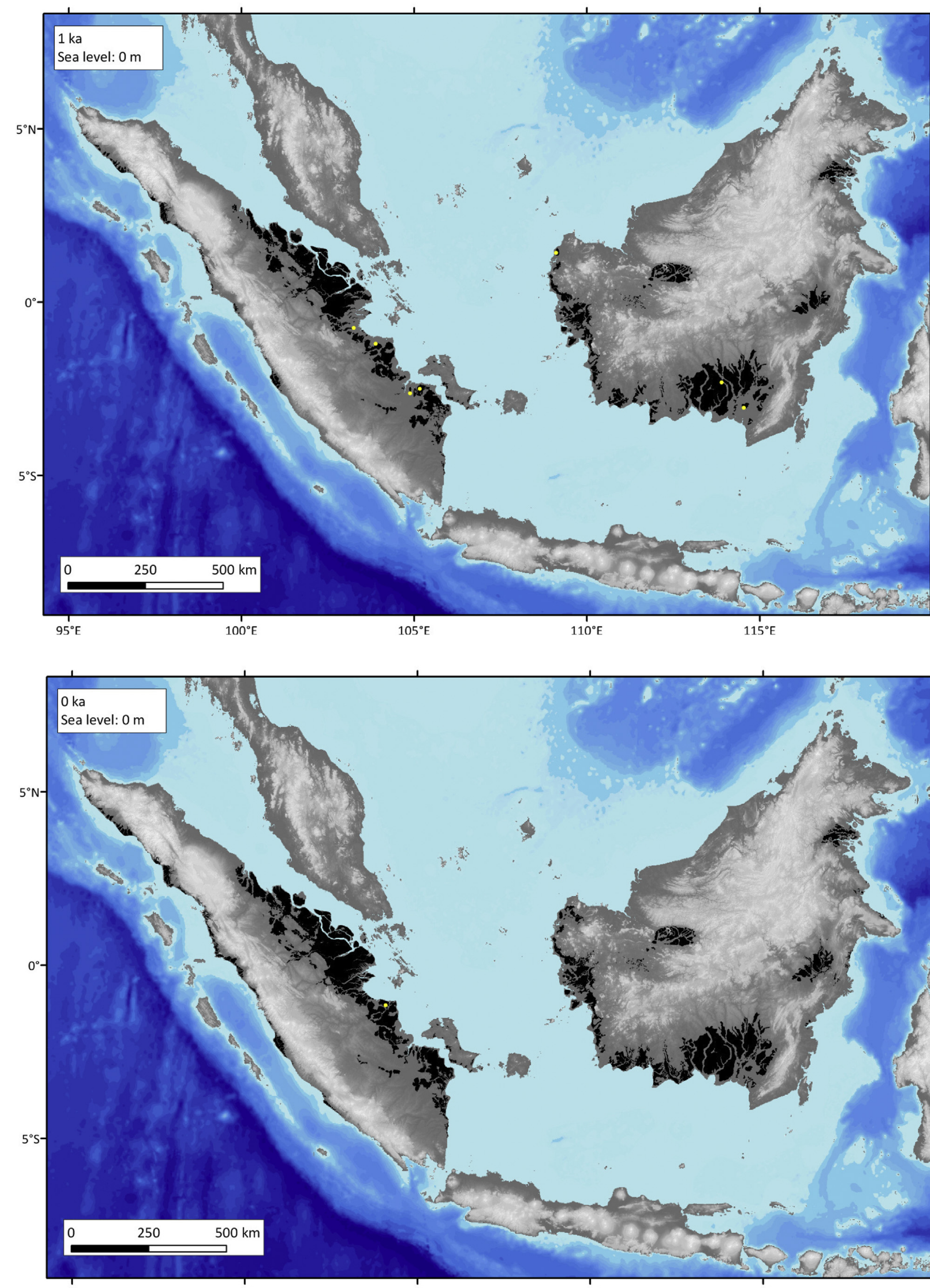

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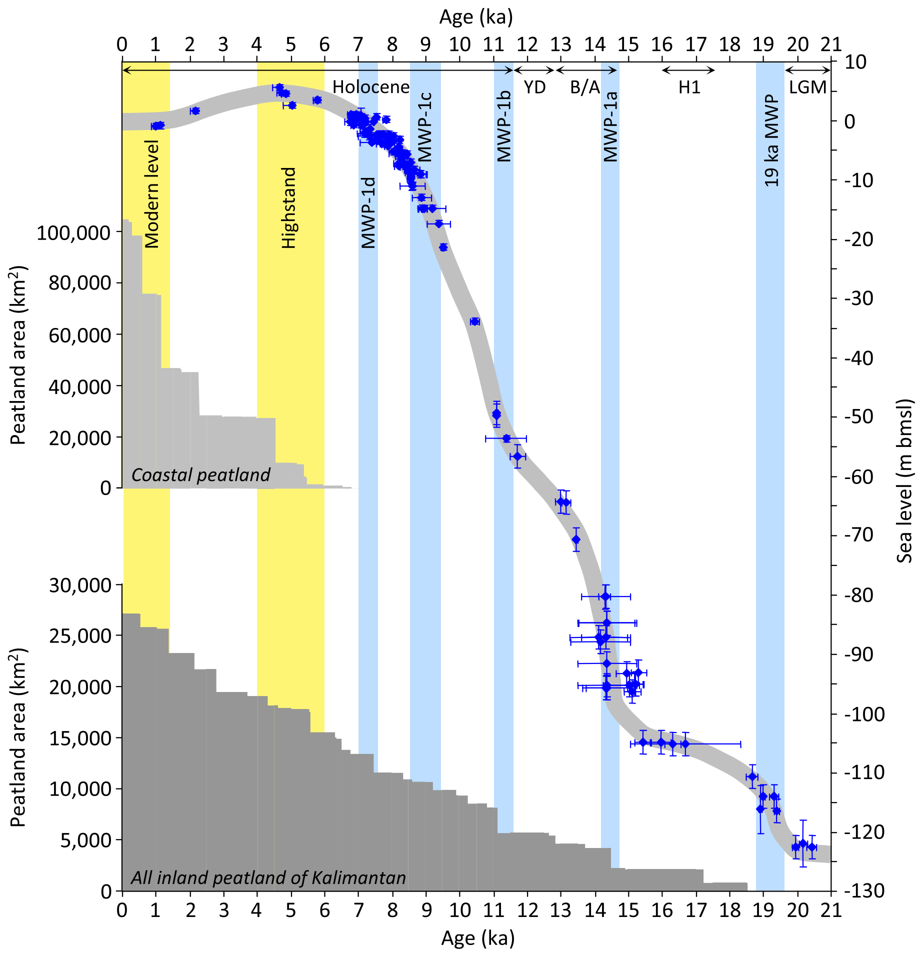


Fig. 7. Sea-level curve of the Sunda Shelf with available radiocarbon dates from Hanebuth et al. (2011) for the past 21,000 years and area of coastal (top) and inland (bottom) peatlands. Note the difference in scale for the two regions. MWP denotes meltwater pulse, LGM Last Glacial Maximum, H1 Heinrich event 1, B/A Bølling/Allerød, and YD Younger Dryas.

lines of evidence support our reconstruction of widespread, recent peat formation in the lowlands of western Indonesia.

5.2. Peatland formation and expansion: the role of sea-level change

During the Last Glacial Maximum (LGM, 26.5–19 ka) sea level was 123 m lower than today and the Sunda Shelf was fully exposed, connecting the islands of Borneo, Sumatra, Java and Bali with mainland Southeast Asia (Hanebuth et al., 2009, Fig. 6). Four major river systems drained the shelf platform as extensions of modern day rivers (Molengraaff and Weber, 1921; Tjia, 1980; Voris, 2000, Fig. 6). Bornean peatlands that existed during Marine Isotope Stage 3 (60–26.5 ka), such as in the Upper Kapuas basin (Anshari et al., 2001, 2004), likely degraded during the LGM when exposure of the Sunda Shelf led to substantially reduced moisture convection and resulting dry conditions in the Indo-Pacific Warm Pool (Bird et al., 2005; Partin et al., 2007; Wurster et al., 2010; Carolin et al., 2013; DiNezio and Tierney, 2013). In addition, the sea-level lowstand steepened the regional hydraulic gradient of land masses, draining the upland areas more effectively through so-called topography-driven groundwater flow (Post et al., 2013) and thereby lowering the water-tables in contemporaneous peatlands. It has been suggested that the exposed shelf with its low gradient could have supported large areas of peatland itself (Kaplan, 2002; Slik et al., 2011). However, the deeply incised river valleys (up to 40 m deep) on the shelf floor (Tjia, 1980; Hanebuth and Stattegger, 2003) with their low hydraulic head would have effectively drained the widely distributed coarse-grained sediments (Bird et al., 2005) of the adjacent watersheds (cf. Post et al., 2013). It therefore seems unlikely that the prevailing dry climate and hydrogeologic setting would have favoured widespread peat accumulation on the Sunda Shelf (Kennett et al., 2003). This inference is supported by the absence of thick LGM freshwater peats on the shelf floor (Geyh et al., 1979; Hanebuth et al., 2011; T. Hanebuth, pers. com.).

The post LGM sea-level history of the Sunda Shelf was characterized by three important developments: 1) rapid sea-level rise during deglaciation from ~19 to 7 ka with associated flooding of the shelf by ~9.5 ka, followed by 2) a slow rise to a Holocene highstand at about 5 ka and 3) the subsequent lowering of sea level by about 5 m (Hanebuth et al., 2000, 2011, Figs. 6 and 7). The deglacial sea-level transgression was punctuated by periods of abrupt rises in sea level associated with so-called meltwater pulses (MWP), the most prominent being MWP 1a from 14.6 to 14.3 ka when sea level rose by 16 m at a rate of approximately 50 mm yr^{-1} (Hanebuth et al., 2000, Fig. 7).

Our reconstruction shows that there was first, very localized, peat formation in inland areas of Borneo at the onset of the deglaciation at 19.9–18.5 ka. At this time sea level rose to about 110 m below modern sea level (mbsl) during its initial pulse-like rise (19 ka MWP; Clark et al., 2004; Hanebuth et al., 2009, Fig. 7). During subsequent Heinrich event 1 (17.5–16 ka), which was the driest period of the past 100,000 years in Borneo (Partin et al., 2007; Carolin et al., 2013), peat accumulation and peatland expansion remained very limited while sea level rose relatively slowly by 4.1 mm yr^{-1} (Hanebuth et al., 2000).

The inland peatland area increased to >4000 km^2 by 14.5 ka at the time of MWP 1a, which raised sea level to about 80 mbsl with an associated lateral transgression of over 450 m yr^{-1} (Hanebuth and Stattegger, 2003). This event flooded the outlet of the East Sunda River into the Makassar Strait as well as large parts of the North Sunda (Molengraaff) River valley by the expanding South China Sea (Fig. 6). These rivers acted as the main drainage ways for the watersheds of the existing peatlands in the Upper Kapuas and Central Kalimantan areas, respectively (Figs. 6 and 8). The deglacial rise in sea level led to a landward migration of the river mouths and reduced the hydraulic gradient within the inland watersheds since sea level represents the ultimate base level for the regional drainage network (Kafri and Yechieli, 2012).

In this way MWP 1a resulted in impeded drainage of the shrinking land masses and in a regional rise in the water table. In addition, the associated flooding of the shelf and the synchronous onset of the Bølling/Allerød (B/A) warm interval led to an abrupt increase in the atmospheric supply of moisture (Kienast et al., 2003; Partin et al., 2007; Tierney et al., 2012). Together, the regional decrease in river discharge and the increase in net recharge (i.e. precipitation minus evapotranspiration) were the principal drivers for the rise of groundwater mounds within the interfluvial divides and the formation of inland domed peatlands (cf. Glaser et al., 2004).

An analysis of the changing hydraulic gradient for the Upper Kapuas (i.e. Paleo-Kapuas River) and Central Kalimantan (i.e. Paleo-Kahayan River) drainage basins is shown in Fig. 8. Thresholds relevant for peatland initiation in these two regions seem to be rises in sea level that surpassed knickpoints on the shelf floor and consequently caused significant landward inundation of the river

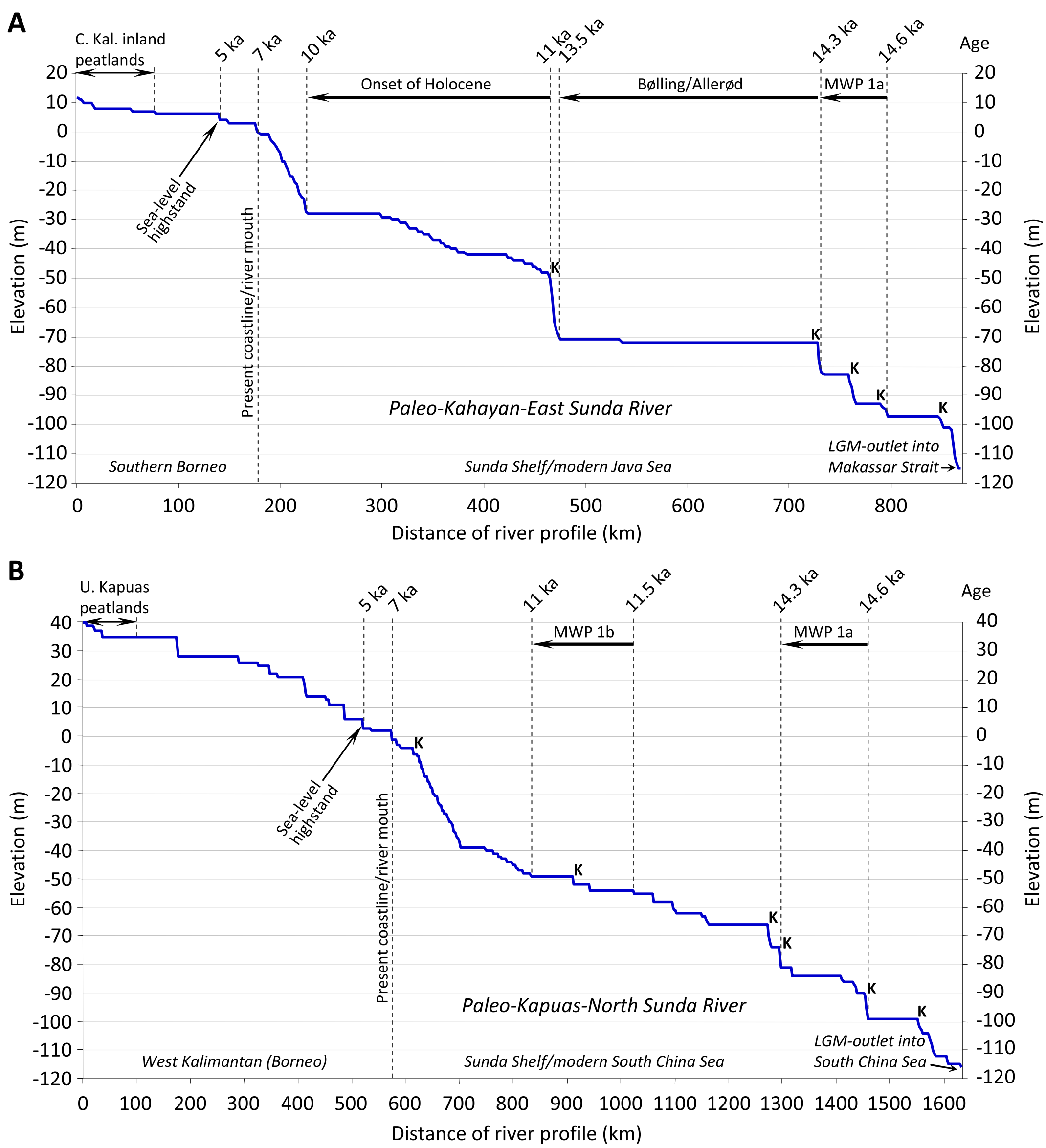


Fig. 8. Elevation profiles for (A) the Paleo-Kahayan-East Sunda River and (B) the Paleo-Kapuas-North Sunda (Molengraaff) River running from the current peatland areas (left) of inland Central Kalimantan (A) and Upper Kapuas (B) down-gradient across the Sunda Shelf. Profiles were extracted from the ETOPO 1 relief model. Dashed vertical lines mark the position of the coastline at specific times discussed in the text. Arrows denote periods of sea-level rise that led to rapid lateral transgression. Note the difference in horizontal scale. K denotes knickpoint. The location of these river systems is shown in Fig. 6.

valleys. The Kapuas-North Sunda river system provides an example of such a threshold at MWP 1a, when the sea level rose to 80 mbsl shifting the river mouth landward by over 170 km while the mouth of the Kahayan-East Sunda River retreated by only 65 km (Fig. 8). Independent evidence for a significant retrogression of river mouths during MWP 1a comes from a biomarker record of the South China Sea that indicates an abrupt reduction in the discharge of terrestrial organic matter into the ocean at this time (Kienast et al., 2003). Another transgression threshold for the Kahayan-East Sunda River system was surpassed later during the B/A interstadial when the sea level reached 70 mbsl (Fig. 8). This rise abruptly flooded the river valley over a 250 km reach and brought the coast within about 320 km of the peatland area of Central Kalimantan.

The expansion of peatlands slowed during the Younger Dryas period (12.9–11.7 ka) when the rate of rising sea level also decreased to about 6 mm yr^{-1} (Hanebuth et al., 2011). This interval was followed by a substantial increase in peatland initiation and expansion in inland Central Kalimantan at the beginning of the Holocene (Fig. 7). By 11 ka the peatland area in this region had almost doubled in size to an estimated 6300 km^2. This areal expansion directly followed a period of accelerated sea-level rise of about 11 mm yr^{-1} between 11.5 and 11 ka (originally termed MWP 1b; Fairbanks, 1989; but see Bard et al., 2010). Whereas no large retrogression in the Kahayan-East Sunda river system is discernable after this sea-level rise to about 49 mbsl, there must have been an associated landward submergence of the Kapuas-North Sunda River valley over a distance of 190 km (Fig. 8). Although basal radiocarbon dates from the Upper Kapuas are absent for this period, it seems very likely that peatlands originated here at around 11 ka.

Sea level continued to rise rapidly by approximately 20 mm yr^{-1} between 11 and 10 ka, which led to the final separation of Borneo from Sumatra and Java at a sea level of ca 30 mbsl (Hanebuth et al., 2011, Fig. 6). This transgression drowned the Kahayan-East Sunda River valley over a substantial distance of 230 km, further decreasing the hydraulic gradient in its catchment (Fig. 8) and promoting the initiation and expansion of peatlands in inland Central Kalimantan (Figs. 6 and 7).

Between about 9.5 and 8.5 ka another pulse-like rise in sealevel is recorded in Southeast Asian seas as MWP 1c (Hori and Saito, 2007; Tjallingii et al., 2010). During MWP 1c the sea level over the Sunda Shelf rose to 8 mbsl at a rate of 10–25 mm yr^{-1} (Fig. 7). Immediately afterwards the first peatland in the Kutai basin originated (8.3 ka; Hope et al., 2005, Fig. 6) presumably as a result of terrestrialisation after the initial inundation of this shallow inland basin (Dommain et al., 2011). A final period of rapid sea-level rise, with a rate of 10 mm yr^{-1}, started at around 7.4 ka (MWP 1d; Liu et al., 2004; Bird et al., 2010, Fig. 7). Concurrent with MWP 1d was the expansion of existing peatlands in Central Kalimantan and Kutai while new peatlands also formed.

Periods of rapid sea-level rise of $\geq$10 mm yr^{-1} (i.e. meltwater pulses) evidently triggered peatland formation in inland Borneo during the last deglaciation (Fig. 7). Each time sea level reached a critical threshold elevation, large portions of the Sunda Shelf were laterally flooded which lowered the hydraulic gradient while the rising base level also reduced the discharge of both ground and surface water from the land masses. These processes produced a rise in the water table, resulting in paludification of the relatively flat hinterland.

By 7 ka the sea level had nearly reached its present modern level and the first coastal peatlands on Sumatra formed (Supiandi, 1988; Dommain et al., 2011, Figs. 6 and 7). Prior to this time the rapid transgression of the Sunda Shelf persistently prevented the development of coastal peatlands. Only when the rate of rising sealevel slowed down to a threshold of 2.4 mm yr^{-1} and continued to decline could coastal peat accumulation keep up with the sea-level rise (Fig. 7).

Between 7 and 5 ka the Kutai peatlands strongly expanded as sea level rose to a highstand of about +5 m above modern sea level (Horton et al., 2005; Hanebuth et al., 2011; Fig. 7). Moreover, by about 5.4 ka the stabilization of sea level initiated the first rapid expansion of coastal peatlands across large areas of central Sumatra at locations such as Siak Kanan, Bengkalis Island, and Batang Hari-Berbak (Neuzil, 1997; Dommain et al., 2011, Fig. 6). These peatlands generally formed a few metres above modern sea level while current land surfaces at lower elevation were still inundated (Diemont and Supardi, 1987b; Supiandi, 1988; Supardi et al., 1993). Around 5 ka extensive areas of the eastern coast of Sumatra were submerged, particularly the South Sumatra basin as well as large parts of the western and southern coast of Kalimantan (Fig. 6). Our reconstruction indicates that the inland peatlands of Central Kalimantan were partly bordered by the invading sea at this time (Fig. 6).

The Holocene sea-level highstand at 4.5 ka and its retreat over the following millennia initiated a rapid increase in the expansion of coastal peatlands. About 10,000 km^2 of peatland had formed in coastal central Sumatra by the time of the sea-level highstand (Figs. 6 and 7). In addition, the first peatlands on the west coast of Kalimantan also originated during this period in areas several metres above modern sea level (Neuzil et al., 1993; Neuzil, 1997). After 4 ka sea level began to fall at a rate of ca 1–2 mm yr^{-1}, and coastal peatlands responded to this regression by a massive expansion across the newly exposed land. For example, Diemont and Supardi (1987b) report that coastal peatlands in West Kalimantan and in Riau (Sumatra) formed directly on former sea beds as indicated by the presence of pyrite, marine shells and the absence of mangrove remains in the mineral sediments underneath these peatlands. The Berbak peatland in Sumatra directly formed over marine clays that can be found up to 45 km inland from the current coast (Diemont and van Reuler, 1984; pers. observation R. Dommain). Another strong seaward expansion of peatlands occurred around 2.3 ka when peatlands formed over an additional 17,000 km^2 along the coasts of west and southwest Kalimantan and east-central Sumatra (Fig. 6).

The marine regression also contributed to peatland expansion in inland areas. In inland Central Kalimantan the regression promoted a “second phase” of peat formation, starting after 2 ka (Sieffermann et al., 1996; Page et al., 1999) when existing peatlands expanded laterally into the adjacent, hitherto flooded low-lying river valleys (Fig. 6). These young inland peatlands in Central Kalimantan represent the so-called “basin peats” of Sieffermann et al. (1988). In the Upper Kapuas basin the peatland area increased substantially by 2.1 ka. This increase could partly be the result of falling water levels in Danau Sentarum and other floodplain lakes, which still cover large parts of this inland basin.

The present sea level in the Sunda region was reached by approximately 1 ka when the modern coastline configuration was established (Figs. 6 and 7), except for the continued buildup of deltaic and estuarine systems (e.g. Sumawinata, 1998). Radiocarbon evidence indicates that this new land surface was rapidly covered by peatland placing the origin of six coastal peatlands that formed over near-shore sediments between 2.3 and 0.2 ka (Dommain et al., 2011). These young peatlands are currently located between 25 and 60 km from the current coastline which together with their basal ages points to a rate of coastal advance of 18–35 m yr^{-1}, consistent with reported rates of 15–30 m yr^{-1} for the Sumatran coast (e.g. Sobur et al., 1978; Tjia, 1980; Diemont and van Reuler, 1984). These independent lines of evidence demonstrate that new land for coastal peatland expansion was continuously formed over the past two millennia, leading to a marked rise in the area of coastal peatland to 76,000 km^2 by $\sim$1 ka.

Coastal advancement thus allowed the explosive expansion of peatlands over 80,000 km^2 in western Indonesia during the Late Holocene. The role of falling sea level as an agent for creating new land surfaces available for the establishment of peatlands is comparable to that of ice sheet retreat and isostatic rebound for peatland initiation in boreal and arctic regions (e.g. Harden et al., 1992; Glaser et al., 2004; MacDonald et al., 2006).

5.3. Rates of long-term peatland carbon storage

The carbon-balance of a peatland is the result of peat addition at the surface and anaerobic decay of deeper, older peat layers. Significant anaerobic decay of older peat further down the peat column leads to a considerable loss of mass over time resulting in a concave mass/age profile (Clymo, 1984). The shape of a carbon mass accumulation curve derived from a dated peat core is therefore indicative of the degree of anaerobic decay integrated over the entire peat column.

The four detailed records of carbon mass accumulation from western Indonesia exhibit either a linear or a convex relationship between carbon mass and age (Fig. 3). The linear mass/age profile of the two coastal peatlands (Fig. 3b) indicates constantly high rates of accumulation at the surface and insignificant anaerobic decay of deeper peat. In contrast, the convex curves for carbon mass accumulation of the two inland peatlands (Fig. 3a) indicate limited anaerobic decay of older peat and reduced apparent rates of peat accumulation during later stages of peatland development. Reduction in the apparent rate of peat accumulation can result either from decreased biomass production and carbon inputs or from increased *aerobic* decay and carbon loss. Pollen records from the Sebangau peatland do not indicate forest decline and substantially reduced net primary production of the peat swamp forest vegetation (Morley, 1981, 2013), suggesting that the convex curve shape is caused by increased aerobic carbon losses.

Collectively, the carbon-mass accumulation curves suggest insignificant anaerobic decay in western Indonesian peatlands. This inference is corroborated by consistently low amounts of released methane – a metabolic product of anaerobic decay – from the peat swamp forests of Southeast Asia (Couwenberg et al., 2010; Pangala et al., 2013) and the young ^{14}C signature of the emitted methane (Nakagawa et al., 2002). Therefore, carbon storage across Indonesian peatlands can be modelled on the basis of reconstructed carbon accumulation rates without adjusting for anaerobic decay rates.

Peat carbon accumulation rates clearly show a distinct regional pattern across western Indonesia where the coastal peatlands represent the most significant carbon sink. Also from a global perspective these coastal peatlands were an important sink for carbon. The Holocene mean CAR of coastal Indonesian peatlands is 77 g C m^{-2} yr^{-1} (Dommain et al., 2011), which is four times higher than that of northern peatlands (18.6 g C m^2 yr^{-1}; Yu et al., 2009) and more than six times higher than that of the subtropical peatlands of the Florida Everglades (12.1 g C m^{-2} yr^{-1}; Glaser et al., 2012).

As a result of the high rate of carbon accumulation, the coastal peatlands of western Indonesia store large amounts of carbon per unit area in spite of their relatively young average age of 2.2 ka. With a carbon pool of 16.5 Pg C over an area of 104,500 km^2, the mean carbon density of coastal peatlands is about 1600 Mg C ha^{-1}. Sheng et al. (2004) report a similar mean carbon density of 1500 Mg C ha^{-1} for the southern part of the West Siberian peat basin, which is, however, largely of Early Holocene age (Smith et al., 2004). Although the average age (4.2 ka) of peatlands in Finland is twice that of the coastal Indonesian peatlands, they store considerable less carbon per unit area ($\sim$1000 Mg C ha^{-1}; Turunen et al., 2002).

The high carbon density of coastal peatlands is further illustrated by a comparison of the Siak Kanan and Teluk Keramat peat cores (Fig. 3b) with cores of the same age from the Red Lake Peatland in northcentral North America. The Red Lake Peatland stores an estimated 1900 Mg C ha^{-1} (Gorham et al., 2003; C content of 51.7% from Gorham, 1991), compared to an estimated 3500 Mg C ha^{-1} in the two Indonesian coastal sites.

The history of the peatland carbon sink of western Indonesia can be divided into two major phases (Fig. 2). During Phase 1 (15–5.5 ka) inland peatlands dominated and the total carbon sink remained below 1 Tg C yr^{-1}. With the onset of significant coastal peatland formation and expansion in Phase 2 around 5.5 ka total carbon storage increased and remained at $\sim$3 Tg C yr^{-1} between 4 and 2 ka. The subsequent doubling of the coastal peatland area during the past two millennia increased the total rate of carbon storage to 7.2 Tg C yr^{-1} (Figs. 2 and 9).

This recent rate of carbon storage of Indonesian peatlands is comparable to the average Holocene rate of 6.1 Tg C yr^{-1} for the West Siberian Lowlands (Smith et al., 2004), but lower than the average 11 Tg C yr^{-1} for the past two millennia in this same region (Beilman et al., 2009). It is important to note, however, that the 590,000 km^2 of peatlands in the West Siberian Lowlands comprise the largest concentration of peatlands in the world (Walter, 1977; Sheng et al., 2004) that are collectively 4.5 times larger in area than the peatlands of western Indonesia. The smaller size of the Indonesian peatland region further emphasizes its effectiveness as a major global carbon sink during the Late Holocene. Shortly before human disturbance, western Indonesia was apparently the most spatially-effective peatland region for carbon storage on Earth. This inference is supported by a comparison to northern peatlands, which accumulate 76 Tg C yr^{-1} on an area basis of 3.3 $\times$ 10^6 km^2 (Gorham, 1991), whereas Indonesian peatlands accumulated 10% as much carbon on an area only 4% as large.

Yu (2011) derives negative net carbon balances for all tropical peatlands for three 1000–2000 year time periods during the Holocene, during which anaerobic decay exceeded carbon accumulation. These model predictions are neither supported by the stratigraphic records from western Indonesia nor by our reconstructions of peatland origin and progressive expansion. Our results demonstrate that the tropical peatlands of western Indonesia were a persistent carbon sink since 15 ka (Fig. 2; see sect. 5.6.).

Fifteen thousand years of persistent carbon storage resulted in an exponentially growing pool of peat carbon in western Indonesia. This carbon pool was only 3.7 Pg C (16%) by the end of Phase 1 (5.5 ka), but increased to 23.1 Pg C during Phase 2, with coastal peatlands responsible for 84% (16.3 Pg C) of total carbon added (Figs. 2 and 4). The expansion of coastal peatlands in Kalimantan and Sumatra maintained the exponential growth of the peat carbon pool in western Indonesia despite the stagnating rate of annual carbon storage in the inland regions of Borneo (Fig. 2).

Our estimate of 23.1 Pg C for the total peat carbon pool in western Indonesia is considerably smaller than the 33.3 Pg C estimate of Wayunto et al. (2003, 2004). In our reconstruction we applied a bulk density value of 0.076 g cm^{-3} for coastal peatlands (Dommain et al., 2011), which represent 80% of the total peatland area. Wayunto et al. (2003, 2004), in contrast, assume a much higher dry bulk density of the peat, particularly for the coastal peatlands of Sumatra (Wayunto et al., 2003). Other estimates for the carbon pool of Indonesian peatlands (Jaenicke et al., 2008; Page et al., 2011) are based on assumed mean values for peat depth that are 1–2 m higher than indicated by the peat atlases. The area-weighted mean peat depth of all polygons is not more than 3.6 m (using the upper boundary of the peat depth classes; Wayunto et al., 2003, 2004). Even if the total peatland carbon pool in western Indonesia is likely below previous estimates, it is still equal to the total carbon pool in above and below ground biomass of all tropical forests of Indonesia (incl. Papua) that, however, cover an over 10 times larger area (Saatchi et al., 2011).

5.4. The role of Indonesian peatlands in the Holocene global carbon cycle

Peatlands have been recognised as globally important carbon sinks over long timescales (e.g. Gorham, 1991; MacDonald et al., 2006; Wang et al., 2009; Kleinen et al., 2010; Yu, 2011) and therefore have had a global, net climatic cooling effect over the Holocene (Frolking and Roulet, 2007).

During the transition from the Last Glacial to the present interglacial the concentration of atmospheric CO_2 increased from 190 parts per million by volume (ppmv) at 17 ka to 268 ppmv at 10.5 ka (Monnin et al., 2001). This rise was followed by a brief decline to 260 ppmv by ~8 ka, after which the atmospheric CO_2 concentration rose to the pre-industrial level of 280 ppmv (Indermühle et al., 1999; Flückiger et al., 2002, Fig. 9).

Indermühle et al. (1999) explained the post 8 ka rise in CO_2 by the release of 195 Pg C from the terrestrial biosphere. Records of ^{13}C–CO_2 from Antarctic ice-cores, however, invalidate the sole attribution of the CO_2 rise to a biospheric source (Elsig et al., 2009). Instead this rise is now primarily ascribed to coral reef formation (Ridgwell et al., 2003; Kleinen et al., 2010) and to carbonate compensation by the world's oceans (Broecker et al., 1999, 2001) as a response to the rapid carbon uptake of the biosphere of about 700 Pg C during the glacial–interglacial transition. In contrast, land-biosphere release is considered to have been only a small source (Elsig et al., 2009; Wang et al., 2009).

Based on our calculations, Indonesian peatlands with a pool of 0.6 Pg C by 11.7 ka made no significant contribution to the estimated land-biosphere carbon uptake of ~700 Pg C prior to the onset of the Holocene. The Early Holocene reduction in atmospheric CO_2 was probably the result of a continued and substantial uptake of carbon by the land biosphere that is estimated by Elsig et al. (2009) to be 290 Pg C with a modelled contribution by peatlands of 180 Pg C (but see Yu, 2011). Accordingly, Indonesian peatlands contributed only 1% (~3 Pg C) to the carbon storage in the global terrestrial biosphere at this time when they were still restricted to a few inland areas of Borneo. Indonesian peatlands were therefore not a significant factor for the Early Holocene decline in atmospheric CO_2.

Elsig et al. (2009) further suggest that peatlands stored another 40 Pg C over the past 5000 years. This estimate is very likely too small considering that western Indonesian peatlands stored almost 20 Pg C alone over this period, while other large peat basins in North America such as the Red Lake peatland, the Hudson Bay Lowland, and the Everglades formed over this same period (Glaser et al., 1981, 2004, 2012) and older northern peatlands continued to expand laterally (Korhola et al., 2010). Over this period the terrestrial biosphere may have lost about 40–90 Pg C, particularly related to the aridification of the Sahara-Sahel region (Brovkin et al., 2002; Elsig et al., 2009; Wang et al., 2009). In storing almost 20 Pg C since 5 ka, the young tropical peatlands of Indonesia contributed considerably to compensating concurrent terrestrial carbon losses elsewhere.

The carbon uptake of western Indonesian peatlands increased substantially after 5 ka from less than 2 to more than 7 Tg C yr^{-1}. This increase together with high carbon accumulation in other Southeast Asian peatlands of similar age (Dommain et al., 2011) may have contributed to the reduced rise in atmospheric CO_2 after 2.5 ka (Fig. 9). Based on a conversion factor of 2.12 Pg C/ppm (Denman et al., 2007) the lowland peatlands of western Indonesia stored an equivalent of approximately 11 ppm CO_2 prior to the onset of modern peatland degradation. The actual atmospheric impact might have been a CO_2 drawdown on the order of 1–2 ppmv, considering equilibration of the ocean to peatland carbon uptake. These results highlight the need to include the significant carbon uptake of Indonesian and presumably other equatorial peatlands into Holocene global carbon cycle models.

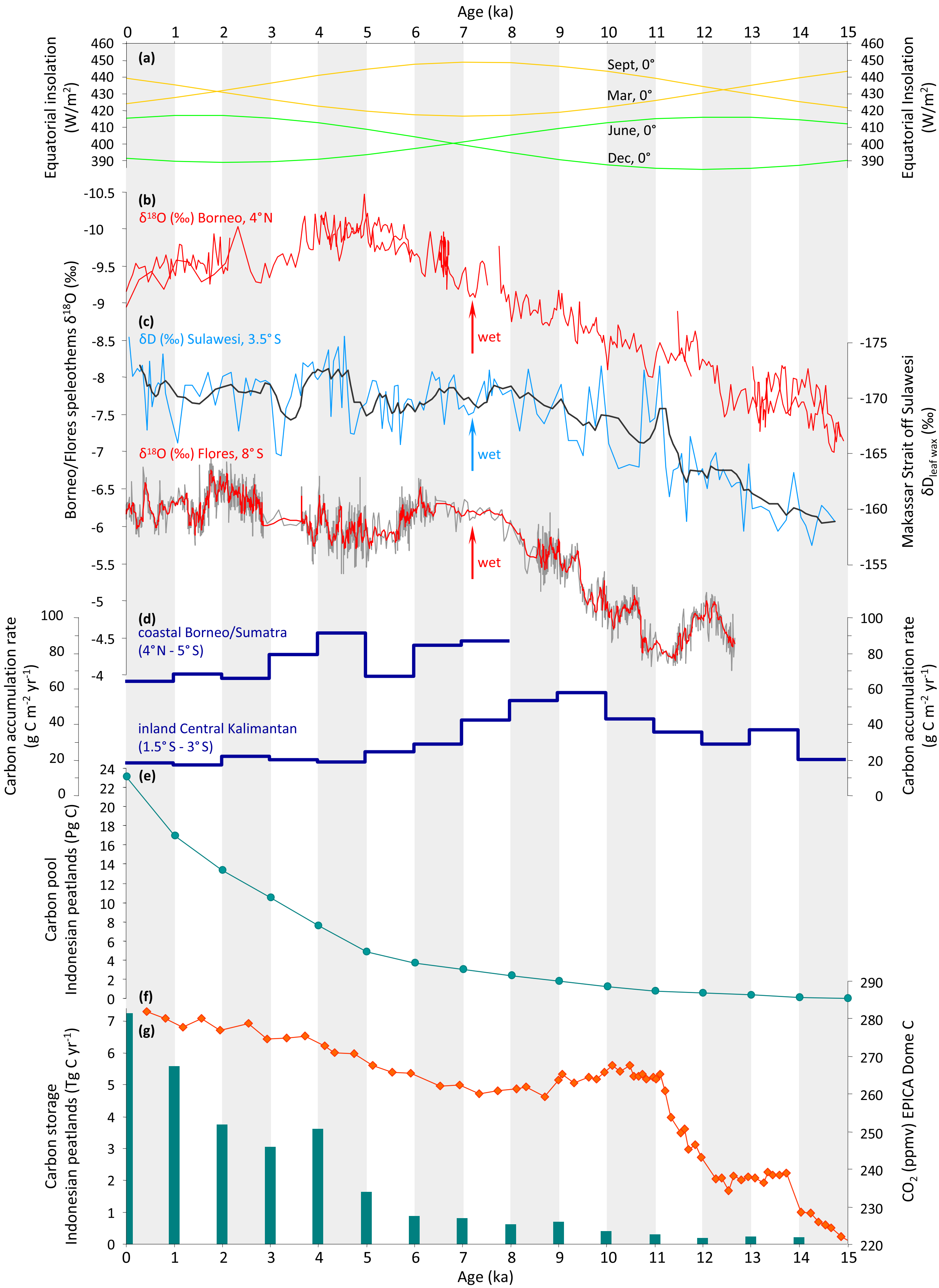
5.5. Climate controls on long-term peat-carbon accumulation

The millennial-scale records of carbon accumulation from Indonesian peatlands allow an examination of climate–carbon relationships by comparison with regional paleoclimate reconstructions. This tropical perspective will contribute to the debate on how climate factors influence carbon cycling in peatlands (e.g. Ise et al., 2008; Dorrepaal et al., 2009; Yu et al., 2009; Glaser et al., 2012).

The continuously high temperatures (26–27 °C) in equatorial Southeast Asia should in theory drive rapid soil carbon turnover prohibiting peat accumulation (e.g. Potonié, 1912; Davidson and Janssens, 2006; Glaser et al., 2012). The very high rates of peat and carbon accumulation in Indonesia, however, strongly suggest that factors other than temperature control peat carbon accumulation rates. Moreover, reconstructions of sea-surface temperatures from the Indo-Pacific Warm Pool (Linsley et al., 2010) imply only small temperature changes of ~0.5 °C over the Holocene. These changes were too small to account for the observed millennial-scale variations in carbon accumulation rates. Only the regional increase in temperature between 15 and 11 ka of around 1.5 °C (Linsley et al., 2010) might have influenced carbon accumulation. These rising temperatures, together with atmospheric CO_2 concentrations rising by > 40 ppmv (Monnin et al., 2001, Fig. 9) most likely stimulated the productivity of existing peat swamp forests in inland Kalimantan via the CO_2 fertilization effect (Prentice et al., 2011), allowing for carbon accumulation to increase from 1 to 30 g C m^{-2} yr^{-1} between 15 and 11 ka (Fig. 2; 9).

The major climatic change during the Lateglacial–Holocene transition in the Indo-Pacific Warm Pool was a constant increase in moisture availability associated with rising sea level and intensification of atmospheric convection and of the monsoon systems (e.g. Partin et al., 2007; Griffith et al., 2009; Tierney et al., 2012, Fig. 9). Carbon accumulation in Central Kalimantan clearly tracked this moisture increase until 8 ka. We therefore suggest that variations in carbon accumulation in general must have been primarily controlled by the regional hydroclimate in close association with rising sea levels.

Long-term hydroclimatic changes in the Indo-Pacific Warm Pool are driven by orbital forcing, dominated by the (half-) precessional cycle (e.g. Kutzbach, 1981; Partin et al., 2007). Over the course of the Holocene hemispheric changes in insolation led to a southward migration of the mean position of the Intertropical Convergence Zone and a reversed hemispheric strengthening of the monsoon systems (e.g. Haug et al., 2001; Wang et al., 2006). During the Early Holocene the summer insolation maximum in the northern hemisphere resulted in a northward shift of the Intertropical Convergence Zone and a strong Asian summer monsoon (Dykoski et al., 2005). This time period corresponded to a weaker Australian (summer) monsoon during boreal winter as recorded in speleothems from Flores (Griffith et al., 2009, Fig. 9). The stronger Asian summer monsoon likely enhanced cross-equatorial air flow and moisture transport, which presumably resulted in wet conditions in southern Borneo and Sumatra during June to September (the modern dry season). Moreover, higher than present September


Age (ka)
(a)
Sept, 0°
Mar, 0°
June, 0°
Dec, 0°
Equatorial insolation (W/m²)
Equatorial Insolation (W/m²)
(b)
δ¹⁸O (‰) Borneo, 4°N
(c)
δD (‰) Sulawesi, 3.5°S
δ¹⁸O (‰) Flores, 8°S
wet
Borneo/Flores speleothems δ¹⁸O (‰)
Makassar Strait off Sulawesi δD leaf wax (‰)
(d)
coastal Borneo/Sumatra (4°N - 5°S)
inland Central Kalimantan (1.5°S - 3°S)
Carbon accumulation rate (g C m⁻² yr⁻¹)
(e)
Carbon pool Indonesian peatlands (Pg C)
(f)
(g)
Carbon storage Indonesian peatlands (Tg C yr⁻¹)
CO₂ (ppmv) EPICA Dome C

insolation at the equator during the Early Holocene suggests enhanced convective activity and wetter conditions during the end of the modern dry season. These probable climate mechanisms may explain the maximum rates of carbon accumulation between 10 and 8 ka in inland Central Kalimantan and the peak in the rate of carbon storage of 0.7 Tg C yr^{-1} at 9 ka (Fig. 9).

However, besides insolation forcing, flooding of the Sunda Shelf has been identified as an important mechanism for controlling moisture supply in the Indo-Pacific Warm Pool and for the intensification in monsoonal rainfall (Griffith et al., 2009, 2012; DiNezio and Tierney, 2013). Coincident with the opening of the seaway between the South China Sea and the Java Sea south of Borneo at ~9.5 ka (Fig. 6) was the abrupt increase in Australian summer monsoon intensity that preceded increasing insolation in the southern hemisphere (Griffith et al., 2012, Fig. 9). The large area of flooded shelf provided a significantly increased moisture source for northwesterly monsoon flow (Nov.–March). Carbon accumulation of the Sebangau peatland (Page et al., 2004) directly responded to this abrupt strengthening of the Australian monsoon with a synchronous increase from ~20 to ~90 g C m^{-2} yr^{-1} at 9.5 ka (Fig. 3). The combined increase in year-round moisture supply (i.e. subdued rainfall seasonality) during the Early Holocene seems to have induced the highest rates of carbon accumulation in inland peatlands.

Rainfall continued to increase after 8 ka in northern Borneo, while it remained rather constant south of the equator (Fig. 9). However, mean carbon accumulation rates in Central Kalimantan decreased by 25 g C m^{-2} yr^{-1} between 8 and 6 ka (Fig. 9). This continuous decline in CAR corresponds to decreasing June insolation (Fig. 9), perhaps pointing at a progressively intensifying dry season in southern Borneo during boreal summer and an associated response of the peat-carbon balance. Interestingly, the Upper Kapuas peatlands, located directly on the equator, do not show a decrease in the mean CAR after 8 ka, but rather constant CAR until 4 ka (Fig. 2).

By 5 ka the mean position of the Intertropical Convergence Zone was centred over equatorial Southeast Asia (Partin et al., 2007) and apparently shifted farther south as seen in the rainfall maxima at 4 and 3 ka in the two southern paleoclimate records (Griffith et al., 2009; Tierney et al., 2012, Fig. 9). Millennial-scale variations in CAR from coastal peatlands largely correspond to the reconstructed rainfall changes from Borneo and Sulawesi (Fig. 9). The Holocene maximum in mean CAR of 91 g C m^{-2} yr^{-1} of coastal peatlands from Kalimantan and Sumatra was synchronous with the recorded Holocene rainfall maxima at 5–4 ka near the equator (Fig. 9). This maximum in CAR was also likely supported by a raised base level due to the sea-level highstand (sect. 5.2; Dommain et al., 2011). The following decline in mean CAR by 25 g C m^{-2} yr^{-1} mirrored reduced rainfall in Borneo and Sulawesi (Partin et al., 2007; Tierney et al., 2012).

Paleoclimate model simulations for Borneo and Sumatra produce higher than present June–November precipitation at 6 ka, most likely as a result of the equatorial September insolation maximum during that time (Tierney et al., 2012, Fig. 9). The declining long-term rates in CAR after 4 ka indeed seem to follow the trend in September insolation (Fig. 9). Increased rainfall seasonality over the past 3000 years thus seems to have influenced the carbon balance of both inland and coastal peatlands. This inference is in line with the precipitation model for peat formation in the tropics by Cecil and Dulong (2003) that predicts maximum rates of peat accumulation under a perhumid climate and its rapid decline with a dry season of more than 2–3 months. As already noted by Anderson (1983) not only the annual sum, but particularly the rainfall distribution over the year is important for peat accumulation in the tropics.

5.6. Natural carbon release

5.6.1. Late Holocene carbon release from aerobic peat decomposition

During the Late Holocene peat growth ceased over wide areas of inland peatlands in Central Kalimantan (Sieffermann et al., 1988; Dommain et al., 2011) and also in the undisturbed Nung peat domes in the Upper Kapuas basin (Anshari et al., 2012). This regional pattern of arrest in peat growth was probably a response to either external forcing factors such as changes in climate, sea level, or geomorphology or to internal peatland growth and decay factors.

A number of different mathematical models predict limits to peat bog growth, including 1) mass balance models in which cumulative mass loss of older carbon mass surpasses the addition of new mass (Clymo, 1984, 1992) and 2) hydrological models, which show that the maximum possible height of a bog is constrained by the elevation and width of its water table mound (Ingram, 1982; Glaser et al., 2004). An internally controlled, mass balance limit to bog height results in continued accumulation of young peat at the surface and a declining apparent rate of accumulation at deeper levels. An externally controlled limit of peat growth as defined by the width of an interfluvial divide and the rate of recharge would in contrast produce a transient reduction followed by a final cessation of peat accumulation as the peat dome approaches the maximum possible height of a water table mound in that specific hydrogeologic setting (Glaser et al., 2004).

The truncated peat profiles from inland Central Kalimantan do not conform to either of these models. Instead, they exhibit steady peat growth up to the surface with a high apparent rate of accumulation (Fig. 10). The truncations therefore suggest an abrupt change in some external hydrological control that causes a nonlinear response of the peatland ecosystem. Apparently peat accumulation did not resume again over the most recent millennium, implying that either the ecosystem could not recover or that the controlling factor preventing peat growth is still active. Today, the water-table in the truncated peatlands is 1–1.5 m below the peat surface in the dry season and the surface peat is strongly humified (Moore et al., 1996), indicating a clear change in the hydrological balance.

The regional hydraulic gradient was altered by the Late Holocene fall in sea level, which lowered base level and water table elevations in the peatland watersheds. With respect to water inputs from recharge (i.e. precipitation), available paleoclimatic records do not show dramatic changes in rainfall for the Late Holocene (Fig. 9). However, abrupt, high-frequency changes in ENSO activity are well documented for this period (Moy et al., 2002; McGregor and Gagan, 2004; Conroy et al., 2008, Fig. 10) leading Dommain et al. (2011) to

Fig. 9. Synthesis of paleoclimate and carbon cycle changes over the past 15,000 years. A) insolation at the equator (from Laskar et al., 2004). B) $\delta^{18}O$-based moisture reconstructions from speleothems from Borneo (upper red line: from Partin et al., 2007) and from Flores (grey line: actual data, red line: 5-point running mean: from Griffith et al., 2009). C) leaf-wax δD record of moisture history from the Makassar Strait off Sulawesi (blue line: actual data, black line 5-point running mean: from Tierney et al., 2012). For all three isotope records lower values indicate wetter conditions. D) mean carbon accumulation rates for coastal peatlands and for inland peatlands of Central Kalimantan. E) build-up of the western Indonesian peat carbon pool. F) atmospheric CO_2 concentration from the EPICA Dome C ice-core record (from Monnin et al., 2001; Flückiger et al., 2002). G) Annual rate of carbon storage of the total of western Indonesian peatlands at each millennial boundary.

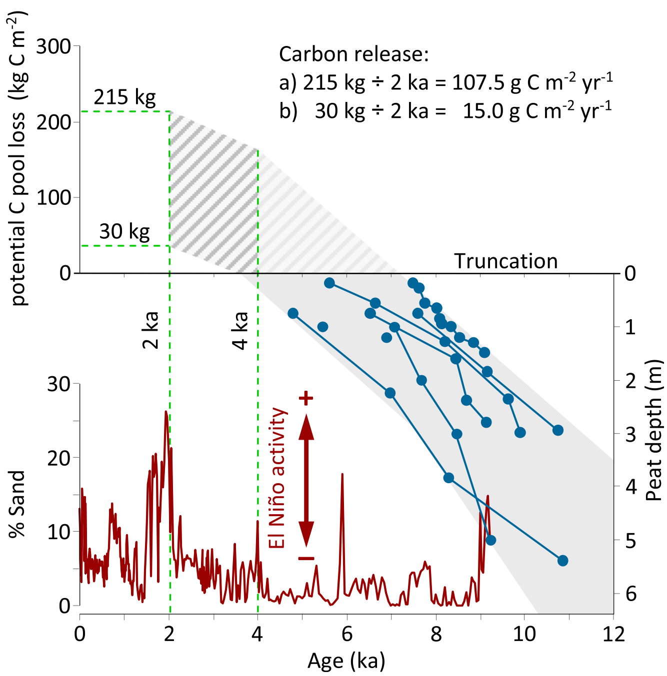


Fig. 10. Conceptual model of natural peat degradation for high-peat areas in inland Central Kalimantan. Age-depth models of truncated cores (core locations shown in Fig. 11) and extrapolated carbon mass accumulation until 2 ka. Carbon accumulation after 4 ka is assumed to decrease as a result of falling sea levels and increased El Niño activity. The abrupt increase of El Niño intensity at 2 ka triggers the release of the additionally accumulated carbon. El Niño record at the bottom is from Conroy et al. (2008).

suggest that inland peatlands in Kalimantan stopped accumulating and, in fact, started to lose surface peat in response to enhanced El Niño activity during the last 2 ka.

The stronger El Niño events were probably only a final trigger for aerobic decay of the surface peats of the inland peatlands. Falling sea level and lower precipitation most likely had begun to lower the water tables on these peatlands after 4 ka leading to slower rates of peat accumulation (see Fig. 3a: scenario b for the Palangka Raya peatland; Fig. 10). Subsequently, El Niño events periodically induced still deeper water-table drawdowns and promoted rapid aerobic decomposition of surface peat under the high ambient temperatures that are even higher during the positive phase of ENSO (Harger, 1995).

The spatial restriction of Late Holocene carbon losses (i.e. truncated peat profiles) to the high peat areas of Central Kalimantan (Fig. 11) can be explained by their specific topographic setting. The interfluvial high peat areas are located on podzol plains that are up to 15 m above adjacent, partly peat-filled river valleys (Sieffermann et al., 1987, 1988). These high peat areas benefited from the sea-level highstand when the adjacent valleys were widely flooded and the water-table mounds on the interfluves were higher. The fall in sea level after 4 ka in conjunction with overall reduced precipitation and increased ENSO activity would have drawn down the water table in these elevated central areas and have led to the decay of previously deposited peat. In contrast, falling sea level led to exposure of land surfaces within the lower adjacent river valleys enabling peat formation (sect. 5.2.).

The reconstructed annual rate of carbon release since 2 ka from the truncated high peat profiles ranges from 15 to 105 g C m^{-2} yr^{-1} (mean = 38 C g m^{-2} yr^{-1}; $n = 7$, Fig. 10). Today, the carbon release from just outside the high peat area in the undrained Sebangau peatland is on average even higher at 174 ± 203 g C m^{-2} yr^{-1} (Hirano et al., 2012, Fig. 11).

Over the past two millennia, annual carbon losses from the 5000 km^2 of degrading peatland in inland Central Kalimantan and the Upper Kapuas amounted to 0.15 Tg C yr^{-1}, resulting in a cumulative carbon release of 0.3 Pg C. The annual losses equal almost one half of the contemporaneous storage in the remaining, still accumulating peatlands of these two regions (Table 1). The cumulative losses correspond to 1% of the total current carbon pool of western Indonesia. Even in the absence of fires, changes in sea level and hydroclimate led to notable carbon release from Indonesian peatlands.

5.6.2. *Late Holocene carbon release from peat fires*

Fires are a major cause of carbon losses from tropical peatlands in the past few decades (Page et al., 2002). During the past 20 years, fires consumed the upper portions of peat profiles in Kalimantan down to depths of 15–150 cm with associated carbon releases of 6–104 kg C m^{-2} (Couwenberg et al., 2010). These carbon losses are equivalent to about 190–3300 years of peat accumulation, highlighting the scale of destruction of recent peat fires. Modern peat fires are most severe and widespread during El Niño events (van der Werf et al., 2008; Langner and Siegert, 2009) and the same can be assumed to hold for the Holocene epoch (Haberle et al., 2001).

Paleo-records of peat fires are restricted to a few localized sites in Kalimantan and no synoptic analysis on the fire history has yet been attempted for Southeast Asian peatlands. Holocene charcoal records exist from the peatlands of the Kutai basin, the Upper Kapuas basin and one site in Central Kalimantan (Anshari et al., 2001, 2004; Yulianto et al., 2004; Hope et al., 2005), but only in the Kutai basin could past fires be related to carbon losses. In contrast, the absence of charcoal indicates that fires were not responsible for the truncation of peat profiles in Central Kalimantan (Dehmer, 1993; Moore et al., 1996). The only evidence for local peat fires from this region exists for the period 9.2–7.5 ka, without any clear indication of associated peat losses (Yulianto et al., 2004).

Holocene fire activity in the Upper Kapuas basin was highest between 3.3 and 1.3 ka (Anshari et al., 2001, 2004) during the strongest El Niño phase (Fig. 10), suggesting a strong climatic control on both local and regional fire occurrence. However, no fire related peat truncations have yet been recorded in the Upper Kapuas basin.

Fires affected the peatlands of the Kutai basin mainly after 5 ka and higher frequencies of both micro- and macro-charcoal were recorded after 2.5 ka with maxima during the last millennium (Hope et al., 2005). These fires were most likely the result of human activities along the rivers of this region (Hope et al., 2005). Nevertheless, the time of highest fire activity coincides with the strong Late Holocene El Niño phase, indicating that associated droughts likely supported the spread of fires.

Severe peat fires during the Late Holocene were the likely cause for peat truncation and the subsequent formation of several large lakes in the Kutai basin as indicated by several metres of woody peat that underlie the lacustrine sediment layers (Hope et al., 2005). This mode of lake formation implies that peat fires must have consumed at least one metre of peat to permit the permanent inundation needed to prevent re-colonisation of the burnt area by trees (cf. Wösten et al., 2006; van Eijk et al., 2009).

The loss of one metre of peat in each of the burnt and truncated profiles from Kutai corresponds to the loss of 560–900 years of additional peat accumulation (see sect. 3.4.). Adding these time spans to the age of the truncations places the time of possible fire occurrences at 4.4 ka, 3.9 ka, 1.9 ka, and 1.4 ka respectively. The El Niño reconstruction of Moy et al. (2002) identifies 5, 5, 3, and 9 El Niño events per 100 years respectively for these four periods of possible fire activity. If fires were directly controlled by El Niño

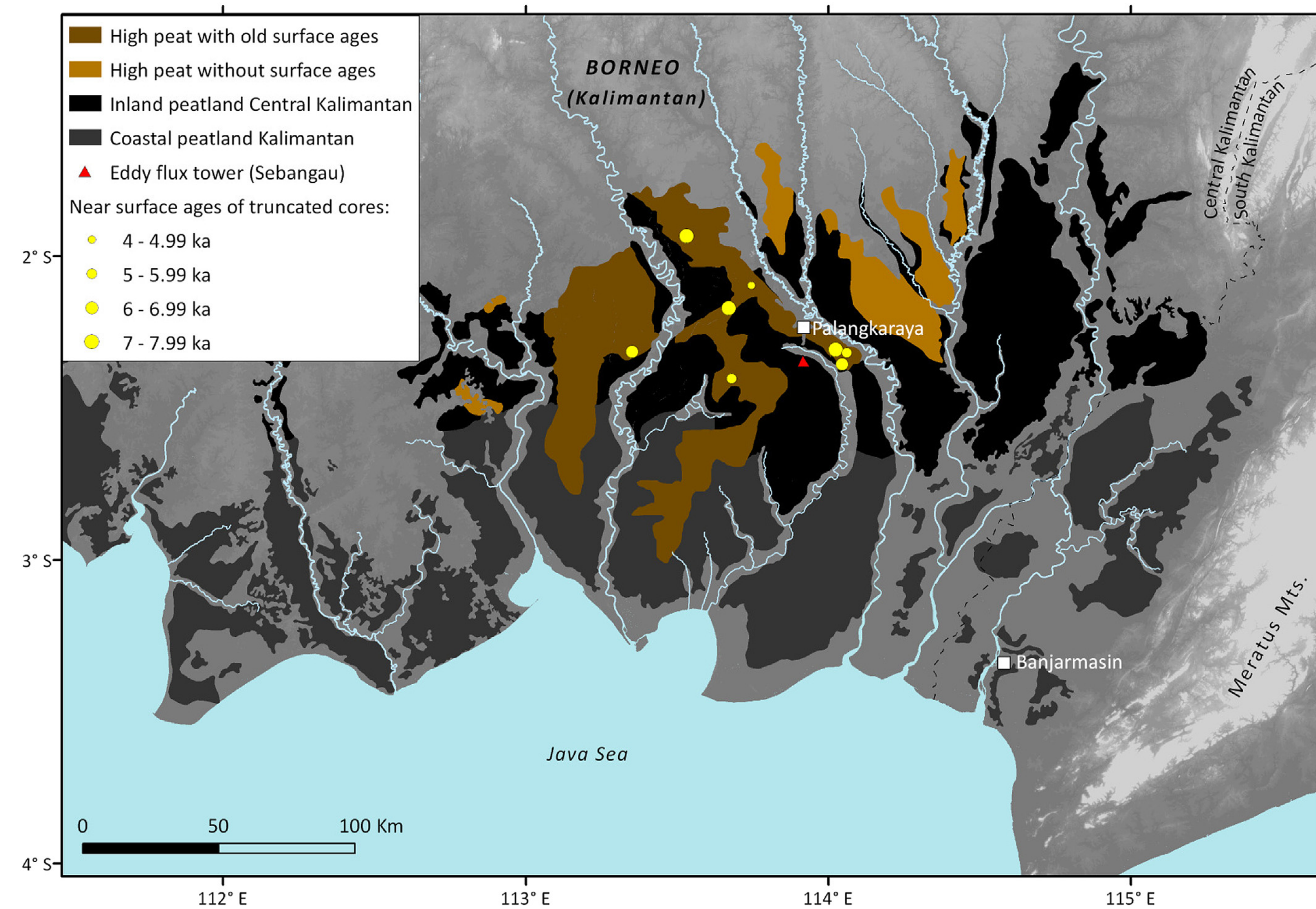


Fig. 11. Peat distribution in Central Kalimantan (southern Borneo) with coastal peatlands in dark grey and inland peatlands in black, brown and orange. Brown and orange are the high peat areas of Sieffermann et al. (1988). Brown coloured areas are assumed to release carbon as indicated by truncated cores (yellow dots). Dated cores are lacking for the orange coloured areas and these are not assumed to release carbon. Age-depth curves of the truncated cores are shown in Fig. 10. Eddy flux tower site of Hirano et al. (2007, 2012).

droughts then 3 to 9 fires per century could have occurred. This fire frequency would have permitted the combustion of one metre peat if single fires burnt peat layers of between 10 and 30 cm depth. The corresponding mass of released carbon would then be around 6–20 kg C m^{-2} per fire.

These potential carbon losses from Late Holocene peat fires are in the lower range of modern fire-related losses (6–104 kg C m^{-2}, Couwenberg et al., 2010). Integrated over burnt peatland areas (i.e. GIS polygons) the average carbon release of single fire events during the four fire episodes at 4.4 ka, 3.9 ka, 1.9 ka, and 1.4 ka would have been 0.66 Tg C, 1.96 Tg C, 3.6 Tg C and 0.1 Tg C, respectively. Despite their local impacts these carbon losses never surpassed contemporaneous carbon storage from all peatlands of western Indonesia (Table 1). Unless other widespread peat fires occurred in other parts of western Indonesia at similar times fire related carbon release did not offset peat carbon storage during the Late Holocene.

Although 10% (440 km^2) of the Kutai peatlands were likely affected by fire, the associated release of 25 Tg C corresponds to only 1% of the current carbon pool of Kutai. This cumulative fire-related carbon release from Kutai – which, at this stage, represents all known carbon losses from paleo-fires of western Indonesia – is less than the average carbon losses from modern peat fires across all of Borneo in one year (van der Werf et al., 2008). Fire related carbon losses from western Indonesian peatlands over the past 4000 years were an order of magnitude lower than the total carbon release from aerobic peat decomposition over the past 2000 years.

5.7. Current and future anthropogenic carbon losses

Today, Indonesian peatlands release massive amounts of carbon as a consequence of rapid deforestation, widespread drainage and recurring peat fires (Page et al., 2002; van der Werf et al., 2008; Couwenberg et al., 2010; Hooijer et al., 2010; Frolking et al., 2011). Dommain et al. (2012) estimated anthropogenic carbon emissions of 123 Tg C yr^{-1} from drainage-based peat decomposition in western Indonesia in 2007 (Fig. 12), constituting 85% of drainage related emissions in Southeast Asia (Dommain et al., in press). Natural carbon release from aerobic peat decay (0.15 Tg C yr^{-1}) in inland Kalimantan is insignificant in comparison to the 820 times higher anthropogenic losses.

Considering the remaining extent of undrained, primary peat swamp forest today, the rate of carbon storage in western Indonesian peatlands has been reduced to 1.2 Tg C yr^{-1} (Dommain et al., 2012, in press; Fig. 12), equalling only 17% of the annual pre-disturbance rate. Thus, peatland degradation not only causes substantial losses from the existing carbon pool, but also a significant reduction in inputs.

In addition to drainage-based peat decomposition, fires currently contribute significantly to regional carbon losses from peatlands. Carbon emissions from fires vary strongly between years. During wet La Niña years carbon emissions in western Indonesia are much lower (~35 Tg C yr^{-1}) than during dry El Niño years (~130 Tg C yr^{-1}; van der Werf et al., 2008). The fires of the major El Niño event of 1997/98 released approximately 700 Tg C (Heil et al., 2007; van der Werf et al., 2010) equalling 100 years of natural carbon accumulation for all western Indonesian peatlands. Between the years 2000 and 2006 peat fire emissions from Borneo (mainly Kalimantan) and Sumatra averaged about 75 Tg C yr^{-1} (van der Werf et al., 2008). The estimated carbon release from Late Holocene peat fires of 0.1–3.6 Tg C yr^{-1} is therefore small in comparison to current peat fire emissions.

The combined carbon losses of modern peatland drainage and peat fires in western Indonesia amount to about 200 Tg C yr^{-1} (range: 160-250 Tg C yr^{-1}; Fig. 12). This carbon loss is 28 times higher than the pre-disturbance carbon uptake indicating that the entire western Indonesian peatland region has switched from a carbon sink to a significant carbon source. If current carbon losses continue at the same rate the total peat carbon pool will be exhausted in 115 years (assuming full drainability). Yet, it seems that the rate of carbon release will continue to increase, at least over the coming 20 years. Miettinen et al. (2012) project the expansion of drained oil-palm plantations in western Indonesian peatlands to 25,000 km^2 in 2020 and 36,700 km^2 in 2030. This land use change alone would cause an increase in carbon emissions to 227 Tg C yr^{-1} by 2020 and to 250 Tg C yr^{-1} by 2030 (Fig. 12). Compared to the multi-millennial timescale it took to build up the Indonesian carbon pool, beginning 19,000 years ago, the ongoing carbon losses are highly disproportionate. Current human activities are leaving a geological imprint in the western Indonesian landscape that also affects the CO_2 concentration in the atmosphere.

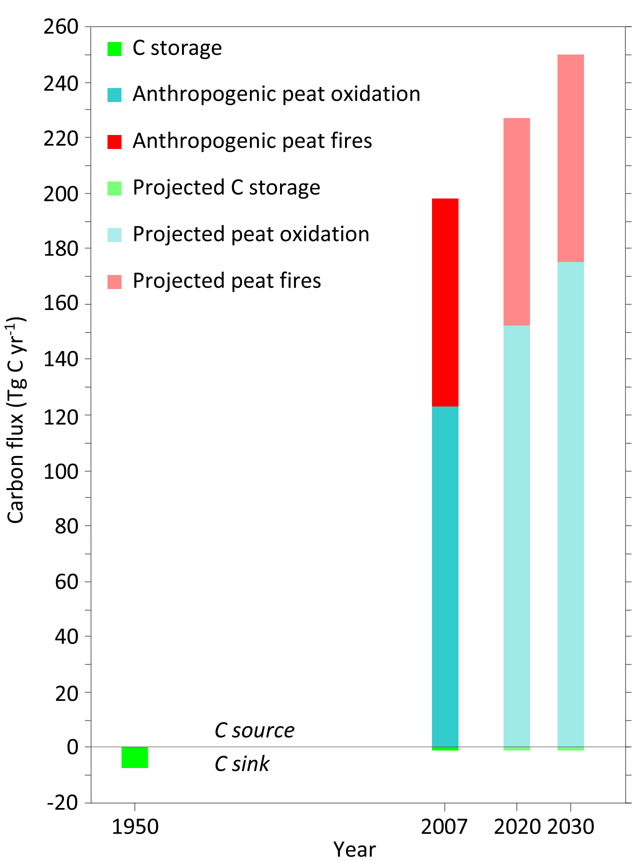


Fig. 12. Modern human impact on the carbon fluxes of western Indonesian peatlands. Fluxes from 1950 (before large-scale human impact) are compared to present fluxes and future projections. Peatlands have switched from a carbon sink to a carbon source.

6. Conclusions

Using a novel approach, based on peat depth maps, peat accumulation rates and available radiocarbon dates, we reconstruct the spatial expansion of peatlands in western Indonesia since the LGM. Our approach expands and improves upon the commonly applied use of basal dates as a proxy for peatland area in the past. The basal dates approach to peatland expansion underestimates young, shallow peat areas, including lateral peatland expansion. The new method shows that western Indonesian peatlands had their greatest expansion during the past 3000 years, several millennia later than suggested by the basal dates approach. The choice of method has significant consequences for the reconstruction of peatland impacts on atmospheric greenhouse gas concentrations. The basal dates approach is probably overestimating Holocene CO_2 uptake by peatland regions.

Sea-level change was the primary driver for peatland formation and expansion in the lowlands of the western Indonesian archipelago as it controls both the regional atmospheric moisture availability and the hydrological gradient on the islands. A sea-level height of 80 m bmsl seems to mark a threshold level for inland peatland formation in Borneo during the Late Quaternary. Periods of rapid deglacial sea-level rise of over 10 mm yr^{-1} induced peatland formation and expansion in inland Kalimantan between 20 and 7 ka. The formation of coastal peatlands in western Indonesia started only when the rise in sea-level was reduced to a rate of 2.4 mm yr^{-1} which marks an important threshold above which any future sea-level rise will likely lead to widespread inundation, particularly in view of anthropogenic peatland subsidence.

Changes in sea level and hydroclimate were not only important drivers for the growth of the carbon pool in western Indonesia but also for the degradation of inland peatlands, which released 0.3 Pg C during the past 2000 years. Moisture supply is the primary climatic control for peat accumulation in the tropics where high temperatures drive rapid carbon turnover whenever water tables are lowered. Perhumid conditions induced the highest rates of carbon accumulation in the past, whereas higher seasonality and El Niño droughts led to reduced accumulation and peat truncation. Peat bog growth in the tropics is thus limited by aerobic decay and not by anaerobic decay integrated over of the entire peat profile, which is insignificant in comparison to northern peatlands. This observation provides an additional explanation for the enormous thickness of Carboniferous and Tertiary coal deposits that are derived from tropical peat swamp forests.

The high rates of carbon accumulation of coastal peatlands and their uninterrupted expansion as a result of falling sea levels enabled continuously increasing rates of carbon storage during the Late Holocene. Increased land availability due to sea-level regression thus compensated for decreased moisture availability – as driven by the southward migration of the ITCZ and enhanced ENSO activity – to maintain the exponential growth of the western Indonesian peat carbon pool. Although Indonesian peatlands were a persistently growing carbon sink between 15 and 0 ka, they were insignificant in contributing to the massive uptake of carbon by the biosphere during the Early Holocene. However, after 5 ka this carbon sink increased substantially and partly compensated for contemporaneous biospheric carbon release from other regions and likely contributed to a reduced increase in atmospheric CO_2 concentrations since 2 ka.

Although Indonesian peatlands have persisted as a carbon sink for 15,000 years in spite of substantial changes in sea level and climate, they cannot withstand the pressure of modern human activities. Within the recent decades this important carbon sink has switched to a significant carbon source that contributes to currently rising atmospheric CO_2 concentrations.

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