Past and present carbon accumulation and loss in Southeast Asian peatlands

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Tropical peatlands store ~75 Pg carbon and have operated as long-term net carbon sinks throughout the Holocene. However, intensive land development is destabilizing these reservoirs, resulting in large carbon emissions to the atmosphere and loss of valuable low-latitude peat paleorecords.

Location and carbon storage

By area, peatlands have their greatest extent in the boreal and temperate zones (Immirzi et al., 1992) but tropical deposits, located in Southeast Asia, Africa, the Caribbean, and Central and South America, are also an important component of the global resource and terrestrial carbon (C) storage in both their above-ground biomass and underlying thick peat mass (Rieley et al., 1996; Page et al., 1999, 2004). A recent study (Page et al., submitted) indicates that tropical peatlands cover ~439,238 km^2 (~11% of global peatland area), with a peat C pool of 88.5 Pg (~17-19% of the global peat C pool (Immirzi et al., 1992)). Globally, the most important tropical peatlands occur in Southeast Asia (57% of total area; 68.5 Pg of C, representing 77% of global tropical peatland carbon stores). In this region, Indonesia holds by far the largest share (57.4 Pg or 65%), followed by Malaysia (9.1 Pg or 10%) (Fig. 1).

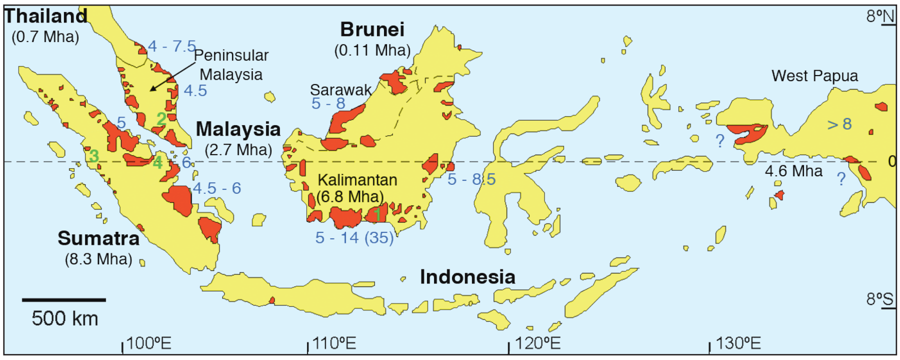

*Figure 1: Distribution (**red shading**; in million ha, after Rieley et al., 1996) and approximate dates of initiation (**blue numbers**; cal ka BP) for peatland in Southeast Asia. Question marks indicate unknown peatland initiation age. **Green numbers** indicate the location of peatlands referred to in text: 1) Sungai Sebangau, 2) Tasek Bera, 3) Tao Sipinggan, and 4) Siak Kanan.*

Peatlands in Southeast Asia: types

Most Southeast Asian peatlands are ombrotrophic (precipitation-fed), although a few basin peatlands are minerotrophic (receiving surface runoff and/or groundwater), and support a vegetation of dense swamp forest. A combination of low topographic relief, impermeable substrates and high effective rainfall have provided conditions suitable for slow decomposition of organic material and the accumulation of thick (often >10 m) deposits of woody peat.

Three categories of lowland peatlands have been proposed: (i) coastal, (ii) sub-coastal or valley, and (iii) high, interior or watershed (Rieley et al., 1996; Page et al., 1999, 2006). Coastal peatlands occur along maritime fringes and in deltaic areas where they have developed over marine sediments, inland of accreting mangrove and Nipa palm swamps (Anderson, 1983; Staub and Esterle, 1994). Sub-coastal peatlands are further inland at slightly higher elevations (5-15 m asl) where peat formation was initiated as a result of rising ground water levels, linked to changes in sea level. High peatlands have been described from Central Kalimantan (Indonesian Borneo; Fig. 1) up to 200 km inland from the coast, where they cover low-altitude, watershed positions (10-30 m asl) (Sieffermann et al., 1988, 1992; Page et al., 1999; Morley, 2000). In addition, some isolated basin deposits have formed in and around lakes (e.g., Anshari et al., 2001, 2004; Wüst and Bustin, 2004; Dam et al., 2001; van der Kaars et al., 2001; Penny, 2001; Maxwell, 2001; Maxwell and Liu, 2002).

Peat and carbon accumulation

Only a few peatlands in Southeast Asia have been investigated for peat structure, age, development, and rates of peat and C accumulation (e.g., Neuzil, 1997; Brady, 1997; Page et al., 2004; Wüst and Bustin, 2004), the onset and development of which range from the Late Pleistocene to the Holocene. Paleoenvironmental studies of peatlands in Borneo reveal initiation dates ranging from Late Pleistocene (~40 ^{14}C ka BP) in Lake Sentarum basin, West Kalimantan (Anshari et al., 2001, 2004) to

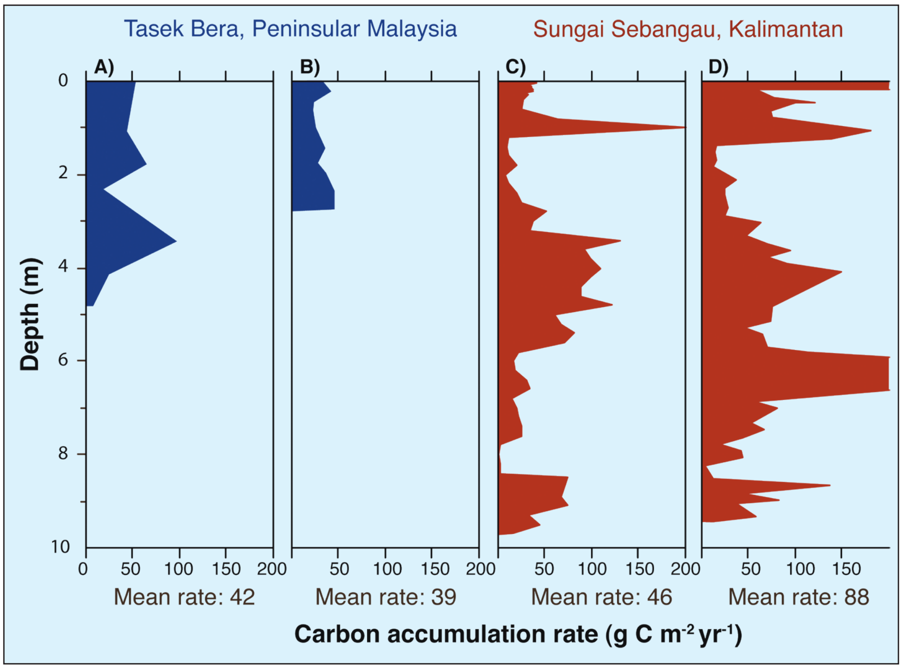

*Figure 2: Carbon accumulation rate down-core for a minerotrophic peat on Peninsular Malaysia (Tasek Bera; **A**) core B53, **B**) core B144) (Wüst and Bustin, 2004) and an ombrotrophic peat on Kalimantan (Sungai Sebangau; **C**) core SA6.5, **D**) core Kal1, located within 1.5 km of each other) (Page et al., 2004; Wüst unpub. data).*

~23 ^{14}C ka BP for high peat in Central Kalimantan (Page et al., 2004), to the early Holocene (10 - 8 cal ka BP) for other high and sub-coastal deposits (Neuzil, 1997; Sieffermann et al., 1988; Staub and Esterle, 1994). In comparison, the extensive coastal deposits are the youngest peatlands in the region, with initiation around 3.5 - 6 cal ka BP (e.g., Anderson and Muller, 1975; Staub and Esterle, 1994).

A detailed record of peat accumulation from Central Kalimantan (Page et al., 2004) (Figs. 1, 2), reveals a relatively rapid initial rate of peat accumulation of 1 mm a^{-1} between 24 - 26 cal ka BP (22 - 23 ^{14}C ka BP), equivalent to a C accumulation rate of ~54 g C m^{-2} a^{-1}. This period probably lasted for several thousand years until the onset of the drier Last Glacial Maximum (LGM) (~18 ^{14}C ka ago), when conditions were less favorable to peat formation. During and after the LGM, until ~13 cal ka BP, peat and C accumulation rates were low at only 0.04 mm a^{-1} and 1.3 g C m^{-2} a^{-1}, respectively. The beginning of the Holocene, however, saw a rapid resurgence: between 8.54 and 7.82 cal ka BP the peat accumulation rate increased from 0.60 to 2.55 mm a^{-1} with an average C accumulation rate of 92 g C m^{-2} a^{-1} and the formation of more than 3.5 m of peat over a ~2.2 ka period (~9.1 - 6.9 cal ka BP). Rapid sea-level rise at the end of the LGM led to the transgressive flooding of the Sunda and Sahul Shelves. Sea level changes were associated with warmer sea surface temperatures (Kienast et al., 2001, 2006), which likely resulted in increased precipitation, and the backing up of rivers owing to reduced drainage (Siefferman et al., 1987). In combination, these conditions favored peat accumulation in coastal areas with low topographic relief, such as along the seaboards of Borneo, Sumatra, E and W Peninsular Malaysia, and further inland in Borneo on interfluvial divides (Fig. 1; Wüst et al., 2007).

Towards the end of this period of rapid accumulation for inland high peats (~6 cal ka BP), large, relatively flat areas of new coastal environments were being exposed throughout the Southeast Asian region as rising sea levels stabilized and fell slightly during the mid-Holocene (Geyh et al., 1979; Hu et al., 2003; Tjia, 1992; Tjia et al., 1984). The combination of favorable topographic and climatic conditions led to rapid peat accumulation across coastal lowlands (Wilford, 1959; Hesp et al., 1998; Staub and Esterle, 1994). In the Rajang Delta of Sarawak (Fig. 1), 4.45 m of peat accumulated between 6.4 and 2.06 cal ka BP (~1.26 mm a^{-1}; Staub and Esterle, 1994), whilst on the east coast of Sumatra, peatlands underwent very rapid accumulation with initial rates as high as 6-13 mm a^{-1} between 5.3 - 4.3 cal ka BP (Neuzil, 1997). A study from inland Tasek Bera on Peninsular Malaysia (Wüst and Bustin, 2004) also indicates peat initiation at this time, with highest rates occurring after 4.3 cal ka BP. The rapid accumulation of inland peats, subsequently followed by the formation of deep coastal peat deposits, must have provided a large regional sink for atmospheric carbon throughout the Holocene.

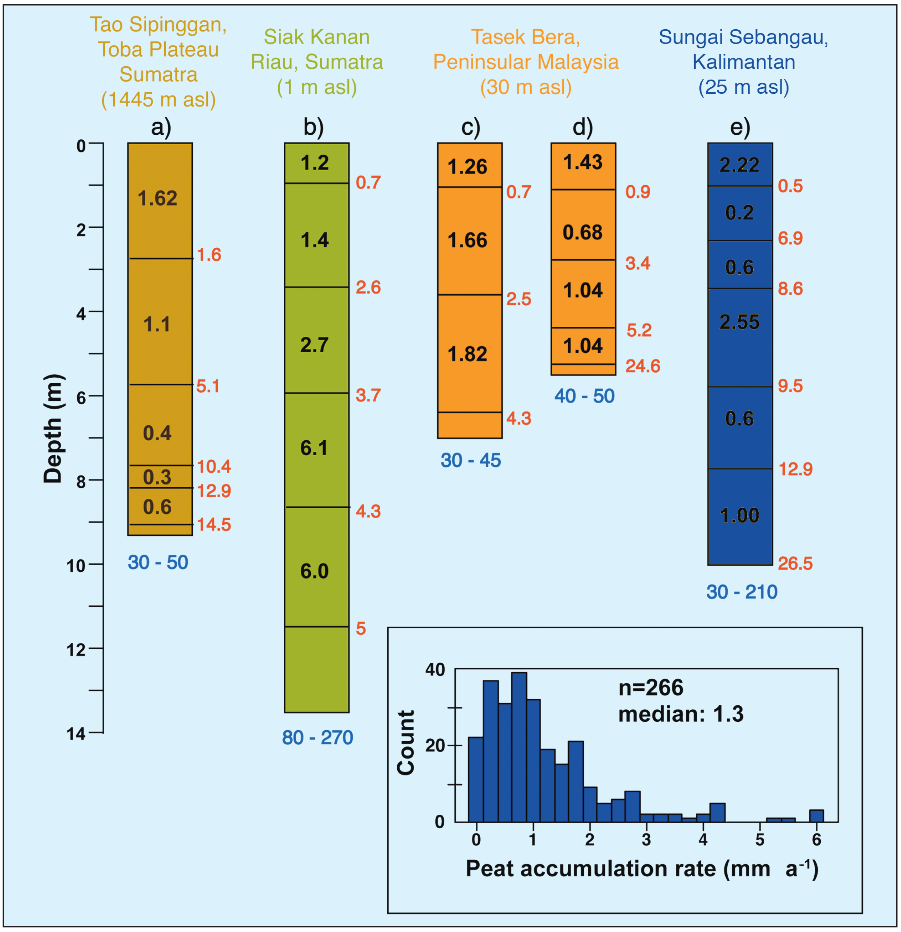

*Figure 3: Selected peat sections from various sites in Sumatra (**a**, **b**), Peninsular Malaysia (**c**, **d**) and Kalimantan (**e**) showing approximate age of peat accumulation (cal ka BP, red numbers), peat accumulation rates (mm a^{-1}, black numbers) and carbon accumulation rates (g C m^{-2} a^{-1}, blue numbers); the latter vary between 30-270 g C m^{-2} a^{-1}. Data from Maloney and McCormac, 1995 (Tao Sipinggan); Neuzil, 1997 (Siak Kanan); Wüst and Bustin, 2004 (Tasek Bera); Page et al., 2004 (Sungai Sebangau). Inset: Histogram of peat accumulation rates of 266 samples across sites in Sumatra, West Java, Kalimantan, Sarawak, Peninsular Malaysia, Thailand, Sulawesi and New Guinea.*

From carbon sink to carbon source

Radiocarbon dating of peat material from sites across Southeast Asia (Fig. 3) reveals a long-term median peat accumulation rate of ~1.3 mm a^{-1} (i.e., 67 g C m^{-2} a^{-1} assuming a peat bulk density of 0.09 and 56% C content), which is about 2-10 times the rate for boreal and subarctic peatlands (0.2-0.8 mm a^{-1}) (Gorham, 1991). Currently, however, most, if not all, remaining peatlands in Southeast Asia are to some extent degraded with many no longer functioning as C-accumulating systems. Anthropogenic activity is the principal cause of this shift, although longer-term climate-induced changes are also important in some locations (Page et al., 2004). Deforestation, drainage, large-scale conversion to plantation agriculture and regular fires have resulted in carbon flux to the atmosphere and loss of carbon sequestration function. Current C emissions are of the order ~360 Mt C a^{-1} (~170 Mt C a^{-1} from drainage-related peat decomposition (Hooijer et al., 2006); 190 Mt C a^{-1} from peat fires (Page et al., 2002; van der Werf et al., 2008)), equivalent to 4.5% of global emissions from fossil fuels.

Further detailed investigations of tropical peatland archives could result in new information about ENSO, monsoons and ITCZ migration, as well as an improved understanding of Holocene climate evolution in Southeast Asia and the long-term role of tropical peatlands in the regional and global C cycle. Unfortunately the opportunities to study these paleorecords are now being compromised by the rapid rate of peatland loss owing to human activities.

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Inception, history and development of peatlands in the Amazon Basin

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The existence of peatlands in the Amazonian lowlands has only recently been confirmed, owing to the remoteness of the area. These peatlands are important for regional carbon cycling and habitat diversity, and represent valuable potential resources for paleoecological research.

The Amazon's floodplain peatlands

Amazonia, the world's largest continuous area of humid tropical lowland rainforest, is famous for its dense river network, large seasonal variations in water level (on average 10 m at Manaus, Brazil), and extensive floodplains and wetlands covered by *Mauritia* palms, floodplain forest or savanna-like vegetation (Irmler, 1977; Junk, 1983; Junk and Piedade, 2005; Keddy et al., 2009). Despite the great extent of wetlands within the Amazon Basin, the existence of tropical peatlands has rarely been considered (but see Suszczynski, 1984; Schulman et al., 1999; Ruokolainen et al., 2001; Guzmán Castillo, 2007). Two studies carried out recently in Peruvian lowland Amazonia (Loreto region, Fig. 1) by members of the Amazon Research Team of the University of Turku (Finland) reveal that peat deposits, up to 6 m thick, are widespread on floodplain wetlands of the Western Amazon Basin (Lähteenoja et al., 2009a, 2009b). Sixteen of seventeen studied wetland sites contained some kind of peat deposit. According to the very rough estimate of Schulman et al. (1999) based on local land-cover maps, satellite images, grey literature and sporadic field observations, Amazonian peatlands may cover up to 150 000 km^2, an area equivalent to half of Finland, and about 75 % of the area covered by the better-known tropical peatlands of Indonesia (Rieley and Page, 2005; Page et al., this issue).

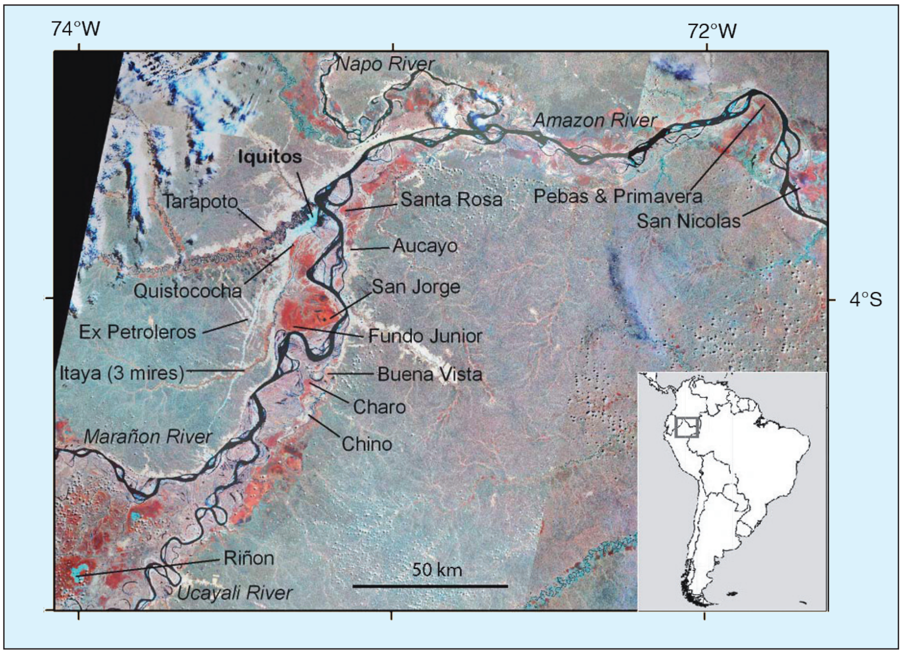

Figure 1: The location of the study sites (from Lähteenoja et al., 2009b, Fig. 1). The map is a mosaic of histogram-equalized Landsat TM satellite images (www.glcf.umiacs.umd.edu/). Palm swamps and forested wetlands have a reddish tone, more or less treeless open areas (like the open peatland Riñón) are blue-green, and other floodplain forests are pinkish to white.

History and development

Since their late Holocene inception, the peatlands identified in Peruvian Amazonia have accumulated peat and carbon at relatively high rates (0.94 - 4.88 mm per year, and 26 - 195 g C m^{-2} per year, respectively) (Fig. 2) and therefore constitute a strong carbon sink (Lähteenoja et al., 2009b). These accumulation rates are comparable to those of the Indonesian tropical peatlands (Page et al., 2004) and are higher than those of the boreal peatlands (Tolonen and Turunen, 1996).

The basal ages of five dated peat deposits varied from 0.588 cal ka BP (at 164 cm) to 2.945 cal ka BP (at 565 cm) (Lähteenoja et al., 2009b), which are considerably younger than basal ages determined in peatlands in many other regions of the world (cf., Korhola et al., 2010). There are several possible reasons for this. A paleoecological study of lake sediments in Peruvian Amazonia suggests that the dry conditions of the middle Holocene were followed by a period of increasingly wet conditions beginning some time between 4.2 and 2.54 cal ka BP (Bush et al., 2007). Although our oldest peat initiation dates coincide broadly with the onset of this wet interval, some of the peat deposits have much younger basal ages (Lähteenoja et al., 2009b), indicating that peat formation was not determined purely by climate. Peat initiation may be controlled by the dynamic lateral migration of western Amazonian rivers, characterized by meandering and avulsion (Kalliola et al., 1992; Neller et al., 1992; Pärssinen et al., 1996), which have the potential to erode and bury peat deposits. Peat accumulation probably began when an area with waterlogged conditions was isolated from the immediate destructive influence of rivers. Consequently, the Western Amazon Basin

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