Late Quaternary palaeoecology, palynology and palaeolimnology of a tropical lowland swamp: Rawa Danau, West-Java, Indonesia

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Abstract

Sedimentological, limnological and palynological analyses of a sediment core from a lowland site in West-Java, Indonesia, provide a detailed palaeoenvironmental record for the Late Glacial and the Holocene. The record suggests open vegetation under inferred drier climatic conditions for the Late Glacial. However, there is no unequivocal evidence for cooler conditions at this time. The onset of the Holocene coincides with a change to more humid climatic conditions, with the development of a fern-rich closed forest vegetation type. Dramatic changes in diatom community composition provide a striking record of habitat change associated with lake shallowing, but this process appears to be a result of basin in-filling rather than variations in precipitation/evaporation balance associated with climatic fluctuations. Evidence for human impact on the vegetation development is restricted to the last few hundred years. © 2001 Elsevier Science B.V. All rights reserved.

Keywords: Pollen; Diatoms; Palaeolimnology; Tropics; Climate change; Lake-levels

1. Introduction

Until recently, the reconstruction of Late Quaternary environmental change in the Indonesian region has been based on the analysis of high altitude sites, and on foraminiferal and oxygen isotope data from marine sediment cores. Palaeoenvironmental records from the high altitude sites are generally interpreted as indicating lower temperatures for the last glacial

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compared with today, without clear evidence for drier conditions (Stuivjs, 1993; Newsome and Flenley, 1988; Maloney, 1979; Maloney and McCormac, 1996), while foraminiferal and oxygen isotope data from marine sediment cores are usually interpreted as indicating similar or only slightly lower glacial temperatures than present (Thunell et al., 1994). Indications for both cooler and drier glacial conditions have been reported from marine palynological records in the region (Van der Kaars, 1991). However, there are very few terrestrial lowland records available and the palaeoenvironmental development of the Indonesian lowlands remains poorly understood.

In this study we present a record of Late Quaternary palaeoecological and palaeoclimatic change from...
Fig. 1. Location of Rawa Danau in relation to Indonesia and West-Java. Shading indicates the approximate extend of exposed Sunda Shelf during the height of glacial periods.
Fig. 2. Stratigraphic cross-section through the Rawa Danau area (after Rimbuaman, 1994).
Rawa Danau, West-Java, Indonesia. This is one of the few lowland sites in the region and may potentially address the ‘temperature paradox’ apparent in the comparison of marine and high altitude terrestrial records, and may also clarify the precipitation history of the lowland environments. Located close to the southern edge of the Sunda Shelf, the site is also expected to be sensitive to any changes in precipitation brought by the northwest monsoon when the Sunda Shelf was exposed at (glacial) times of lowered sea level.

2. Environmental setting

The Rawa Danau swamp is located in the caldera of the Danau volcanic complex, on the north-western tip of West-Java (Fig. 1), at an elevation of around 100 m above sea level (asl). The surrounding caldera rim is at an altitude about 140 m asl, adjacent high volcanic terrain is mostly between 400 and 700 m asl, but Mt. Karang, some 15 km to the south-east, reaches 1778 m asl. The Danau swamp area is drained by the Cidanau (or ‘Ci Danau’) River that originates on the northern lower slopes of Karang volcano, and the caldera forms a large part (70 km$^2$) of its catchment. The drainage level is fixed by the presence of an andesitic lava flow in the narrow outlet in the western part of the caldera. Early this century the river outlet level was artificially lowered by ca. 1 m to improve drainage of agricultural land in the caldera (Endert, 1932).

The Danau volcanic complex is of Plio–Pleistocene age and formed in a number of eruptive phases, culminating in the eruption of the voluminous Banten pumice tuffs (also known as ‘Bantam’ tuffs) and formation of the large caldera (Van Bemmelen, 1949; Santosa, 1991; Rusmana et al., 1991). Andesitic lava and volcaniclastic material form the steep northern-eastern rim of the caldera (Fig. 1). To the south occur volcanic footslopes and alluvial fans of younger volcanic cones (notably the active Mt. Karang). It appears that all other volcanic centres are presently dormant, but reliable information on possible Holocene eruptions in the Danau complex is not available. Within the caldera and swamp area proper, some active hot springs occur, notably in the vicinity of Mt. Djamungkal (or Mt. Jumungkal), where the swamp is said to be “of immeasurable depth” by the local population. The geomorphology of the inner caldera is indicative of the youngest sedimentation, with nearly all sediment derived from the southern volcanic slopes and deposited in alluvial fans and fluvial systems in the caldera. The eastern and northern caldera rim is largely inert, with the exception of some localised talus cones at the base of the escarpment. Organic lacustrine and swamp deposits and fine-grained fluvial sediment occur through the northern half of the inner caldera.

A survey of the swamp deposits by the Indonesian Geological Survey (Rimbaman, 1994) shows that deeper basin deposits (>6 m depth) consist of fine-grained lake deposits (organic silty clay and silty gyttja with shells), with minor intercalations of peat and fluvial sands (Fig. 2). The organic lacustrine deposits show a very homogeneous facies laterally and with depth, with very soft, extremely unconsolidated clayey gyttaes. Peat deposits prevail in the central swamp area (in the vicinity of the present dry-season lake, east of Mt. Djamungkal) with sequences of up to 4–5 m thickness just below the surface. Peaty facies with sandy intercalations occur also in the margins of the basin. A thin veneer of recent sandy clay floodplain deposits occurs at the surface. Thin volcanic ash layers, which show up due to pale colours and anomalous sandy textures, are evident throughout the sedimentary profile. The character of swamp and lacustrine deposits changes gradually towards the south, where facies are less organic and fine- to medium-grained clastic sediments prevail. These deposits also show more pronounced ripening.

Climate in the Indonesian region is dominated by the monsoonal circulation, the migration of the ITCZ, as well as the land-sea distribution in the Malay Archipelago (see Verstappen, 1975; Whitten et al., 1996). During the southern summer, the northern monsoon gathers large amounts of moisture while crossing the sea from the Asian high-pressure belt on its way to the ITCZ, which reaches northern Australia in January. At the ITCZ the moisture laden air rises, resulting in heavy rains in Indonesia (and northern Australia). During the southern winter, the southeast monsoon originates from the Southern Hemisphere high pressure belt and is relatively dry and cool, but gathers moisture as it crosses the
Indonesian waters on its way to the ITCZ, which in July reaches mainland Southeast Asia. The present climatic conditions in the area are typical for the Southeast Asian equatorial lowlands, but some local trends are noteworthy. Generally, average annual rainfall increases strongly with altitude (i.e. 1500–2000 mm in the northern coastal lowlands to 2500–3500 mm in the mountains). Rainfall distribution on the western tip of Java shows a strong south–north gradient from more than 3000 mm to less than 1500 mm annually. In the northern coastal zone, the dry season may last up to 6 months (Schmidt and Ferguson, 1951). Monthly rainfall data recorded at Anyer (approximately 5 km to the northeast of Rawa Danau) and Padarincang (on the southern slopes of the caldera itself) are presented in Table 1.

### Table 1
Monthly rainfall data recorded at Anyer, approximately 5 km to the northeast of the Danau complex, and Padarincang, on the southern slopes of the caldera itself (Berlage, 1949)

<table>
<thead>
<tr>
<th></th>
<th>Jan</th>
<th>Feb</th>
<th>Mar</th>
<th>Apr</th>
<th>May</th>
<th>Jun</th>
<th>Jul</th>
<th>Aug</th>
<th>Sep</th>
<th>Oct</th>
<th>Nov</th>
<th>Dec</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>Anyer (coast, 3 m asl)</td>
<td>361</td>
<td>278</td>
<td>244</td>
<td>181</td>
<td>135</td>
<td>90</td>
<td>79</td>
<td>70</td>
<td>53</td>
<td>107</td>
<td>175</td>
<td>328</td>
<td>2101</td>
</tr>
<tr>
<td>Padarincang (Rawa Danau, 98 m asl)</td>
<td>527</td>
<td>396</td>
<td>381</td>
<td>330</td>
<td>245</td>
<td>156</td>
<td>119</td>
<td>110</td>
<td>126</td>
<td>221</td>
<td>319</td>
<td>496</td>
<td>3426</td>
</tr>
</tbody>
</table>

3. Vegetation

The Rawa Danau area contains the largest and most significant area of freshwater swamp forest on Java (Endert, 1932; Whitten et al., 1996). Although these forests are at present highly disturbed, they are of enormous conservation value as the last representatives of a community type that once covered as much as 72,000 km² of Java (Scott, 1989, cited in Whitten et al., 1996; p. 487). The main vegetation types are mixed swamp forest, *Ficus retusa* swamp forest, open herbaceous swamp and dryland, largely deciduous, hill forest (Endert, 1932; Melisch et al., 1993; Whitten et al., 1996). Endert (1932) provides a detailed overview of the vegetation in the Danau swamp area. As this work is published in Dutch, and to our knowledge has not been reproduced in English, it is reviewed here in some detail, particularly given its relevance to the palaeoecological interpretation.

The vegetation in the southern part of the Rawa Danau area has been cleared for rice fields and villages. In the southeast the freshwater swamp forest still remains. Along rivers tall Poaceae, some Cyperaceae (usually more inland) and *Polygonum pulchrum* (syn. *Polygonum tomentosum*; Polygonaceae) are common. The freshwater swamp forest is dominated by *Ficus retusa* (Moraceae) while *Elaeocarpus littoralis* (Elaeocarpaceae), *Horsfieldia irya* (Myristicaceae) and *Mangifera gedebé* (Anacardiaceae) are common. In more open areas the characteristic knee roots of *Elaeocarpus littoralis* are covered by the climbing fern *Stenochlaena palustris* (Blechnaceae). Also present are *Barringtonia racemosa* (Lecythidaceae), *Ficus truncata*, *Eugenia polyantha* (Myrtaceae) and *Sarcocephalus cordatus* (Rubiaceae). Shrubs include *Ilex cymosa* (Aquifoliaceae), *Scolopia rianthera* (Flacourtiaceae), *Ficus hispida* and *Glochidion cyrostylus* (Euphorbiaceae). Epiphytes include *Trichosporum radicans* (Tiliaceae) and two bird’s nest ferns *Asplenium nidus* (Aspleniaceae) and *Polypodium punctatum* (Polypodiaceae). The herb layer is characterised by tall Cyperaceae, *Lasia spinosa* (Araceae) and *Helminthostachys zeylanica* (Ophioglossaceae). In more open places the climber *Columella trifolia* (syn. *Vittis trifolia*; Vitaceae) is common. Along streams, relatively tall *Gluta renghas* (Anacardiaceae) trees are common, with smooth stems and few epiphytes. *Lagerstroemia speciosa* (Lythraceae) is also present. Where the vegetation is disturbed by human impact and/or fire, the original forest is usually replaced by grasses and sedges with the occasional shrub or tree.

A mosaic of secondary and deciduous high forest vegetation dominates the hilly northwestern slopes of the caldera. The upper strata of these forests are characterised by *Pterocymbium javanicum* (Sterculiaceae), *Pterospermum javanicum* (Sterculiaceae), *Pangium edule* (Flacourtiaceae), *Ficus variegata*, *Alstonia scholaris* (Apocynaceae), *Artocarpus elastica* (Moraceae), *Anthemochalus cadamba* (Rubiaceae),
(Leguminosae) and *Sandoricum koetjape* (Meliaceae). Small trees and shrubs include *Mallotus blumeanus* (Euphorbiaceae), *Laportea* sp. (Urticaceae), *Villebrunnea rubescens* (Urticaceae), *Kleinhovia hospita* (Sterculiaceae), *Antidesma bunius* (Stilaginaeae), *Semecarpus heterophylla* (Anacardiaceae), *Cinnamomum* sp. (Lauraceae), *Myristicaceae*, *Carallia lucida* (Rhizophoraceae), *Saurauia* sp. (Actinidaeae), *Picrasma javanica* (Simaroubaceae) and *Strombosia javanica* (Olacaceae). A dense undergrowth of *Zalacca edulis* (Areaceae), Araceae, and tall Zingiberaceae is common.

The swamp vegetation in the north is partly replaced by rice fields, where *Pistia stratiotes* (Araceae) and *Phragmites karka* (Poaceae) are common at the edges of the remaining freshwater swamp forest. The swamp forest in this part of the caldera differs from that in the south-east. The soil is wetter and aerial roots are much more abundant. Trees are generally higher, and *Elaeocarpus littoralis* and to a lesser extent *Horsfieldia irya* are dominant. In contrast, *Ficus retusa* is almost absent. The undergrowth consists of palms (*Calamus* spp. and *Licuala* sp.), a tall *Cyperaceae*, *Alocasia bantamensis* (Araceae), and *Lasta spinosa*. Epiphytes are abundant, particularly *Asplenium nidus* and *Polypodium punctatum*, with *Stenochlaena palustris*, *Flagellaria indica* (Flagellariaceae) and a small *Lygodium* (Schizaceae) also common.

The broad transition zone between swamp forest and floating vegetation consists of a 1–1.5 m high vegetation comprised mostly of *Scleria multifoliolata* (Cyperaceae), *Dryopteris callus* (Aspleniaceae) and some *Stenochlaena palustris*. This gives way to vegetation dominated by Poaceae, which in turn changes gradually from a rooted to a floating habit towards the lake. Other taxa within this floating and rooted vegetation include *Alocasia bantamensis*, *Polygonum barbatum*, *P. dichotoma* and *Colocasia esculenta* (Araceae). Scattered throughout this floating grass vegetation are isolated trees and shrubs, or small groups of trees and shrubs, usually on slightly higher ground. These taxa include *Alstonia spathulata*, *Eugenia polyantha*, *Phyllanthus reticulatus* (Euphorbiaceae) and *Glochidion cyrtostylum*, accompanied by the climbers *Stenochlaena palustris*, *Flagellaria indica*, *Columella trifolia* (*Vitis trifolia*) and some *Calamus*. The lake itself is partly covered

*Gossampinus* sp. (syn. *Bombax* sp.; Bombaceae), *Spondias pinnata* (Anacardiaceae), *Dysoxylum* sp. (Meliaceae), *Dracontomelon mangiferum* (Anacardiaceae), *Pometia pinnata* ( Sapindaceae), *Erythrina* sp.
Table 2
Radiocarbon ages were calibrated to the bidecadal tree-ring/marine coral calibration curve using CALIB v. 3.0 (Stuiver and Reimer, 1993). Ages are calibrated in both years B.P. and B.C., and expressed at one sigma (88.3% confidence), and two sigma (95.4% confidence) ranges. Where multiple intersections of the calibration curve occur, the largest relative area of the probability distribution in each case was chosen to represent the most probable age range (where this occurs, the portion of the total range is presented in brackets and italicised). The median age of the two sigma calibrated range is given in bold type.

<table>
<thead>
<tr>
<th>Lab. No.</th>
<th>Depth (cm)</th>
<th>Uncalibrated age</th>
<th>Calibrated age B.P.</th>
<th>Calibrated age B.C./A.D.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wk-5229</td>
<td>270–280</td>
<td>350 ± 80</td>
<td>1s: 468 (391) 314</td>
<td>1s: AD 1482 (1559) 1636</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2s: 526 (400) 275 (0.97)</td>
<td>2s: AD 1424 (1550) 1675 (0.97)</td>
</tr>
<tr>
<td>Wk-5230</td>
<td>380–390</td>
<td>1810 ± 60</td>
<td>1s: 1813 (1752) 1691 (0.81)</td>
<td>1s: AD 137 (198) 259 (0.81)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2s:1839 (1704) 1571 (0.97)</td>
<td>2s: AD 111 (245) 370 (0.97)</td>
</tr>
<tr>
<td>Wk-5231</td>
<td>490–500</td>
<td>3890 ± 160</td>
<td>1s: 4459 (4272) 4084 (0.86)</td>
<td>1s: BC 2509 (2322) 2134 (0.86)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2s:4655 (4268) 3881 (0.94)</td>
<td>2s: BC 2705 (2318) 1931 (0.94)</td>
</tr>
<tr>
<td>Wk-5232</td>
<td>860–870</td>
<td>5450 ± 150</td>
<td>1s: 6405 (6280) 6036</td>
<td>1s: BC 4455 (4331) 4086</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2s:6532 (6280) 5915</td>
<td>2s: BC 4582 (4331) 3965</td>
</tr>
<tr>
<td>Wk-5233</td>
<td>1400–1410</td>
<td>11460 ± 160</td>
<td>1s: 13571 (13373) 13199</td>
<td>1s: BC 4455 (4331) 4086</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2s:13788 (13373) 13035</td>
<td>2s: BC 11838 (11424) 11085</td>
</tr>
<tr>
<td>Wk-5234</td>
<td>1640–1650</td>
<td>13200 ± 240</td>
<td>1s: 16096 (15743) 15353</td>
<td>1s: BC 14146 (13794) 13403</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2s:16425 (15743) 14922</td>
<td>1s: BC 14475 (13794) 12972</td>
</tr>
</tbody>
</table>

by free-floating aquatic plants such as *Pistia stratiotes*, *Eichhornia crassipes* (Pontederiaceae) and *Sesamum malayanum var. aquatica* (syn. *Hanguana malayanum var. aquatica*; Hanguanaceae).

Observations made during the 1996 field season suggest that the local vegetation has been substantially modified since Endert (1932) made his surveys in the Rawa Danau caldera. In particular, *Elaeocarpus littoralis*, described by Endert as a dominant component of the fresh-water swamp forest, has been replaced largely by *Ficus retusa*.

4. Materials and methods

Sediment cores were collected from the site using a hand-operated gouge auger. The main core (RD-3) was extracted from the approximate centre of the caldera, some 2 km to the south-west of the modern lake (Figs. 1 and 2). This site was chosen for coring as the survey of swamp deposits indicated the greatest thickness of lacustrine deposits occurred in this part of the caldera. The core was described and sub-samples were taken in the field and transported to the laboratory for analysis.

Core RD-3 is 17 m in length (Fig. 3). Five distinct sedimentary units are evident. Unit 1 (1700–1610 cm depth) is a black organic clay. Unit 2 (1610–480 cm depth) is a fine-grained green lacustrine clay (Munsell colour 10–7 Y 4//2) with abundant shells. Unit 3 (480–290 cm depth) is a woody peat unit. Unit 4 (290–50 cm depth) is black organic clay inter-bedded with woody peat deposits. Finally, Unit 5 (50–0 cm depth) is a crumbly brownish-grey (2.5 Y 4//2 to 10 YR 4//2) clay deposit, resulting from (sub)recent overbank sedimentation during wet season floods, and including the modern soil. Root penetration is evident through the upper 3 m of the core. Layers of volcanic ash are evident at 300 cm, 530 cm, 772 cm and possibly 1250 cm depth.

Six conventional radiocarbon dates on bulk sediment are available for the RD-3 sequence (Table 2). Dates were calibrated to calendar years using the program CALIB v.3.0 (Stuiver and Reimer, 1993). Unless otherwise stated, ages discussed below will be expressed as the median age of the two sigma calibrated age range, taken to be the most probable age of the sample. The age/depth relationship (Fig. 4) indicates a coherent sequence of increasing age with depth, and relatively stable sedimentation rates through the sequence, with exception of the upper 2.5 m.

Samples (1 cm$^3$) for diatom analysis (64) were taken at 20 cm intervals. Samples were digested in 10% hydrogen peroxide ($\text{H}_2\text{O}_2$) and 10% hydrochloric acid (HCl). The remaining material was diluted by a measured amount of distilled water and mounted with Naphrax. Counts were undertaken on Olympus BH-2
Fig. 4. Age-depth relation for Rawa Danau core RD-3.

Fig. 5. Areola density measurements for the main forms, variants and species of *Aulacoseira* recorded in the Rawa Danau record. All measurements were made on a Zeiss Axioskop with DIC at ×1000 magnification (Plan-Achromat 100×/1.40 oil DIC objective).
Rawa Danau, West-Java, Indonesia
core RD-3

Fig. 6: Diatom diagram for Rawa Danau core RD-3, showing taxa that account for at least 15% of the diatom variation.
Rawa Danau, West-Java, Indonesia
core RD-3

Fig. 8. Pollen diagram for Rawa Danau core RD-3, showing relative frequency curves for individual taxa, shading indicates 5 x exaggeration, a summary diagram, charcoal/pollen ratio, non-siliceous algae estimates and abundance of Ceratophyllum leaf-spines.
and Zeiss Axioskop microscopes with Differential Interference Contrast. Diatom taxonomy followed Krammer and Lange-Bertalot (1986, 1988, 1991), and all taxa were identified to at least species level. *Aulacoseira granulata* valves were identified to either recognised subspecies (i.e. *Aulacoseira granulata* var. *angustissima*), or to a morphotype based on the density of areolae on the valve girdle. Two forms of *A. granulata* are distinguished in this manner (Gómez et al., 1995; Krammer and Lange-Bertalot, 1991; Hustedt, 1930, 1938/39; Müller, 1903). The least coarse form recorded in the Rawa Danau material (*A. granulata* γ or ‘gamma form’) has a mean areolae density of 11.1/10 μm, whereas the coarse form (*A. granulata* or ‘alpha form’) has a mean areolae density of 6.27/10 μm. The mean and range of the areolae density of the four most common forms, variants and species of *Aulacoseira* recorded in this study are presented in Fig. 5. This procedure was undertaken as variations in valve ornamentation between morphotypes of *Aulacoseira* are related to environmental conditions (Gómez et al., 1995; Kilham et al., 1986; Stoermer et al., 1985). A minimum of 300 diatoms were counted per sample. Results are expressed as relative abundances of the total diatom community (excluding aerophilous taxa). Only taxa that occur above 15% in at least one sample are included in Fig. 6, although the habitat summary diagram (Fig. 7) includes values for all taxa.

Samples for pollen and charcoal analysis (42) were selected at 40 cm depth intervals. From every selected sub-sample 2 cm² of sediment was processed. Samples were initially treated with hot 10% Na-pyrophosphate (Na₃P₂O₇) and sieved over a 160 and an 8 μm mesh. The material retained in the 8 μm mesh was then treated with 10% hydrochloric acid (HCl) prior to acetylation (9 parts (CH₃CO)₂O:1 part H₂SO₄, 10 min). Organic material was isolated from the remaining inorganic fraction using heavy liquid separation with sodium polytungstate (Na₄[H₂W₁₂O₄₆] H₂O; s.g. 2.0, 20 min at 2000 rpm). The organic fraction was then treated with 40% hydrofluoric acid (HF) to remove any remaining silica. Samples were then dehydrated with ethanol (C₂H₅OH). Slides were mounted with glycerol (CH₂OHCHOHCH₂OH) and sealed with paraffin.

All slides were counted along evenly spaced transects using an Olympus BH-2 microscope at ×600 magnification. The minimum pollen sum was approximately 200 arboreal pollen grains. Pollen taxa from the open swamp vegetation (mainly Cyperaceae and Poaceae) were calculated as a percentage of the total arboreal and non-arboreal sum. The abundance of *Ceratophyllum* leaf-spines is also expressed as a percentage of this total arboreal and non-arboreal pollen sum, whereas non-siliceous algae estimates are based on a subjective ‘rare’ ‘present’, ‘common’, ‘abundant’ classification. Charcoal particles >10 μm in size were counted along three evenly spaced transects. Charcoal particle values per unit volume were calculated using the dilution of a known number of *Lycopodium* spores added to each sample prior to chemical processing.

5. Palaeoecological results

5.1. Diatoms

A total of 184 diatom taxa were recorded. Stratigraphically constrained classification (Grimm, 1987) of the results identifies eight sample groups which are used to define zone boundaries and provide a framework for diagram description and interpretation (Fig. 6).

5.1.1. Zone 1: 1700–1600 cm depth (ca. 16,300–15,000 years B.P.)

The diatom flora of this zone is dominated by the euplanktonic taxa *Aulacoseira granulata* var. *angustissima* and *A. granulata* γ, and the tychoplanktonic Fragilaria pinnata and *F. construens*. *Aulacoseira muzzanensis* and *Nitzschia paleacea* occur commonly also, while *Aulacoseira aff. alpigena* is abundant in the 1675 cm sample. The diatom flora suggests the occurrence of a fresh, alkaline, productive lake within the caldera. The black organic clay (Unit 1) deposited at this time is strongly suggestive of an anoxic depositional environment which in turn suggests poor circulation and a photic zone, which does not penetrate to the lake bottom. We interpret this as reflecting a deep water phase. However, the presence of non-planktonic diatoms appears to contradict this interpretation, being more indicative of a relatively shallow lake environment.
5.1.2. Zone 2: 1520–1220 cm depth (ca. 14,500–10,900 years B.P.)

Sediment between 1620 cm (approximately 15,000 years B.P.) and 1520 cm (approximately 14,500 years B.P.), coincident with a sedimentological change to green lacustrine clays, was not recovered during coring. The planktonic diatom *Aulacoseira granulata* is the dominant species above this missing section, with *Fragilariopsis* spp., *Nitzschia amphibia*, *N. paleacea*, *N. subacicularis* and *Synedra acus* v. *angustissima* occurring as common minor elements. We interpret this flora as indicating an open water, relatively deep, possibly fluctuating lake at the core site. The dominance of the meso-eutrophic species *Aulacoseira granulata* indicates relative productivity within the lake.

5.1.3. Zone 3: 1220–1120 cm depth (ca. 10,900–9600 years B.P.)

This zone was identified largely on the strong representation of *Aulacoseira granulata* var. *angustissima*, which reaches a maximum of 70% of the total diatom assemblage, replacing the *γ* form as the dominant planktonic diatom. *A. granulata* *γ* maintains a mean representation of approximately 20% throughout the zone, with *Fragilariopsis construens* and *F. pinnata* both reaching mean values less than 12% relative abundance. Minor taxa include *Nitzschia subacicularis* and *Synedra acus* var. *angustissima*.

5.1.4. Zone 4: 1120–900 cm depth (ca. 9,600–6700 years B.P.)

A second major shift within the phytoplankton occurs at approximately 9600 years B.P., with *Aulacoseira granulata* var. *angustissima* replaced by *Aulacoseira granulata* *α* and *A. muzzanensis*, reflecting a shift toward larger and more coarsely areolate diatoms.

*Aulacoseira* aff. *alpigena* is well represented between 995 and 955 cm (approximately 8000 and 7500 years B.P.), but is absent from other samples in the zone. *Fragilariopsis construens* and *F. pinnata* have maximum values of 15 and 14% respectively at 995 cm, with mean values of < 5.5% in the zone. Other taxa recorded include *Nitzschia gracilis*, *N. lacuum*, *N. paleacea*, *N. subacicularis*, and *Synedra acus* var. *angustissima*.

5.1.5. Zone 5: 900–720 cm depth (ca. 6700–5400 years B.P.)

A marked increase in tychoplanktonic species *Fragilariopsis construens* and *Fragilariopsis pinnata*, at the expense of previously dominant euplanktonic species, indicates substantial hydrological changes at the core site from approximately 6700 years B.P. The increased representation of tychoplanktonic taxa — that is, taxa that spend part of their life cycle on the benthos — suggests that changes in habitat, notably lake shallowing, competitively advantaged *Fragilariopsis*. Importantly, the absence of obligate benthic taxa within the diatom assemblage suggests that photic depth may have remained too shallow to support such taxa, with tychoplanktonic taxa thus competitively advantaged.

The genus *Aulacoseira*, which had dominated the record for over 9000 years, is substantially reduced at this time, presumably as a result of increasing competition with tychoplanktonic species for light and nutrients. The representation of *A. granulata* *α* form falls to less than 10% of the assemblage, while *A. muzzanensis* is largely absent from the diatom assemblage. *A. granulata* *γ* form is the most strongly represented species within the genus, reaching a maximum representation of 82% at 895 cm, before falling to a mean representation of approximately 23% through the rest of the zone. *Nitzschia subacicularis* and *Synedra acus* var. *angustissima* have variable representation through the zone, both reaching maximum values of 18%.

5.1.6. Zone 6: 720–390 cm depth (ca. 5400–2000 years B.P.)

*Fragilariopsis construens* and *F. pinnata* remain the dominant species within zone 6, with a mean representation of 29 and 21% of the total diatom sum, respectively. The representation of *Aulacoseira* within this zone is extremely variable. The most common species, *A. granulata* *γ* form, reaches 53% of the total diatom assemblage at 625 cm (ca. 5000 years B.P.), but has a mean representation of only 10% through the zone. The epiphytic diatoms *Achnanthes minutissima* and *Cocconeis placenta* increase in abundance between approximately 5000–4000 years B.P., the latter reaching in excess of 30% of the diatom sum.
5.1.7. Zone 7: 390–275 cm depth (ca. 2000–500 years B.P.)

Periphytic taxa dominate the diatom assemblage between 375 and 335 cm depth (a mean value of 80%). They include *Achnanthes levanderi*, *A. minutissima*, *Cocconeis placentula*, *Epithemia adnata*, and *Eunotia bilinaris*. Diatoms with an aerophilous habit, such as *Navicula confervacea*, increase to around 8%. In contrast, both planktonic and tychoplanktonic taxa are poorly represented in this zone (both with a mean value of around 6%). We interpret the dominance of attached diatoms (Fig. 7) as evidence for a further decrease in lake depth, allowing the temporary expansion of aquatic and littoral plants close to the core site.
Tychoplanktonic taxa increase markedly in the upper two samples of this zone (315 and 285 cm depth), concomitant with a decrease in the relative abundance of periphytic and aerophilous species, possibly reflecting a return to deeper water conditions from around 870 years B.P.

5.1.8. Zone 8: 125–50 cm depth (ca. 181–80 years B.P.)

Diatom frustules did not preserve between 255 and 125 cm depth (ca. 370–180 years B.P.). After this period periphytic and aerophilous taxa dominate the record, suggesting relatively shallow, possibly fluctuating lake levels at the core site.

No diatoms are preserved in the upper 50 cm of the record. This is most likely due to the lowering of the outflow level early this century, substantially reducing the size of the permanent lake and thus initiating pedogenesis at the core site. Conditions are likely to have become unsuitable for both diatom growth and preservation at this time.

5.2. Palynology

The pollen diagram (Fig. 8) has been divided into four pollen zones, based on major changes in the relative abundance of taxa in the diagram.

5.2.1. Zone 4: 1700–1425 cm depth (ca. 16,300–14,500 years B.P.)

Zone 4 is initially characterised by high Poaceae and charcoal particle values, and low Pteridophyta values. Poaceae pollen decreases substantially through the zone, falling from over 60% of the non-arboreal pollen sum to less than 15%. Macaranga-Mallotus and Moraceae-Urticaeae (probably Athrocarpus) pollen dominate the arboreal pollen sum, with Elaeocarpus, Celtis, Gironniera-Trema, and Ficus strongly represented also. Non-siliceous algae, particularly Coelastrum and Pediastrum, are common at this time.

5.2.2. Zone 3: 1425–1210 cm depth (ca. 14,500–10,800 years B.P.)

The relative abundance of Poaceae stabilises in this zone, while arboreal pollen types increase in both taxon diversity and relative abundance. Elaeocarpus and Moraceae-Urticaeae are the most strongly represented arboreal taxa, with Carallia-Rhizophora, Celtis, Dipterocarpaceae, Gironniera-Trema, and Macaranga-Mallotus pollen well represented. Charcoal particle values decline initially, with a return to higher values in the upper samples. The non-siliceous algae Botryococcus and Coelastrum are both abundant in these sediments.

5.2.3. Zone 2: 1210–460 cm depth (ca. 10,800–3400 years B.P.)

Macaranga-Mallotus pollen dominates the arboreal pollen count throughout this zone, progressively increasing to a maximum of 44% at 500 cm depth. In contrast, Carallia-Rhizophora type declines through the zone from a maximum relative abundance at 1200 and 1160 cm depth (ca. 10,600 years B.P.). Elaeocarpus pollen increases to 26% of the arboreal pollen sum at 487 cm depth (ca. 4100 years B.P.). Antidesma-Baccaurea type, Celtis, Dipterocarpaceae, Ficus, Gironniera-Trema, Moraceae-Urticaeae and Nauclea type are all consistently represented within the arboreal pollen assemblage throughout the zone. Poaceae and charcoal values are low, while Pteridophyta values and non-siliceous algae are abundant. Leaf-spines of the floating aquatic plant Ceratophyllum (Hydrocharitaceae) are common, particularly in the upper samples of this zone, becoming most abundant at 607 cm depth (ca. 4900 years B.P.).

5.2.4. Zone 1: 460–50 cm depth (ca. 3400–80 years B.P.)

Elaeocarpus is strongly represented in this zone, replacing Macaranga-Mallotus as the dominant arboreal pollen type. Eugenia, Ilex and Ficus are also common. From around 400 years B.P. the arboreal pollen signal is greatly reduced, replaced largely by Asteraceae Tubuliflorae, Cyperaceae and Poaceae. This is broadly synchronous with an increase in charcoal values. Non-siliceous algae are virtually absent from these sediments. From around 250–300 years B.P., pollen of the cultivars Arenga (sugar palm), and Cocos nucifera type (coconut) become increasingly important, as do Selaginella spores.
6. Discussion

6.1. Palaeoenvironmental reconstruction and climate history

6.1.1. Late Glacial (ca. 16,300–11,250 years B.P.)

The palynological data for the Late Glacial at Rawa Danau indicate the presence of open herbaceous swamp vegetation dominated by grasses, while the character of the sediment deposited at this time suggests a deep lake. The diatom communities indicate a slightly alkaline and productive lake but, in contrast with the sedimentological evidence, the presence of periphytic and aerophilous diatom taxa suggests shallower conditions. Pteridophyte were not a major component of the vegetation, which indicates relatively dry conditions. This interpretation is supported by the charcoal record, which indicates that fires were relatively frequent.

The majority of the pollen deposited at the core site is likely to have been derived from local swamp forest communities and dryland forests within the caldera. Indeed, there is little evidence of any long-distance pollen rain. Exceptions to this are high altitude elements such as Daucycarpos, Distyllium, Podocarpus, and possibly Ulmus, although these taxa never reach high values, and are largely restricted to this late glacial period (before 11,250 years B.P.). The swamp forest which occurred within the caldera during the Late Glacial was probably restricted to the margins of the lake proper, and was largely composed of Moraceae-Urticaceae (mostly Artocarpus), Macaranga-Mallotus, Elaeocarpus (probably E. littoralis) and Ficus (probably F. retusa).

From around 15,000–13,000 years B.P., grass cover gradually reduced while Elaeocarpus forest expanded. The sedimentological change to greenish, highly organic clays, abundant mollusc fauna and planktonic, meso to eutrophic diatom taxa indicate that Rawa Danau was a productive, neutral to slightly alkaline fresh-water lake from ca. 14,500 years B.P. The total forest cover in the Rawa Danau caldera increased strongly from around 15,000 years B.P., indicating a change to wetter climatic conditions. Ferns become an important element of the vegetation from around 11,250 years B.P., also indicating the development of a more humid environment. Macaranga-Mallotus and Dipterocarpaceae seem to have dominated the swamp forest communities, partly replacing Moraceae-Urticaceae and Elaeocarpus, while planktonic diatoms are at their most abundant.

The presence of Syedra acus var. angustissima and Aulacoseira aff. alpigena in these late glacial sediments is noteworthy. In tropical African lakes Kilham et al. (1986) have shown that elongate taxa from the Fragilariaeae, including the Syedra acus group, are competitively advantaged at high Si:P ratios relative to a number of other planktonic genera, including Aulacoseira species. The relatively strong occurrence of this species at Rawa Danau may therefore reflect relatively high concentrations of dissolved silica in the lake. Aulacoseira aff. alpigena (which in Australian systems has similar nutrient requirements to A. granulata, Tibby unpublished data) may be more abundant as a result of competitive advantage offered by smaller cell size and slower sinking rates. In this context A. aff. alpigena’s greater abundance may reflect factors such as longer periods of stratification.

6.1.2. Early Holocene (ca. 11,250–6700 years B.P.)

The presence of Carallia-Rhizophora pollen in the Rawa Danau record during the early Holocene could be ascribed to the presence of Carallia in the swamp forest vegetation (probably C. lucida; Endert, 1932). However, the shape of the curve and its timing suggests that it may actually reflect the development of mangrove vegetation in the Sunda Strait area, through long-distance transport of Rhizophora pollen. Many of the marine palynological studies in the region show a similar pattern, with the maximum expansion of Rhizophora around the transition from oxygen isotope stage 2 to 1 (Van der Kaars, 1991, 1998; Van der Kaars et al., 2000). The presence of Sonneratia within the pollen record at this time provides some corroborative evidence for this interpretation.

Lake water chemistry during the early Holocene appears largely unchanged from the late glacial period; that is, fresh, slightly alkaline and most probably meso to eutrophic. The shift toward coarser forms of Aulacoseira is difficult to interpret unequivocally with the data available. Kilham et al. (1986) argue that shifts between small pored forms of Aulacoseira granulata (e.g. γ form) and coarse forms (e.g. α form) are likely to result from increases in lake nutrient status. Gómez et al. (1995) also indicate that α form occurs in waters with higher nutrient
status. Their data indicate that, when contrasted to the γ form, α form occurs in waters with higher dissolved Si concentrations. It seems, therefore, that the broad shifts from *Aulacoseira* valves with high areolae density to those which are more coarsely areolate in the Rawa Danau record may reflect lake shallowing, sediment entrainment and a concomitant increase in nutrient availability.

6.1.3. Middle Holocene (ca. 6700–3650 years B.P.)

Vegetation within the Rawa Danau caldera appears to have been relatively stable through the middle Holocene. Mixed swamp forest communities were dominated by *Macaranga-Mallotus*, with *Celitis*, Dipterocarpaceae, *Elaeocarpus*, Moraceae-Urticaceae (*Ficus*) and *Trema* as minor elements. The strong representation of fern spores at this time, in concert with low charcoal values, implies a relatively stable humid climate. Interestingly, a sharp increase in representation of the epiphytic diatom *Cocconeis placentula* from ca. 5000 to 4500 years B.P. (625–545 cm depth) is associated with the maximum representation of *Ceratophyllum* leaf-spines recorded in the pollen samples. While the pollen of *Ceratophyllum* is not produced in large numbers and does not preserve readily (Warner, 1989), Birks (1973) has shown *Ceratophyllum* macrofossil abundance is concomitant with the plant’s representation in lakes. Given that macrofossils are generally not found in abundance at a great distance from the source plant (Birks, 1973), the abundance of *Ceratophyllum* leaf-spines, in concert with the increased representation of epiphytic diatoms implies the presence of floating and submerged aquatic plants at the core site. The sudden decline of *Ceratophyllum* and the epiphytic diatoms *Cocconeis placentula* and *Achnanthes minutissima* at ca. 4500 years B.P. is coeval with the deposition of volcanic ash (530 cm depth). Gasse (1986) found *C. placentula* growing abundantly on *Ceratophyllum*, and there appears to be a similarly strong association between this diatom and *Ceratophyllum* in the Rawa Danau record. We interpret the reduction of *Ceratophyllum* to be a result of volcanic ash fall, perhaps representative of a larger ecological impact, which is not visible in the pollen record, and that the dramatic reduction of epiphytic diatom taxa from the record reflects the destruction of this habitat.

Interestingly, the 530 cm depth ash layer corresponds to the decline in *Syndra acus* var. *angustissima*, a taxon apparently adapted to high Si:P ratios. Given the high Si content of volcanic tephra (Eastwood et al., 1998), this decline is intriguing. Abella (1988) notes a similar phenomenon with the decline of *Fragilaria crotonensis* in response to a Mt. Mazama tephra deposition on Lake Washington. Changes in lake-water pH may be associated with tephra deposition (Birks and Lotter, 1994), while short-term shading of phytoplankton is another possible explanation for elimination of *Syndra acus* var. *angustissima*.

The change in the diatom flora from euplanktonic to tychoplanktonic species at ca. 6700 years B.P. reflects a critical ecological threshold within the diatom community. Metcalfe (1988) notes that habitat characteristics are important in determining species composition in tropical freshwater systems, hence the change from an *Aulacoseira* dominated community to one dominated by *Fragilaria* may be a result of competitive advantage provided by tychoplanktonic life strategy. Bennion (1995) has shown that *Fragilaria* species, including those recorded in the Rawa Danau sediments, may dominate as a result of the penetration of the photic zone to the lake bottom which they inhabit. However, these taxa also dominate the plankton in highly turbid, shallow lakes with a very low secchi depth (<50 cm) (Fluin, unpublished data). In such shallow, turbid environments these taxa may be advantaged by their chain forming habit, which allows them to be suspended into the photic zone, but also inhabit nutrient-enriched environments in the lake bottom. Whether light penetrated to the bottom sediments at Rawa Danau from ca. 6700 years B.P. or not, it appears that the transition to *Fragilaria construens* and *F. pinnata* dominated assemblages is related to competitive advantage afforded by lake shallowing as a result of vertical sediment accretion.

6.1.4. Late Holocene (ca. 3650–80 years B.P.)

The initiation of peat formation, the disappearance of non-siliceous algae and the replacement of the previously common *Macaranga-Mallotus* and Dipterocarpaceae forest with *Elaeocarpus* forest, occurred from ca. 3650 years B.P., in response to lake shallowing. While the increased representation of periphytic and aerophilous taxa in this part of the sequence may reflect further shallowing at the core site, there is no
evidence for oxidation or pedogenesis. Alternatively, these taxa may be occurring above the water level on the trunks and buttressed roots of swamp forest trees occurring at the core site. The localised occurrence of swamp forest vegetation is confirmed by the contemporary deposition of woody peat and abundant leaf macrofossils. From ca. 1750 years B.P. *Ilex* became an important element of the swamp vegetation, while the appearance of aerophilous diatoms such as *Navicula confervacea* and *N. contenta*, associated with fluctuating water levels (Vyverman, 1992) in the record are indicative of further terrestrialisation of the core site. During the last few hundred years, as a result of a marked intensification of land use (burning and crop cultivation), and the lowering of the outflow level, the vegetation in the Rawa Danau area became much more open. Forest cover had been reduced and herbs such as Asteraceae, Cyperaceae, Poaceae and the fern *Selaginella* became important elements in the vegetation, as did Anacardiaceae, *Arenga*, and *Cocos nucifera*. The remaining swamp forest vegetation included *Elaeocarpus*, *Glochidion*, *Ilex*, *Macaranga-Mallotus*, *Stenomurus*, and ferns (*Cytisere* and undifferentiated psilamonolete spores), a community structure similar to the closed, fern-rich swamp forest described by Endert (1932) in the north of the caldera.

6.2. Regional comparisons and local environmental change

The pollen, diatom and sediment records presented here present a largely coherent reconstruction of palaeolimnological and palaeoclimatic conditions from the lowlands of Indonesia from the Late Glacial to the present. In particular, hydrological changes, documented by changes in the diatom community, have had a direct impact upon swamp forest vegetation in the Rawa Danau caldera, as clearly reflected in the pollen record.

The only published record from the Indonesian lowlands that extends beyond the Holocene is the Bandung Basin core, West Java (Van der Kaars and Dam, 1995, 1997), which has provided a 135,000 year palynological record. However, the Late Glacial and Holocene are not well represented in the Bandung Basin record, and meaningful comparisons with the much younger Rawa Danau record cannot be made.

The present study indicates that, during the Late Glacial period, the vegetation at Rawa Danau was much more open, suggesting a relatively dry climate. This contrasts strongly with contemporary records from higher altitudes in Java. Stuijts (1993), for example, records the dominance of ever-wet *Dacrycarpus* forest in West Java until 12,000 years B.P. from a series of sites between 1015 and 2200 m asl, inferring precipitation values comparable with today’s but temperatures at least 3°C lower than present. Temperatures are thought to have increased between 12,000 and 10,000 years B.P., evidenced by a rise in *Engelhardia* pollen and the subsequent development of Fagaceous forests during the Holocene (Stuijts, 1993). Similarly, palaeoclimatic evidence from central (Newsome and Flenley, 1988) and northern Sumatra (Maloney, 1979; Maloney and McCormac, 1996), suggests that temperatures during the most recent glacial period were lower than present (up to 5.2°C lower; Newsome and Flenley, 1988) but, at least in northern Sumatra, conditions may have been as wet or possibly wetter than present.

By contrast, there is little unequivocal palynological evidence of depressed temperatures at Rawa Danau during the late glacial period. The presence of high altitude elements prior to 11,250 years B.P., possibly derived from the nearby Mt. Karang area, may be taken as evidence of the expansion of mid-upper montane forest elements into lower altitudes (reflecting lower temperatures), but might equally be explained by an increased influx of long-distance pollen rain during periods when vegetation cover was relatively open and/or when there was a large area of open water within the caldera. In any case, little can be deduced from the presence of these montane elements at Rawa Danau given their poor representation within the pollen spectra. The data presented here are, then, more in keeping with climate reconstructions based on marine cores which indicate little variation from modern values in sea surface temperatures during the most recent glacial maximum (Thunell et al., 1994), than with terrestrial records of climate change from high altitude sites which demonstrate dramatic temperature change over this period. Similarly, Kutzbach and Guetter (1986) predict little change in temperature and only minor declines in precipitation in the West Java region during the glacial maximum based on synthetic climate modelling experiments. Although the RD-3 core does not
incorporate the glacial maximum, our late-glacial data does not conflict with Kutzbach and Guetter’s (1986) temperature model.

The palaeoecological data presented here are equivocal with regard to climatic conditions during the Holocene, as vegetation development appears to be largely controlled by local hydrological changes within the caldera. However, palynological data, particularly the strong representation of fern spores, do suggest that climatic conditions were relatively humid and apparently stable through the early and middle Holocene.

While the pollen data provide some evidence of palaeoprecipitation change over time, there is little convincing evidence that biotic or physico-chemical changes in the lake itself have responded to, or been forced by, changes in prevailing climate. Throughout most of the 16,000 year record presented here, Rawa Danau was a fresh, alkaline, meso-eutrophic water body. The overriding mechanism forcing change in the algal community within the lake is clearly water depth and associated physico-chemical variations. Bradbury and Diterich-Rurup (1993) have shown that shifts between planktonic taxa, such as the shift from *Aulacoseira* to *Synedra* in the Rawa Danau record, may reflect short-term climatic fluctuations (and their effect on nutrient availability), superimposed upon a diatom flora which is largely unresponsive to longer term climatic changes. The role Si availability and limitation has played in the development of the diatom communities in this volcanically active region is uncertain. It is only with the decline of *Cocconeis placentula* at ca. 4500 years B.P. that any direct response by the diatom community to a recorded ash fall event can be observed.

While the diatom data provide a striking record of lake depth variation from the late glacial period, lake depth decreases are not manifestly engendered by climate change. It is our interpretation that the lake level data presented here reflect the sedimentological and ecological process of basin in-filling, or terrestrialisation, rather than direct response to variations in the precipitation/evaporation balance associated with climate change since the Late Pleistocene. The geological and geomorphological setting of the lake, which has maintained deep lake conditions even during periods of relative aridity, has apparently isolated the lake from the environmental changes that were clearly taking place within the caldera and in the surrounding landscape.

6.3. Human/environment interaction

People clearly had a large impact on the vegetation development in the Rawa Danau area during the last few hundred years, as evidenced by increased burning, the appearance of food crops, the apparent opening of the vegetation with many grasses, sedges and ferns, and presence of weeds. However, it is more difficult to identify the impact of people, as distinct from climate change, in earlier parts of the record. The vegetation in the Rawa Danau caldera during the Late Glacial period was relatively open, much like the vegetation of the last few hundred years, although with a significantly different character; that is, dominated by grasses yet with very few of the ferns and sedges, that are presently a characteristic element of the open vegetation in the caldera (Endert, 1932).

Burning levels were high during the Late Glacial, but there is no evidence for food crops or the presence of weeds. Our record does not enable us to establish whether the open vegetation around the lake was caused by people through forest clearance and burning, or whether people moved on to the lake shore area after the vegetation had become more open. Although resources around the lake may have been attractive to people, they may only have moved there once ease of access had increased, such as when dense forest cover had changed to open vegetation at the onset of the last glacial period. For instance, charcoal and pollen evidence from the Bandung basin record suggests that people may have moved into the lake shore area around 65,000 years B.P., but only after dense forest cover had been replaced by a much more open vegetation at around 80,000 years B.P. (Kershaw et al., 1997; Van der Kaars and Dam, 1995).

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