The Eocene–Oligocene transition in Nanggulan, Java: lithostratigraphy, biostratigraphy and foraminiferal stable isotopes

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Supplementary material: Supplementary figures, tables and appendices are available at https://doi.org/10.6084/m9.figshare.c.5429453

Received 13 January 2021; revised 8 May 2021; accepted 18 May 2021

Abstract: The Nanggulan section in south central Java comprises open marine sediments and volcanic deposits of Eocene–Oligocene age that accumulated in a marginal basin within the young Sunda Arc complex. A new borehole captures the stratigraphy and showcases the exceptional preservation of calcareous microfossils across an apparently complete Eocene–Oligocene Transition (EOT), a time interval significant for the initiation of continental-scale glaciation on Antarctica. Low-resolution benthic and planktonic foraminifera oxygen and carbon stable isotopes (δ18O and δ13C) record increasing δ18O and δ13C in the basal Oligocene, allowing correlation to global records. Isotopic values imply warm temperatures and relatively high nutrients along the SE Java margin. The Nanggulan EOT is a valuable archive for reconstructing ocean–climate behaviour and plankton evolution and extinction in the Indo-Pacific Warm Pool. The borehole also adds to understanding of the early stages of Sunda Arc volcanism.

Miocene and younger volcanic rocks dominate much of the surface geology of the Indonesian island of Java. However, in the Nanggulan region of central Java, patches of calcareous marine and volcanoclastic sediments of Paleogene age crop out in small pockets where there has been limited burial following by uplift and erosion (Oppenoorth and Gerth 1929; Hartono 1969; Lunt 2013a). These calcareous sediments, which comprise chalky marls, claystones and clays, and contain diverse and in places very well-preserved ‘glassy’ microfossils of middle Eocene to late Oligocene age, are renowned to palaeontologists (Verbeek 1891; Rutten 1927; Hartono 1969; Saito et al. 1981; Pearson et al. 2006; Lunt 2013b). Of particular interest is the presence of a sedimentary sequence spanning the Eocene–Oligocene Transition (EOT), an interval widely recognized as the time when Earth’s climate switched from a globally warm greenhouse state to an icehouse ‘coolhouse’ state, and continental-scale glaciation was initiated on Antarctica (Shackleton and Kennett 1975; Prothero 1994; Zachos et al. 1998; Coxall and Pearson 2006; Francis et al. 2008; Westerhold et al. 2020). High-latitude studies are important for determining the effect and scale of EOT climate change but, equally, conditions in the tropics, especially sea surface temperatures (SSTs), are needed to constrain the global impact of Antarctic glaciation and test hypotheses about the drivers of change (Hutchinson et al. 2021). If declining levels of CO2 were the dominant driver, as modelling studies predict (DeConto and Pollard 2003; Goldner et al. 2014), we should see cooling in the tropics as well as the high latitudes. In contrast, if Southern Ocean tectonic changes, that is, opening of Drake Passage, causing thermal isolation of Antarctica, forced the shift (Kennett 1977), then we might expect a latitudinal or global redistribution of heat (Hutchinson et al. 2019). Furthermore, constraining how hot tropical regions became under strong greenhouse climate conditions is relevant to concerns about modern global warming. A continuous cored section through the upper Eocene–lower Oligocene part of the Nanggulan Section has potential to provide new insights into these questions from the tropics.

In this paper we describe the lithology, biostratigraphy and some initial foraminiferal stable isotope results from scientific coring of the Eocene–Oligocene sedimentary succession in the Nanggulan region. The recovered cores provide the first continuous sedimentary sequence spanning the EOT from the Indo-Pacific Warm Pool, thus sampling global change in a climatically revealing part of the globe. Additionally, occurrence of volcanic deposits in the cored section, which are well dated by biostratigraphy, provide new insights into early stages of arc volcanism in the ‘young’ and well-preserved Sunda Arc province. The coring was conducted by an international team of researchers from UK and Irish universities and the Indonesian Geological Research and Development Centre (GRDC), for stratigraphic and palaeoclimate studies.

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Interest in the Nanggulan Paleogene past and present

There is a long history of interest in the geology of the Nanggulan region because of mineral and hydrocarbon exploration (van Dijk 1872; Martin 1882; Verbeek 1891; Martin 1915; Oppenoorth and Gerth 1929; Leupold and van der Vlerk 1932; Rothpletz 1944; van Bemmelen 1949; Hartono 1969; Okada 1981; Purnamaningsih and Harsono 1981; van Gorsel et al. 2014; Hartono and Sudradjat 2017). Much of this work remains outside the peer-reviewed literature (Oppenoorth and Gerth 1929; Lelono 2000). Previous investigations include extensive surface sampling, drilling of boreholes in 1864–1865, 1969 and 1998, and trenching across the Eocene–Oligocene boundary (E/O) (see review by Lunt 2013a). The rich microfossil and mollusc assemblages are renowned for the high quality of preservation and stratigraphic utility, with already proven suitability for geochemical proxy reconstructions from outcrop samples (Evans et al. 2018). A number of foraminiferal taxa have been described from the Nanggulan section, including Hantkenina nanggulanensis Hartono and various species of Nummulites e.g. N. nangoelina Verbeek and N. djokdjokartae (Martin), the latter being the subject of sophisticated 3D morphometric analysis (Renema and Cotton 2015). Moreover, the Nanggulan Formation has played an important role in the development of the Indo-Pacific marine biostratigraphy and may be regarded as the birthplace of the ‘East Indian Letter Classification’, which remains an objective division of time over interpreted chronostratigraphic timescales (Leupold and van der Vlerk 1932; Lunt 2013b). Most recently, the taxonomy of Nanggulan calcareous nannofossils from the cores described herein has been investigated in detail and their assemblages used as indicators of oceanic change (Jones et al. 2019; Jones and Dunkley Jones 2020).

Geological and oceanic setting

Java is situated at the convergence of the Eurasian and Indo-Australian plates in the Indonesian archipelago and is part of the island chain forming the Sunda Magmatic Arc (Fig. 1). The island sits on the SE edge of the Eurasian plate, locally known as the Sunda Plate or Sundaland (Fig. 1) (Smyth et al. 2008). The Indo-Australian plate, which has been moving north since initial rifting and separation from Antarctica c. 100 myr ago, has a margin of oceanic crust that has been subducting along the modern Sunda–Java Trench since the middle Eocene (c. 45 Ma), from which time the first evidence of arc volcanism appears on Java (Smyth et al. 2005, 2008). The modern volcanic arc, active since the late Miocene (Soeria-Atmadja et al. 1994, 1998), is expressed as a chain of more than 30 NW–SE-trending volcanoes. The string of topographic highs paralleling this to the south, known as the Menoreh Hills, also the West Progo Mountains, Kulon Progo Mountains or the Southern Arc Mountains, are the remnants of mostly Oligocene to early Miocene-age volcanoes that formed an earlier arc complex (Smyth et al. 2008; Widagdo et al. 2018) (Fig. 1). These rocks are referred to as the ‘Old Andesites’ and are represented by major, kilometre-thick volcanic deposits lasting until about 20 Ma (van Bemmelen 1949; Lunt 2013a).

Sedimentary basins on Java include remnants from the Paleocene, with deep marine clastic deposits incorporating
carbonate olistoliths around Karangsambung (Lunt 2013a). Most basins developed in the later middle Eocene as the result of strike-slip faulting and extension of the Makassar Straits (Lunt 2019). These basins developed in an episodic fashion with variable uplift and subsidence, while initially weak volcanism incorporated thin tuff beds into the sedimentary record. Eocene and early Oligocene strata contain many richly quartzose sandstones but by the late Oligocene Old Andesite volcanic material dominated, and quartz sands are not known in southern Java from this time onwards. A period in the later early and middle Miocene had fewer volcanic beds and many limestones developing on extinct volcanic highs, but from the early late Miocene, volcanism once again predominated over sediments on Java.

Eocene sediments occur in many wells and outcrops (Lunt 2013a) but the best-preserved material is in the Nanggulan region, situated NW of the regional capital Yogyakarta (Fig. 1). The Nanggulan Formation crops out on the eastern slope of the Menoreh Hills NW of the village of Kenteng. It comprises a c. 500 m thick transgressive sequence of coals and conglomerates, sandstones, siltstones, mudstones and chalk underlying extensive volcanic deposits. According to Lunt (2013a) the Oligocene intrusion of the three large volcanic plugs under the Menoreh Hills formed a buttress preventing further subsidence and burial of the Nanggulan area on its eastern flank.

From an ocean–climate perspective, Java today sits at the western edge of the Indo-Pacific Warm Pool, where SSTs are among the warmest on the planet (c. 29°C, De Deckker 2016) (Fig. 1). The Indonesian archipelago configuration restricts inflow from the Pacific, resulting in slightly cooler SSTs in the Indian Ocean compared with the western Pacific, where peak Warm Pool SSTs occur (Fig. 2). During Eocene–Oligocene time Australia was further south and model simulations show the Warm Pool extending unimpeded into the Tethys or proto-Indian Ocean (von der Heydt et al. 2006; Huber and Caballero 2011; Hutchinson et al. 2018) (Fig. 2). How hot and expanded this Warm Pool became under greenhouse climates of the Eocene is uncertain. Modelling predicts SSTs of the order of 38°C (Fig. 2), far exceeding modern limits (Huber and Caballero 2011; Hutchinson et al. 2018) but proxy data are lacking.

The high quality of calcareous microfossil preservation in the Nanggulan cores, which is similar to that previously described from Tanzania and elsewhere (Norris and Wilson 1998; Pearson et al. 2001), will allow quantitative proxy measurements of important climate variables across the EOT, including SST, thus filling this crucial gap in reconstructing Earth’s past temperature maxima for the warmhouse to coolhouse transition. The recent nannofossil studies on the new core include a brief outline of the borehole stratigraphy (Jones et al. 2019; Jones and Dunkley Jones 2020). Here we present an expanded treatment of the Nanggulan NKK-1 borehole lithostratigraphy, assessments of the planktonic and benthic foraminifera, an updated summary of the nannofossils and initial insights into the palaeoceanography using benthic and planktonic foraminiferal δ18O and δ13C.

Stratigraphy of the Nanggulan section from previous studies

Early studies of the Nanggulan Section utilized surface samples, often from outcrops in small stream sections, including Kali Watu Puru, Kali Songo and Kali Kunir, but several boreholes have also been drilled in the region previously (see reviews by Smyth et al. 2008 and Lunt 2013a). The first borehole, drilled by van Dijk in 1864–1865, was 161 m (van Dijk 1872). A second is referred to in a micropalaeontological study by Hartono (1969). This was drilled during construction of a bridge at Kebon Agung. However, there is no systematic report and no trace of the core at the GRDC repository where it was thought to be. The third drill core was taken by Lundin Blora b.v. through the lower part of the succession in 1998 close to the van Dijk site and reached 130 m subsurface depth (Lunt 2013a).

Various lithological units have been recognized in the Nanggulan section over time (Table 1). The lower beds contain terrestrial to marginal marine sediments, evidenced by abundant molluscs, larger benthic foraminifera and coals (‘Axinea’ and ‘Jogjakarta’ beds of Oppenwoorth and Gerth 1929 and Hartono 1969; Songo Beds of Lunt 2013a). Up-section, beds were found to contain diverse planktonic microfossils including planktonic foraminifera, radiolarians and calcareous nannofossils, together with deep-sea benthic foraminifera, implying deep water during the late middle Eocene to Oligocene (Lunt 2013a; van Gorsel et al. 2014). Known as the ‘Globigerina marls’, these deep marine sediments have been variously subdivided, most recently into the ‘Watu Puru Beds’, ‘Jetis Beds’ and the ‘Tegalsari Marls’ (Lunt 2013a). The Watu Puru Beds are the first strata in which volcanic material predominates over quartz, and are locally rich in planktonic foraminiferal marls. The Jetis Beds are poorly fossiliferous volcanioclastic turbidites. A thin (1 m thick) unit dominated by shallow water larger benthic foraminifera has also been identified in outcrop at the top of the Jetis beds close to our drill site, probably a debris flow from a thicker turbidite. Axinea chalky marl changing down-section into late Eocene clays (including the Tegalsari Marls and ‘cunialensis Clays’). The onset of this carbonate, and its distal debris flow equivalents in Nanggulan, identifies an unconformity very close to the middle to
late Eocene boundary, which is traced across the southern Makassar Straits and eastern Java (Lunt and van Gorsel 2013).

Methods

Drilling operations

Field surveying by P. Lunt identified an appropriate coring locality and in 2005 funds were obtained from the UK Natural Environment Research Council (NERC) to support shallow drilling. The expedition, which took place in January 2006 with support from Gadjah Mada University and the Indonesian Geological Research and Development Centre (GRDC), resulted in the first core to be drilled through the upper part of the Nanggulan succession. The borehole was cored on an open plot next to the Jetis Road at the top of the slope above the Kali Kunir Stream, a tributary of the western Kali Songo River (7.73844°S; 110.18574°E), providing the borehole name, NKK-1. Examination of a Google Earth Map at the time of writing (January 2021) indicates that this plot has now been built upon. During EOT time the NKK-1 site was closer to the Equator, with latitudinal reconstructions ranging from 6°S to 5°N depending on the rotation framework used. Here we use 2.155°N, 123.788°E as the late Eocene palaeo-coordinates after Hutchinson et al. (2018) and Jones et al. (2019).

Site NKK-1 was rotary cored using a ground-stabilized rig (Fig. 3), with a core diameter of 5 cm and typical core lengths of 1.2–1.5 m, using water and mud circulation to avoid unnecessary organic contamination. After recovery each core was laid in a plastic pipe split lengthways and cleaned of adhering drilling mud using a knife to expose the internal material for lithological description. Cores were described immediately after recovery. All samples for future analysis were taken in the field. For foraminiferal analysis, 10 cm long samples were taken to obtain sufficient specimens for palaeoceanographic studies. After sampling cores were wrapped in polythene, transferred to wooden core-boxes ready for archiving at the GRDC repository based at the Geological Museum in Bandung. To date (May 2021) P. Sanyoto (recently retired) and G. Marliyani confirm that the material is still accessible and available for study. Sample identification numbers include the site ID followed by core number and the position of the sample in centimetres from the top of Table 1.

| Oppenorth/Hartono 69 | Stratigraphy | Unit name/ approx. thickness | Age/ Stratigraphy | Lithology | Palaeontology | Environment | Map | NKK-1 drilling, this study |
|----------------------|--------------|------------------------------|------------------|-----------|---------------|-------------|-----|--------------------------|
| Andesitic breccias   | Old Andesites (many 100s of m) | late Olig. to early Miocene | Volcanic andesite breccias & agglomerates | NA | Deep marine | Agglomerates | Not drilled |
| Globigerina marls, (Hartono, 1969) | Tegalsari Marls (~1 m) | late Eocene/early & late Olig. P16-P22 E15-O6 CP15-19 | Planktonic-rich marls, clays & chalky marls. Condensed. Contains the E/O | Planktonic foraminifers. H. nanoguineensis described from upper Globigerina marls’ | Deep marine | Chalky marls | Chalky marls Unit I (~5.5 m recovered) |
| | “Palliatispina transgression” (seen at Kali Kunir in outcrop, 1 m) | late Eocene early to middle part of Letter Stage Tb | Bioclastic Palliatispina limestone’ | Dominated by larger benthic foraminifers (Palliatispina). | Deep marine expression of transgression. Down slope transport | Not seen | Not seen in the core |
| (Unit 3e of Oppenorth) | Jogjakartase Beds | middle Eocene P14 (E13) CP14 | Thinly bedded tuffaceous andesitic sandstones. | Planktonic foraminifers throughout, lower fossil content than older beds. | Deep marine. Distal high stand system. Slow deposition, air-fall &/or turbiditic volcanics | Olive grey marly clays & interbedded tuffaceous sandy clays | Unit VI Inter bedded clays & tuffaceous sands 11 m recovered, middle Eocene (E13 = ca. P14, N16|P17, ONE15) |
| | Water Puru Beds ~250 m | middle Eocene P12-P14 (E10-E13) ~ 3 m yrs of time CP14, N16 | Pale mudstone (calcareous marls), relatively low volcanic content | Excellent preserved planktonic foraminifers & radiolarians | Marine transgression. Deepening Stumps, provide shallow benthic inputs. | Marine activity from 4-2 Ma | ? | ? |
| | Discocyclina Beds (Zone 2e of Oppenorth) | middle Eocene P12-P14 (E10-E13) ~ 3 m yrs of time CP14, N16 | Arkosic, tuffaceous marine marls/clays with higher calcareous planktonic content than below & multiple (~ 5 cm thick) Discocyclina beds | Upper: planktonic foraminifers & radiolarians. Lower: abundant larger benthics Discocyclina, Astrodiscocyclina & planktonic foraminifers. | Not drilled | Not drilled |
| | Songo Beds ~200 m | middle Eocene P12 (E10) CP14 | Fine quartz sandstone/clayey-occasional conglomerates/coal layers, 200 m. | Rare Nummulites sp. | Delta plain. Aggrading clastic facies. Shallow marine. High sed. rate | Not drilled | Not drilled |

After Oppenorth and Gerth (1929), Hartono (1969) and Lunt (2013a). The NKK-1 core spudded into the upper Tegalsari Marls.

Table 1. Summary of the stratigraphic units recognized in the Nanggulan region based on previous coring, outcrop sampling and the results of this study, including coring and field mapping
the core; for example, Sample NKK-1/48, 98–108 cm was taken from NKK-1, Core 48, between 98 and 108 cm. Letter suffixes (e.g. a, b, c) refer to the part of the core (See Supplementary material, Appendix 2, core photographs).

**Lithostratigraphy**

Lithological descriptions are based on terminology used by the International Ocean Discovery Program (IODP) to facilitate comparison with marine cores. US Geological Survey lithological patterns or swatches were used for graphic logging. Sedimentary structures and accessory elements were noted and obvious evidence of drilling disturbance was recorded. Logging was carried out by different teams. Cores 1–21 were logged primarily by C. Nicholas, T. Dunkley and A. O’Halloran whereas Cores 22–82 were logged by P. Sanyoto from the GRDC team, who oversaw completion of the drilling. Although there is a general consistency between the logging methods there is greater detail for Cores 1–21. Scanned copies of the original field logs and core photographs for Cores 1–21 are available in Supplementary material, Appendices 1 and 2. Logs for Cores 22–82, which were used to construct the summary log (Fig. 4), are currently not available. Basic field mapping was undertaken in the immediate vicinity of the drill site using a hand-held global positioning system (GPS) device to record locality coordinates using a World Geodetic System (WGS) 84 Universal Transverse Mercator (UTM) 49M grid (Square 49 M) as a base map.

**Planktonic foraminifera**

Preparation and examination of foraminifer samples was performed at Cardiff University. Clay samples were placed in beakers and gently soaked overnight in deionized water. The clay was then washed over a 63 μm mesh sieve to separate the sand fraction, which was then oven-dried at 40°C before storage in vials. Calcareous microfossil preservation was qualitatively assigned as: Excellent (E), containing mostly ‘glassy’-standard; Good (G), some glassy specimens; Moderate (M) ‘frosty’-standard; Poor (P) evidence of strong dissolution and/or recrystallization of the test wall; or Barren (B). Assemblages were analysed to species level based on examination of the >150 and 63–150 μm sieve fractions, and a stratigraphic range chart was constructed. Full assemblage counts were not made but recognized taxa were qualitatively assigned to the following categories: abundant (A), common (C), rare (R) and single (S). Biostratigraphically important events were identified and used together with nannofossils, to build an age–depth model. Taxonomy follows the Atlases of Eocene and Oligocene planktonic foraminifera (Pearson et al. 2006; Wade et al. 2018). Selected foraminifera were imaged using the scanning electron microscopes (SEM) at Cardiff University and Stockholm University for taxonomic purposes and to assess the state of preservation. Prior to imaging, specimens were immersed in an ultrasonic bath for 10 s to remove adhering clay. A suite of crushed and cleaned foraminifera were SEM-imaged to evaluate the cleaning method. Specimens were gold-coated prior to imaging. Planktonic foraminifera microslides
Fig. 4. Stratigraphic log of the NKK-1 borehole. The biostratigraphic zonation follows Jones et al. (2019), with updates herein. DWAF, deep-water agglutinated foraminifera. Symbols document the general presence of common fossil or mineral components throughout the core.
and SEM stubs are archived at Stockholm University, Department of Geological Sciences in the collections of H. K. Coxall.

**Benthic foraminifera**

Benthic foraminifer assemblages were studied at Cardiff University using 48 of the planktonic foraminifera preparations. The taxonomic results are reported in descriptive form. Relative proportions of planktonic and benthic foraminifera (subdivided into calcareous and agglutinated forms) were counted in the coarse fraction (>125 μm) by spreading each whole sample on a picking tray and counting the first 300 foraminiferal specimens. Attempts were made to optimize the sample evenness, and thus minimize gravity sorting by including the whole sample and sprinkling in multiple sweeping circles. Nevertheless, results should be regarded as semi-quantitative. Representative forms were imaged using the Cardiff University SEM. SEM stubs and benthic foraminifera microslides were in the collections of Ian MacMillan. Unfortunately, their current location is unknown.

**Calcareous nannofossils**

Calcareous nannofossils in the NKK-1 core were studied previously for biostratigraphy and to investigate palaeoceanographic dynamics across the EOT (Jones et al. 2019; Jones and Dunkley Jones 2020). This involved fully quantitative assemblage counts and morphometric analyses using the ‘simple smear slide preparation technique’ and the ‘extended count technique’ of Bown and Young (1998). An additional 11 samples from the upper 40 m of the core (from Cores, 3, 9, 15, 21 and 24–30) were studied here to provide biostratigraphic control in the upper part of cored section. This involved qualitative examination of smear slides to identify zonal markers. The slides are archived in the collections of T. Dunkley Jones at the University of Birmingham, School of Geography, Earth and Environmental Sciences.

**Biostratigraphy and age modelling**

Biostratigraphic zonation and age modelling for the NKK-1 core utilizes the integrated calcareous nannofossil and planktonic foraminiferal biostratigraphy of Jones et al. (2019), with additional constraints in the upper 40 m added here. The bioziation schemes of Wade et al. (2011) are used for planktonic foraminifera (‘E’ and ‘O’ Zones) and those of Agnini et al. (2014) and Martini (1971) for calcareous nannofossils. (‘CN’ and ‘NP’ Zones respectively). The age model is here updated after Jones and Dunkley Jones (2020). It applies only to the late Eocene–early Oligocene part of the section (Cores 30–67) where preservation was optimal. All ages are adjusted to the Geologic Timescale 2012 (Gradstein et al. 2012).

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**Fig. 5. NKK-1 foraminifera preservation and the test fragmentation method used for planktonic foraminifera isotopic analysis.**

(a) NKK1-32, 112–120 cm, 44.66 m, whole, unfilled specimen (*T. ampliapertura*) showing exceptional (glassy) preservation; (b) NKK1-32, 112–120 cm, 44.66 m, cross-section of a broken test wall showing the pristine preservation including evidence of the bilamellar wall structure; (c) NKK1-42, 62–71 cm, 56.67 m, whole, ‘filled’ specimen with excellent external test calcite preservation despite sparry calcite infill; (d) NKK1-42, 62–71 cm, 56.67 m, cross-section of a broken, infilled test showing the distinction between test calcite and infill; (e) NKK1-42, 62–71 cm, 56.67 m, test fragments including some with remaining infill; (f) NKK1-32, 112–120 cm, 44.66 m, test fragments free of infill ready for isotopic analysis. Scale bars: a, c and e, 100 μm; b and d, 20 μm; f, 200 μm.
Oxygen and carbon stable isotopic analysis

Foraminiferal δ18O and δ13C analysis used *Turborotalia ampliapertura* and *Cibicidoides* sp. to trace surface (planktonic) and bottom water (benthic) conditions respectively (Pearson et al. 2008). Isotopic analysis was performed at Cardiff University using a MAT252 gas source mass spectrometer with an automated KIEL carbonate preparation unit. Results were calibrated to the VPDB scale using the international standard NBS19. Analytical precision was ±0.05% for δ18O and ±0.02% for δ13C.

For much of the studied interval, isotopic samples were composed of 5–10 whole, empty specimens of *T. ampliapertura* or 3–4 specimens of *Cibicidoides*. In the basalt Oligocene foraminifera are affected by secondary crystalline infilling (Fig. 5). We addressed this by separating test calcite from sparry calcite infill. The infilling affected globular-shaped *T. ampliapertura* most severely. Primary test wall calcite was collected by first crushing 15 whole, empty specimens of *T. ampliapertura* and then separating the fragments, excluding those where infill remained attached (Fig. 5), using a fine paint brush in a small amount of deionized water under a binocular reflected light microscope. Fragments were pooled and excluding those where infill remained attached (Fig. 5), using a fine test wall calcite was collected by first crushing 15 whole, empty specimens of *T. ampliapertura* and then separating the fragments, excluding those where infill remained attached (Fig. 5), using a fine paint brush in a small amount of deionized water under a binocular reflected light microscope. Fragments were pooled and excluding those where infill remained attached (Fig. 5), using a fine paint brush in a small amount of deionized water under a binocular reflected light microscope. Fragments were pooled and excluding those where infill remained attached (Fig. 5), using a fine paint brush in a small amount of deionized water under a binocular reflected light microscope. Fragments were pooled and excluding those where infill remained attached (Fig. 5), using a fine paint brush in a small amount of deionized water under a binocular reflected light microscope. 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Fig. 6. Surface geological map and sketch cross-section of the region surrounding the NKK-1 rig site mapped by C. Nicholas during the 2006 Nanggulan coring campaign. Shown with a 50 m WGS 84 UTM 49 M grid. Planktonic foraminifera biozone assignments for the various lithologies are indicated.
suspected horizon of the E/O boundary. Bioturbation occurs throughout the recovered clays, manifest as greenish grey spots. In Core 57 numerous chips of brown cemented carbonate occur, interpreted to have been derived from a hardground surface, suggesting a short-lived interruption of sedimentation. Hardening towards its base, Unit VB ends abruptly at 89 mbs to be underlain by alternating beds of bioturbated greenish grey sandy mudstone with radiolaria and dark greenish grey volcanic sandstone (Unit VI), in which the borehole terminates at 100 mbs.

**Surface geology**

A simple geological map was produced over a 2 day period (8–9 January 2006), which identifies a repetitive sequence of chalky marls and interbedded light blue–grey to olive grey clays (Fig. 6). The regional dip of the strata is c. 25° towards the NW, essentially into the hillside. The repetition of chalky marls and blue–grey clays occurs around a pronounced bench and spring line towards the base of the hillside. This, combined with the fact that no such repetition occurs in the NKK-1 borehole, suggests that this bench is a slumped block rather than a thrust or angular unconformity (Fig. 6). Given the high rainfall and mountainous terrain in a sequence dominated by clays, this seems the most likely explanation. To the SE, a gently folded package of thinly interbedded clays, arkosic sandstones and tuffs marks the base of the Paleogene in the map area. The sediments at Nanggulan crop out as an infiller, unconformably overlain by younger volcanic rocks, mostly consisting of agglomerates with feldspar porphyry clasts. Previous researchers (Rothpletz 1944; Hartono 1969; Lunt 2013a) identified a distinctive horizon referred to as the *Pellatispira* Limestone found cropping out as isolated thin slabs that are indicated on Lunt’s field map. We did not find these; however, their estimated position inferred from maps of Lunt (2013a) is shown in Figure 6, at the base of the main hill, and this may simply have been covered by undergrowth or during recent cultivation.

**Micropalaeontology and biostratigraphy**

Planktonic and benthic marine microfossils occur throughout the cored sedimentary section, apart from a c. 3 m interval surrounding the 1.5 m thick basalt, with altered sediment contacts above and below and evidence of extensive calcite dissolution within the encasing sandstone (Fig. 6). Foraminifera in the upper 20 m of the core, especially the marly chalk, are dissolved, sometimes highly. At other horizons there occurred overgrowth and/or infilling of shells by secondary calcite or pyrite. This is similar to the microfossils from comparable hemipelagic clays in coastal Tanzania (Pearson et al. 2004; Jiménez Berrocoso et al. 2015). Benthic foraminifera are outnumbered by planktonic foraminifera (typically 1–10% benthic foraminifera), although they are more common in the upper Eocene compared with the Oligocene (Fig. 7). Pyrite-filled burrow casts, up to 1 cm long, are common at several horizons and fish teeth also occur fairly commonly at lower levels. Radiolaria are abundant in the volcaniclastic unit extending from Core 67. No time has been spent on the radiolaria here or previously. Palynomorphs, also known from Nanggulan (Takahashi 1982; Lelono and Morley 2011), have not been investigated in the current study.

**Planktonic foraminifera**

We examined 127 core samples for foraminifera content, with increased detail (closer sample spacing) in the silty clays occurring c. 41–88 mbs (Unit V) (Supplementary Table 2). In the most carbonate-rich upper portion of the hole (Unit I) planktonic foraminifera are abundant but tests are somewhat dissolved and recrystallized (Fig. 8). They are less common in the clay-rich sediments below but preservation is good to excellent, with common glassy specimens in which close to original wall textures are preserved (Figs 8–12). Infilling with sparry calcite is common, becoming extensive in lower Oligocene samples (Fig. 5). Despite infillings, wall textures remain intact, suggesting that the secondary calcite crystallized from carbonate-saturated porewater rather than in situ corrosion and reprecipitation.

The chalky marl samples of Unit I (Cores 1–5) were difficult to break down. Nevertheless, they yield diverse assemblages suitable for biostratigraphic purposes. Among the common taxa are inflated *dentoglobigerinids*, including *Dentoglobigerina selli*, *Globigerina archaeobulloides*, the toothed *Subbotina projecta*, members of the *Ciperoella* group and *Globorotalia*-like morphologies with flattened spiral sides and five chambers in the final whorl suggestive of the *Paragloborotalia semivera/mayeri* group. Rare examples of *D.
are also found (Fig. 8, 4) along with a few Catapsydrax unicavus and C. dissimilis. The presence of Ciperoella angulisuturalis (Fig. 8, 2) in the absence of Chiloguembelina leads to an O5 zonal assignment. Sample NKK1-5, 40–54 cm at the base of the chalky marl, was similar in general assemblage composition but the presence of few to common Chiloguembelina in the 65–125 μm
Fig. 9. SEM images of selected lower Oligocene and upper Eocene planktonic foraminifera from the NKK-1 borehole. 1, *Catapsydrax unicavus*, reproduced from the Oligocene Atlas pl. 4.3 fig. 15; 2, *Dentoglobigerina tapuriensis*; 3 and 4, *Turborotalia ampliapertura*, including wall close-up; 5, *D. galavisii/globularis*; 6, *D. tapuriensis*; 7, *Paragloborotalia nana*; 8, *Globoturborotalita pseudopraebulloides/ouachitaensis*; 9, *Cribrohantkenina inflata*; 10, *Hantkenina alabamensis*; 11 and 12, *H. nanggulanensis*; 13, *H. primitiva*; 14–17, *H. nanggulanensis*. Sample information: 1–8, NKK1-33b, 45–105 cm (Zone O1); 9–13, NKK1-43b, 100–109 cm (Zone E16); 14 and 15, NKK1-45, 49–59 cm (Zone E16); 16 and 17, NKK1-39, 50–61 cm (Zone E16). Scale bars: 1–3, 5–8 and 13–16, 100 μm; 4 and 17, 10 μm; 9–12, 200 μm.
Fig. 10. SEM images of selected upper Eocene planktonic foraminifera from the NKK-1 borehole. 1, *Dentoglobigerina galavisi*; 2 and 3, *Turborotalia cerroazulensis/coccoaensis*; 4 and 5, *T. cuniulensis*; 6, *Dentoglobigerina pseudovenezuelana*; 7 and 8, *Hantkenina nanngulanensis*; 9, *H. primitiva*; 10, *Streptochilus* sp.; 11, *Dentoglobigerina tripartita*; 12, *Subbotina projecta*; 13, *S. goritani*; 14, *T. ampliapertura* or *T. pomeroli* transition; 15, *S. angigoroides*; 16, *S. linaperta*; 17, *S. utilisindex*; 18, *Pseudohastigerina micra*; 19, *P. naguewichiensis*. Sample information: 1–3, NKK1-39, 50–61 cm (Zone E16); 4–6, NKK1-43b, 100–109 cm (Zone E16); 7 and 8, NKK1-46a, 40–49 cm (Zone E16); 9–11, NKK1-50, 38–47 cm (Zone E16); 12–14, NKK1-65, 34–45 cm (Zone E15); 15–19, NKK1-63, 0–10 cm (Zone E15). Scale bars: 1, 2, 7, 8 and 12–14, 200 μm; 3–6, 9, 11, 15 and 16, 100 μm; 10 and 17–19, 50 μm.
Fig. 11. SEM images of selected middle Eocene planktonic foraminifera from the NKK-1 borehole. 1 and 2, *Globigerinatheka semiinvoluta*; 3, *G. mexicana*; 4, *Subbotina frontosa*; 5, *Morozovelloides crassatus*; 6, *Acarinina praetopilensis/topilensis*; 7, *Acarinina cf. rohri*; 8, *Acarinina mcgowrani/pseudotopilensis*; 9, *Pseudohastigerina micra*; 10, *Globigerinatheka index*; 11, *G. barri*; 12, *Paragloborotalia pseudowilsoni*; 13, *Hantkenina dumblei*; 14, *H. compressa*; 15, *Turborotalia altispiroides*; 16, *T. pomeroli*. Sample information: 1–3, NKK1-66, 24–33 cm (Zone E14); 4–12, NKK1-75, 32–41 cm (Zone E13); 13, NKK1-77, 4–13 cm (Zone E13); 14–16, NKK1-82, 90–97 cm (Zone E13). Scale bars: 1–12, 100 μm; 13, 500 μm; 14–16, 200 μm.
Fig. 12. SEM images of selected middle Eocene benthic foraminifera from the NKK-1 borehole. 1, *Hanzawaia* sp.; 2, *Cibicidoides* sp.; 3 and 4, *Gyroidinoides* sp.; 5, *Nodosaria*?; 6, *Cibicidoides* sp.; 7, *Pullenia* sp.; 8, *Anomalinooides* sp.; 9, *Oridorsalis umbonatus*; 10, *Chloostomelloides* sp.; 11, *Odogenerina* sp.; 12, *Dentalina* cf. *solata*; 13, *Nodosaria* sp.; 14, *Stiphonodosaria*?; 15, *Dentalina* sp.; 16, *Nodosaria*?; 17, *Pleurostomella* sp.; 18, pyrite-filled burrow; 19, *Pleurostomella* sp.; 20, *Quinqueloculina* (transported) sp.; 21, *Fissurina* sp.; 22, *Tritaxilina* sp.; 23, *Karreriella* sp.; 24, *Lenticulina* sp. Sample information: 4, 9, 18, 20, 23 and 24, NKK1-33B, 95–105 cm (Zone O1). Others; NKK1-82, 90–97 cm (Zone E13). Scale bars: 1, 2, 5, 6, 9–19, 22 and 23, 200 μm; 3, 4, 7, 8, 20 and 21, 100 μm; 24, 500 μm.
fraction implies a Zone O4 assignment. Zone O4 continues into the claystone of Unit IIA (Cores 6–11). Chilognemelina are small and rare, occurring only in the 65–125 μm fraction. Calcareous benthic species are notably more common than in open ocean deep-sea cores, especially in the coarse fraction.

Preservation improves slightly in the silty claystone of Unit IIB (Cores 12–21). Benthic species continue to be relatively common, but better preserved than in the previous core. The preservation improves more substantially in the 65–125 μm fraction. Calcareous benthic species are notably more common than in open ocean deep-sea cores, especially in the coarse fraction.

The poorly consolidated sands of Unit III are barren of planktonic foraminifera, apart from occasional poorly preserved specimens in Core 29. From Core 30, beneath the basaltic lava flow (Unit IV), the last three samples examined above the E/O boundary on the other side of the Indian Ocean (Wade and Pearson 2008). Assemblages attributable to Zone E16 continue to Core 52. Preservation is variable, typically good to moderate, at times excellent or glassy with or without infills of sparry calcite and commonly with pyrite. Dengotoglobigerinids, especially D. galavisí and D. pseudovenezuelana, are consistently abundant (Fig. 10, 18 and 19) consistent with previous observations of a sharp size reduction in the genus at the E/O boundary on the other side of the Indian Ocean (Wade and Pearson 2008).

Similar assemblages attributable to Zone O1 persist to the bottom of Core 38 (52.62 mbs). Preservation generally varies between good and moderate, although poorly preserved tests also occur, implying variable calcite preservation at a small scale. The relative abundance of subbotinids, dengotoglobigerinids and paragloborotaliids fluctuates. Cataspis oxyrhis is a consistent element, without C. dissimilis. Pyrite infills and lose pyrite grains become common at times. Samples NKK1-33a, 32–43 cm and NKK1-33b, 95–105 cm are particularly notable for the occurrence of abundant red-stained ‘oxidized’ grains in the coarse fraction, together with abundant pyritized filled burrow casts, with some test calcite infilling. Sample NKK1-35a, 25–35 cm contains the first down-core occurrence of Acarinina meditex. Here, a few larger C. cubensis and P. naguewichiensis occur in the 150–250 μm fraction as well as being abundant in the smaller fraction. Samples from Cores 36–38 stand out because of common large pyritized burrow casts, unilocular calcareous benthic foraminifera in the >250 μm fraction (Fig. 12), and abundant large globular planktonic foraminifera, sometimes with strong iron staining. Relatively large and common fish teeth also occur. These observations imply some extreme in primary production and export to the seafloor. The last three samples examined above the E/O boundary (Core 38, basal O1) are different again, containing abundant dark grey mineral grains, probably manganese oxide, together with abundant and large deep-water agglutinated benthic foraminifera tubes of Bathysiphon and common unilocular calcareous benthic foraminifera in the >250 μm fraction. Sample NKK1-38, 108–116 cm sitting directly above the E/O boundary contains abundant deep-water agglutinated benthic foraminifera (Fig. 7). Here dissolution and pyrite infills are especially noticeable. An abrupt yet subtle colour change from greenish grey to a lighter greenish grey at 146 cm in Core 38 falls in the exact window identified by biostratigraphy to contain the E/O boundary and we predict this might be the boundary itself at 52.96 ms.

Still in Unit V, the first down-core occurrence of Hantkenina, and thus the Eocene, is in sample NKK1-39, 40–50 cm, which was found to have a few hantkeninids in the 125–250 μm range (H. alabamensis and H. primitiva). Here infilling with sparry calcite is common, as is again iron oxide staining. Many specimens, of various species, exhibit atypical test morphologies, including kummerform or irregular-shaped chambers suggestive of environmental stress (see discussion by Pearson and Wade 2015 for comparable observations in Tanzania). Turborotalia ampliapertura remains common, the angular T. cerroazulensis group (i.e. T. cerroazulensis, T. cunialensis and T. cocaensis) less so, although good examples of the latter occasionally occur (Fig. 10, 2–59). In the absence of globigerinathekids this sample is assigned to Zone E16. Pseudoasthgitina, mostly P. micro, is immediately larger and more frequent in Zone P16 compared with O1 above (Fig. 10, 18 and 19) consistent with previous observations of a sharp size reduction in the genus at the E/O boundary on the other side of the Indian Ocean (Wade and Pearson 2008).

Table 2 we show the occurrence, in one sample (NKK-1/52, 70 m) subbotinids, we also show the occurrence of abundant red-stained ‘oxidized’ grains in the coarse fraction, together with abundant pyritized filled burrow casts, with some test calcite infilling. Sample NKK1-35a, 25–35 cm contains the first down-core occurrence of Acarinina meditex. Here, a few larger C. cubensis and P. naguewichiensis occur in the 150–250 μm fraction as well as being abundant in the smaller fraction. Samples from Cores 36–38 stand out because of common large pyritized burrow casts, unilocular calcareous benthic foraminifera in the >250 μm fraction (Fig. 12), and abundant large globular planktonic foraminifera, sometimes with strong iron staining. Relatively large and common fish teeth also occur. These observations imply some extreme in primary production and export to the seafloor. The last three samples examined above the E/O boundary (Core 38, basal O1) are different again, containing abundant dark grey mineral grains, probably manganese oxide, together with abundant and large deep-water agglutinated benthic foraminifera tubes of Bathysiphon and common unilocular calcareous benthic foraminifera in the >250 μm fraction. Sample NKK1-38, 108–116 cm sitting directly above the E/O boundary contains abundant deep-water agglutinated benthic foraminifera (Fig. 7). Here dissolution and pyrite infills are especially noticeable. An abrupt yet subtle colour change from greenish grey to a lighter greenish grey at 146 cm in Core 38 falls in the exact window identified by biostratigraphy to contain the E/O boundary and we predict this might be the boundary itself at 52.96 ms.

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Assemblages attributable to Zone E16 continue to Core 52. Preservation is variable, typically good to moderate, at times excellent or glassy with or without infills of sparry calcite and commonly with pyrite. Dengotoglobigerinids, especially D. galavisí and D. pseudovenezuelana, are consistently abundant (Fig. 10, 1 and 6) whereas subbotinids come and go, with Subbotina linaperta being most common and S. angiporoides and S. utilisindex less so (Fig. 10). Hantkeninids show variability in abundance and diversity with some samples containing all five late Eocene species. Here the large and infilled species H. nanggulanensis described by Hartono (1969) from the Nanggulan section and its descendant Cribrohantkenina inflata with additional areal apertures are notable (Fig. 9, 9–17; Fig. 10, 7–9). Hantkeninids are at times unusually abundant; for example, in Sample NKK1-1/47, 90–100 cm (62.95 mbs). Foraminiferal preservation tends to be poorer (‘frosty’ preservation) where there are more hantkeninids, and relatively few hantkeninids occur in very well-preserved (more clay-rich) samples. In general, diversity appears to be lower in ‘glassy’ samples.

It should be noted that in our range chart (Supplementary Table 2) we show the occurrence, in one sample (NKK-1/52, 70–78 cm), of the species Globotorotalitida barbula Pearson and Wade, known for its covering of tiny prickle-like ‘barbules’. This typically small species (150–200 μm) was described after our
general NKK-1 assemblage work, and thus was not originally recognized in the census. However, because paratypes of *G. barbula* are derived from the NKK-1 material we can now acknowledge its occurrence and anticipate it being common in late Eocene and early Oligocene NKK-1 samples, as indicated by the statement by Pearson and Wade (2015): ‘*Globoturborotalita barbula* n. sp. is a common component of the Tanzanian assemblages both before and after the Eocene/Oligocene boundary and we have also confirmed that it is equally common in the Eocene/Oligocene transition of Java (unpublished data).’ Most probably *G. barbula* has here been described within *Globoturborotalita martini* and/or *G. oachitaensis*.

Core 53 sees a change in colour from greenish grey to rusty brown clay. Extending down to Core 56, this zone is broadly associated with an interval of often glassy foraminiferal preservation, although samples with good preservation come and go down to Core 65. *Globigerinatheka* index is rare and discontinuous, thus its highest stratigraphic occurrence in sample NKK-1/53, 30–42 cm (70.86 msb) provides only a tentative delineation of Zones E16–E15 (Fig. 1). Abundant pale orange and olive green crystalline lumps, in samples NKK1-55a, 25–36 cm and NKK1-57, 22–32 cm are probably remains of a volcanic ash. Brown-coloured cemented carbonate aggregates in NKK1-57, 22–32 cm may be fragments of a broken-up carbonate hardground that indicates a break in sedimentation. Abundant small pyrite burrow casts at this level are consistent with this idea. Sample NKK1-61, 91–101 cm exhibits rather good preservation, together with abundant crystal masses indicating another ash bed. *Paragloborotalia nana* is often abundant, as are *Chiloguembelina* and *P. micro* in the fine fraction. *Globoturborotalita ouachitaensis* and *G. martini* are also consistent elements. Sample NKK1-65, 35–45 cm exemplifies the good preservation and assemblage diversity, with common *Turborotalia* spp., *D. galavisii* and *D. tripartita*. Here the subbotinoids are also well represented, with high-spired *S. gortanii* and ‘buck tooth’ *S. projecta* being conspicuous components (Fig. 10, 12 and 14). The E14 biomarker *G. seminovolatus* is found in a single sample, NKK-1/66, 24–33 cm, where it is fairly common (Fig. 11, 1 and 2). This sample also records an abrupt decline in preservation quality and a shift in the assemblage composition. Notable is a marked increase in globigerinathekids, the occurrence of rare *Hankenia dumblei*, fewer dentoglobigerinids and a shift in turborotalids to just *T. cerroazulensis*. *Chiloguembelina*, *P. micro* and globoturborotalitids remain abundant. We place the E15–E14 boundary at 87.09 msb but note that the associated preservation and assemblages shift is suggestive of a hiatus (Supplementary Table 2).

The lithological transition from the clays of Unit V to the interbedded clays and tuffaceous sands of Unit VI occurs somewhere in Core 67. Sample NKK-1/67, 36–45 cm (88.90 m below seafloor (msb)) required multiple wet–dry sieve cycles to disaggregate grains and foraminiferal preservation is poor (frosty and dissolved specimens), although still suitable for biostatigraphic purposes. Microfossils are heavily diluted by coarse (>250 μm) unwashed volcanic grains. Radiolaria and their fragments are also abundant in the 63–125 μm fraction. Here *Turborotalia* is limited to *T. frontosa* and *T. pomeroli*. Dentoglobigerinids and globoturborotalids are absent and hantkeninids are limited to *H. dumblei* and a single *H. compressa* in this and samples below, *Parasubbotina griffithiae* occurs together with rare *P. hagni* and *P. pseudowilsoni*. Globigerinatheka is noticeably more common in this sample, including forms with multiple small bullae capping supplementary apertures, attributable to *G. barri*, a possible *G. kugleri* and *G. euganea*. Among the muriates are found a few *Morozovelloides crassatus*, *Acarinina collaepta*, *A. bullbrooki*, *A. mcgowrani* and *A. praetopilensis* (Fig. 11), although generally *Morozovella* and *Morozovelloides* are never common in our samples. Similar assemblages diluted by volcanic material extend to the bottom of the hole. Other notable species are *T. altispiroidea*, persistent parasubbotinids, a few morozovelloidids, the occasional *H. liebasi* together with *H. dumblei* and a few globigerinathekids (Fig. 11).

Biozonation from sample NKK-1/67, 36–45 cm (88.90 msb) to the base of the hole is somewhat uncertain from a planktonic foraminifera perspective. In our original census and range chart we record scarce ‘*Guembelitrioceras nutalli*’ in sample NKK1-73, 37–46 cm and the one below (Supplementary Table 2), which would imply a Zone E10 assignment. However, this is inconsistent with the evolutionary character of the hantkeninid assemblages (i.e. *H. dumblei* and *H. compressa* without *H. mexicana*), which are suggestive of a younger age (Coxall and Pearson 2006). Furthermore, *Turborotalia* and its descendant *T. cocoensis* and *pomoriello*, which evolved in upper E10, c. 42.2 Ma, are common in outcrop samples taken from below the base of the NKK-1 core (Lunt 2013a), implying that the base of the NKK-1 borehole cannot be as old as E10. The short-ranging *Orbulinoides beckmannii* Zone (E12) was not found in any of the basal NKK-1 samples, thus the planktonic foraminifera alone imply undifferentiated middle Eocene Zones E11–13. The nanofossil zonal assignment CNE15 (Agrini et al. 2014), implied by the presence of *Dictyococcites scrippsi*, also known as *Reticulofenestra bisecta*, and occurring to the bottom of the hole, combined with previous constraints (Okada 1981; Lunt 2013a), supports the younger Zone E13 assignment. Closer inspection of the *G. nutalli*-like morphologies reveals a lack of supplementary apertures diagnostic of the species. Similar ‘deceptive’ high–spired *nutalli*-like forms have been found in other middle Eocene sections above the calibrated range of *G. nutalli* (e.g. ODP 865; Young et al. 2019; Edgar et al. 2020). It is possible that these are in fact *Subbotina gortani*.

**Benthic foraminifera**

The 48 studied samples provide an overview of Nanggulan EOT benthic foraminifera. Identifications are mostly to genus level. There is as yet no comprehensive publication on the smaller benthic foraminifera of the Javanese Paleogene and it is probable that many of the observed species are new. The NKK-1 benthic foraminifera assemblages are unlike those from Tanzania (Pearson et al. 2008; Cotton and Pearson 2011, 2012), especially at the species level, which is not surprising given the provincialism seen in benthic foraminifera (Murray 1991; Pawlowski and Holzmann 2009). Benthic test preservation is moderate to excellent with the best preservation occurring in the clays-rich intervals (Figs 5 and 12). The semi-quantitative counts show that planktonic foraminifera generally outnumber benthic foraminifera, although the latter are relatively more common in the late Eocene (c. 5–40%) compared with the Oligocene (c. 1–5%) (Cores 36–40, Unit VA) (Fig. 7, Supplementary Table 3). The assemblages are composed mostly of calcareous deep-water taxa, with deep-water agglutinated benthic foraminifera being a consistent background component, becoming temporarily very abundant (almost 40% of the total foraminiferal assemblage) around the EOT and in the basal early Oligocene.

Despite some changes in lithology, the benthic foraminifera assemblages throughout the succession are very similar at the generic level but not at the species level. Typically, only one or two specimens per species in a sample occur in the NKK-1 samples (planktonic foraminifera Zones E13–OS) and these are often abnormally large. No plant material or organic debris has been found in any studied samples, although black particles suggestive of organic matter were seen in cores (Fig. 6). Apart from a few miliolids (*Quinqueloculina* and *Triloculina* spp.) in Cores 36–42 and 82 (Fig. 12), there is no significant amount of transported shallow water species (neither miscellaneous shell, smaller foraminifera tests, nor larger foraminifera debris) in the clay-rich samples studied. Small numbers of miliolids have been found in modern (box-core) material.
from eastern Indonesia in the Timor Sea, all the way into upper bathyal settings (van Marle 1988), supporting some small degree of offshore transport. Tiny fragments of fish bone, with occasional fish teeth and scales, occur through much of the upper Eocene succession (Cores 43–57, Units VA and VB) and occasionally elsewhere. This predominantly oceanic setting implied by the NKK-1 benthic foraminifera assemblage is different from that of the Eocene of coastal Tanzania, where frequent influxes of well-preserved material from the shelf occur throughout the clays (Pearson et al. 2004, 2006; Cotton and Pearson 2012).

In the Oligocene unit (Unit VA) (Cores 38 and 37), there is a basal interval rich in benthic foraminifera, including agglutinated species, and an overlying interval in which benthic foraminifera gradually fade away upwards, becoming few or rare (Cores 36 to 1) (Fig. 7). This suggests an interval of progradation. The agglutinated genus *Thalmannammina* is limited to the basal unit, and *Pullenia* and *Cassidulina* range sporadically through the basin unit and the overlying beds, but neither these genera, nor the remainder of the benthic assemblage provide accurate age indications. In addition, no species appear restricted to particular planktonic foraminifera zone intervals, and there appears to be no marked foraminiferal change indicative of depth change through the Oligocene, during which large eustatic sea-level changes of ±50 m sea-level fall occurred associated indicative of depth change through the Oligocene, during which large eustatic sea-level changes of ±50 m sea-level fall occurred associated with episodes of Antarctic growth and decay (Miller et al. 2020).

Typical deep-water agglutinated benthic foraminifera taxa include species of *Bathysson* (particularly numerous in earliest Oligocene Cores 37 and 38), *Thalmannammina* (limited to Cores 37 and 38) and, scattered throughout the entire succession, *Volulina, Spiroplectinella, Karreriella, Repmanina charoides, Clavulina, Bolivinopsis* and possible *Tritaxilina*. At several levels occasional attached vagile agglutinated tubes of *Lituotuba* encrust both benthic (*Lenticulina*) and planktonic tests. There are several distinct species of *Karreriella, Volulina* and *Spiroplectinella* in particular, and some of these may be stratigraphically limited. Typical calcareous benthic foraminifera include species of *Oridorsalis, Globocassidulina, Osangularia*, strongly ornamented spinose and ribbed *Uvigerina, Euuvigerina, Gavelinella, Cibicidoides, Stilostomella, Siphonodarosa, Dodogerina, Mattauella, Nodosarella, Pleurostomella, Ellipsoidina, Cassidulina, Lenticulina, Dentalina, Nodosaria, Brizalina, Fontbotia, Pullenia, Hanzaowia* and *Gyroidinoidea*. The stratigraphic ranges of many of these species are unknown, although already it can be seen that the genera *Cassidulina* and *Pullenia* are more common in the Oligocene part of the succession.

The clay-rich sediments of Cores 66 to 39, inferred by biostratigraphy to be late Eocene, show minor variation in benthic foraminifera. The lower part (Cores 66 to 61, lower part of Unit VA) contains relatively abundant benthic foraminifera. From Cores 59 to 39 benthic foraminifera are generally rare or very rare, with pyritized burrow infills and tiny fish bone fragments occurring widely. Again, none of the diverse benthic foraminifera species in this interval are age-diagnostic. This basal interval of the borehole (Unit VI) is characterized by rare benthic foraminifera throughout, only one of which, a possible *Tritaxilina*, appears restricted to this interval. Probable *Tritaxilina coleii* occurs in the topmost samples of TDP 13 (in Tanzania, unpublished), in an interval dated as E11 on the basis of the planktonic foraminifera, but the Javanese and Tanzanian species are clearly different. No *Discocyclina* or other larger benthic foraminifera were found in the samples studied from Unit VI.

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**Fig. 13.** Biostratigraphic age model (Option 2) for the NNK-1 borehole. Age–depth tie-points are plotted as event mid-points with associated depth errors (Table 3). T, top; B, base; LCO, lowest common occurrence; HCO, highest common occurrence. Planktonic foraminifer (PFZ) and calcareous nannofossil (CNZ) follow Wade et al. (2011) and Agnini et al. (2014) respectively. Green circles identify the datum events used to constrain the age model. Other planktonic foraminiferal events are also named. Dashed biozone separators indicate low-certainty bioevents in NKK-1. The basalt occurs at c. 39 mbsf (O1–O2 range), and was emplaced between 30.5 and 33 Ma. (See Fig. 6 for lithology legend and Supplementary Fig. 1 for the O1–1 age model.) The radiometric age of the Gajah lavas is from Bellon et al. (1989). Sedimentation rates were calculated using linear regression between the selected tie-points. The grey line indicates the alternate sedimentation curve implied by age model Option 1 (see Supplementary Fig. 2).
suggesting that the borehole did not extend into the upper Lutetian-dated shallower water *Discocyclina*-rich beds (the Watu Puru Beds) described by Lunt (2013a).

**Nannofossils**

Calcareous nannofossils are abundant or common within the NKK-1 succession with preservation varying between typically good in middle Eocene, good to excellent in the upper Eocene to lower Oligocene, moderate in the volcanic-sand rich intervals of Unit VI and moderate to very poor in the early Oligocene (Jones et al. 2019; see Supplementary Fig. 1). The 11 new samples studied here provide nannofossil biostratigraphic constraints in the upper part of the core (Units I and II), which was not included in the previous studies.

Assemblages are characterized by diverse tropical florals that allow the documentation of an almost complete sequence of low-latitude bioevents through Cores 30 to 82 (Zones CNE15 to CNO3) (Table 3. Datum Events). The poorer preservation, combined with low sampling resolution and less detailed analysis, provides few control points in Cores 5–30. For example, identifying the base of CNO4 was not possible because the lowest occurrence of *Sphenolithus distentus* is not found in the entire NKK-1 core. The top of *S. predistensum*, defining the base of Zone CNO5, does occur between 5.85 and 13.60 mbs. However, this zonal assignment is tentative as preservation is poor in the chalky marls and relies on one sample. Detailed counts (Jones et al. 2019) reveal that from the middle Eocene to early Oligocene species biodiversity declined, with the most pronounced species loss occurring among the latest Eocene rosette-shaped discoasters, an event named the ‘Discaster extinction event’. This occurs at c. 60.53 mbs (c. 34.44 Ma).

Overall, the observed nannofossil occurrences in NKK-1 work well with the biozonation scheme of Aignini et al. (2014). Considered. Option 1 (Supplementary Fig. 2; also shown in Fig. 13) incorporates the *R. umbilicus* datum as ‘correct’ and interprets the ‘Top *P. naguevichiensis*’ as the highest common occurrence of this species below the sandstone–basalt, ignoring the rare occurrence as possibly reworked above this unit (Table 3, Supplementary Table 2). This requires a decrease in sedimentation rate between 45 and 55 mbs and assumes continuous deposition from Unit II and IV. However, application of this age model to our stable isotope stratigraphy (see below) resulted in the early Oligocene becoming ‘compressed’ and the δ18O and δ13C isotopic maxima were 255–300 kyr too young compared with the global standards. We thus used an alternative approach whereby the *R. umbilicus* datum is excluded from the age model and the rare occurrence of *P. naguevichiensis* above Unit IIIA was included as the ‘Top *P. naguevichiensis*’ tie-point. Here we extrapolated the sedimentation rate calculated between the Top *Hantkenina* spp. and Top *D. saipanensis* segments to the top of Unit VA, which make sense considering it is the same lithology, and accepted a disruption in sedimentation through units III and IV (Fig. 14). This alternative, which we call Option 2, is our preferred age model, as it results in a realistic calibration of the isotope stratigraphy. It implies that *R. umbilicus* went extinct >300 kyr earlier in Java compared with other places, although there is large depth uncertainty in this bio-event as it was found only in one sample at the level of its highest occurrence, and its occurrence up-section is uncertain because carbonate preservation quality plummeted in the overlying tuffaceous sandstone. Such diachronism among calcareous nannofossils in the late Eocene is not uncommon (Berggren et al. 1995). Moreover, the rarity of *R. umbilicus* in the Java succession is consistent with its well-documented preference for mid- to high latitudes at this time (Villa et al. 2008). According to our preferred age model the 1.5 m thick basalt recovered in Cores 28 and 29 at c. 39 mbs occurs in early Oligocene Zone O1, and was emplaced between 33 and 32 Ma.

**Stable isotopes**

The δ18O of *T. ampliapertura* ranges from −8.5 to −2.6‰ and from −3.5 to +3‰ for *Cibicidoides* sp. across the EOT interval (Figs 14 and 15). *Turborotalita ampliapertura* values in the late Eocene and early Oligocene younger than 33.6 Ma are relatively stable (4.8 to −3.7‰ and −4 to −2.6‰ respectively). Clear positive δ18O and δ13C offsets of the fragment samples from the spartein infill (0.8–2.9‰ for δ18O and 2.3–3.3‰ for δ13C in the four sample pairs analysed) implies that through much of the section primary δ18O and δ13C signals can be obtained using the fragment separation method. However, in the basal Oligocene during the c. 34–33.8 Ma (50–53 mbs) interval, *T. ampliapertura* δ18O temporarily shifts to extremely low values (−8.5 to −5.6‰) (Figs 14 and 15 and Supplementary Fig. 3 (plots in the depth domain)), that are extreme even for this tropical surface-dwelling species (compare values of −3 to −2‰ at EOT time in the western tropical Indian Ocean; Pearson et al. 2008). Such values appear too low to be explained by ocean conditions alone (i.e. SST or salinity). Parallel isotopic depletions in the δ13C signals, as well as in the δ18O and δ13C of benthic foraminifera, suggest diagenetic bias at this horizon.

Importantly, we observe that the most negative δ18O and δ13C values cluster where the most extensive sparry calcite infilling occurs. Shells are not dissolved, test-wall calcite appears glassy and there is no obvious difference in the lithology. Moreover, the few paired analyses of separated spartein infill reveal that the isotopic composition of the sparite is also changing through the core and is most heavily depleted (−8‰ δ18O, −5.7‰ δ13C) at this level, where even test fragments appear to be affected. We thus exclude these data from the general interpretation. Data are excluded through comparison with an ‘unaltered’ late Eocene *T. ampliapertura* δ18O ‘baseline’, by reference to a δ18O v. δ13C cross plot (see Supplementary Figs 3 and 4); *T. ampliapertura* values lower than this baseline, also having relatively
| Type | Datum event | CK 95 Agnini et al. (2014) | Sample ID bottom | Sample depth bottom (m) | Sample ID top | Sample depth top (m) | Datum depth midpoint (m) | Depth error (±) | Comment |
|------|-------------|-----------------------------|-------------------|-------------------------|--------------|----------------------|-------------------------|----------------|---------|
| NF   | T Sphenolithus predistentus | B. CNO5 — 26.93 — 26.93 | NKK-1/10, 35 | 13.60 | NKK-1/3, 35 | 5.85 | 9.73 | 3.88 | Poor preservation, uncertain, not used in age model |
| PF   | HCO Chiloguembelina cubensis | B. O5 28.40 — 28.40 | NKK-1/5, 40–54 | 7.97 | NKK-1/1, 32–34 | 3.38 | 5.68 | 2.30 | Poor preservation, uncertain, not used in age model |
| PF   | B Ciperoella angulisuturalis | B. O4 29.40 — 29.40 | NKK-1/16, 0–12 | 23.06 | NKK-1/14, 0–15 | 21.08 | 22.07 | 0.99 |
| PF   | T Subbotina angiporoides | Within O3 30.00 — 30.00 | NKK-1/34, 19–29 | 46.74 | NKK-1/33, 95–105 | 46.00 | 46.37 | 0.37 |
| PF   | T Turborotalia ampliapertura | T. O2 30.30 — 30.30 | NKK-1/19, 179–189 | 29.34 | NKK-1/18, 0–13 | 26.07 | 27.70 | 1.64 |
| PF   | B Paragloborotalia opima | Within O2 30.06 — 30.06 | NKK-1/21, 103–113 | 31.08 | NKK-1/19, 179–189 | 29.34 | 30.21 | 0.87 |
| PF   | T Pseudohast. naguewichiensis | B. O2 32.00 — 32.00 | NKK-1/30, 89–96 | 41.43 | NKK-1/30, 34–43 | 40.89 | 41.16 | 0.27 |
| NF   | T Reticulofenestra umbilicus? | B. CNO3 — 32.02 — 32.02 | NKK-1/33a, 44–44 | 45.44 | NKK-1/33a, 53–54 | 41.04 | 41.24 | 2.20 |
| NF   | T Coccolithus formosus | B. CNO2 — 32.92 — 32.92 | NKK-1/33b, 120–121 | 46.21 | NKK-1/33a, 43–44 | 45.44 | 45.83 | 0.39 |
| NF   | Be Classicococcus subdistichus | B. CNO1 — 33.88 — 33.87 | NKK-1/36a, 43–44 | 49.94 | NKK-1/35b, 109–110 | 49.10 | 49.52 | 0.42 |
| PF   | T Hantkenina alabamensis = E/O | B. O1 33.7 — 33.70 | NKK-1/39, 40–50 | 53.45 | NKK-1/38, 108–116 | 52.62 | 53.04 | 0.42 |
| PF   | HCO Pseudohastigerina micra | (B. O1) 33.7 — 33.70 | NKK-1/39, 40–50 | 53.45 | NKK-1/38, 108–116 | 52.62 | 53.04 | 0.42 |
| PF   | T Turborotalia cerroazulensis | Within E16 33.8 — 33.80 | NKK-1/39, 50–61 | 53.56 | NKK-1/39, 40–50 | 53.45 | 53.50 | 0.05 |
| NF   | T Discaster saipanensis | B. CNE21 — 34.44 — 34.44 | NKK-1/46a, 55–56 | 61.06 | NKK-1/45, 48–49 | 60.53 | 61.06 | 0.54 |
| NF   | Tc Reticulofenestra reticulata | B. CNE20 — 35.24 — 35.40 | NKK-1/46b, 124–125 | 61.75 | NKK-1/46a, 55–56 | 61.06 | 61.41 | 0.34 |
| PF   | T Globigerinatheka index | T. E15 34.5 — 34.50 | NKK-1/53, 30–42 | 70.86 | NKK-1/51, 32–42 | 67.87 | 69.37 | 1.50 |
| PF   | B Turborotalia cunialensis | Within E15 35.3 — 35.30 | NKK-1/61, 35–47 | 81.91 | NKK-1/60, 33–42 | 80.51 | 81.21 | 0.70 |
| PF   | T Globigerinatheka semiinvoluta | T. E14 35.8 — 36.80 | NKK-1/66, 24–33 | 87.79 | NKK-1/65, 35–45 | 86.40 | 87.09 | 0.69 |
| NF   | Be Reticulofenestra isabellae | B. CNE19 — 36.13 — 36.49 | NKK-1/65, 84–85 | 86.85 | NKK-1/64, 23–24 | 84.25 | 85.55 | 1.30 | Continues to the top of the section |

Last occurrence (including rare occurrences in Unit IIB where preservation is poor (age Option 2)
| NF T Oolithoid/aerosol grains | PF T Sphenolithus obtusus B. CNE16 | PF T Morozovelloides crassatus B. E14 | PF T Globigerinatheka semiinvoluta PF B | PF T Acarinina mcgowrani | PF T Reticulofenestra bisecta (= >99.98 — 99.8) | PF T Dictyococcites scrippsae < 10 µm |
|-----------------------------|-----------------------------------|-------------------------------------|-------------------------------------|-------------------------|-----------------------------------------------|----------------------------------|
| 37.78                       | 38.63                             | 39.30                               | 39.62                               | 39.80                   | 40.34                                         | 40.64                            |
| NKK-1/65                    | NKK-1/66                          | NKK-1/67                            | NKK-1/67                            | NKK-1/66                | NKK-1/67                                       | NKK-1/66                         |
| 37.98                       | 38.02                             | 38.42                               | 38.62                               | 38.82                   | 39.34                                         | 40.64                            |
| NKK-1/65                    | NKK-1/66                          | NKK-1/67                            | NKK-1/67                            | NKK-1/66                | NKK-1/67                                       | NKK-1/66                         |
| 38.02                       | 38.42                             | 38.62                               | 38.82                               | 38.82                   | 39.34                                         | 40.64                            |
| NKK-1/65                    | NKK-1/66                          | NKK-1/67                            | NKK-1/67                            | NKK-1/66                | NKK-1/67                                       | NKK-1/66                         |
| 38.42                       | 38.62                             | 38.82                               | 38.82                               | 39.34                   | 40.64                                         | 40.64                            |
| NKK-1/65                    | NKK-1/66                          | NKK-1/67                            | NKK-1/67                            | NKK-1/66                | NKK-1/67                                       | NKK-1/66                         |
| 38.62                       | 38.82                             | 38.82                               | 39.34                               | 40.34                   | 40.64                                         | 40.64                            |
| NKK-1/65                    | NKK-1/66                          | NKK-1/67                            | NKK-1/67                            | NKK-1/66                | NKK-1/67                                       | NKK-1/66                         |
| 38.82                       | 39.34                             | 40.34                               | 40.64                               | 99.98                   |
| NKK-1/67                    | NKK-1/67                          | NKK-1/67                            | NKK-1/67                            | NKK-1/67                |

Discussion

**Integrated lithostratigraphy, biostratigraphy and age modelling**

The studied section, which sampled the upper part of the Nanggulan Formation, spans c. 13 myr, from the middle Eocene (Lutetian) to the early late Oligocene ( Chattian), and includes an apparently continuous succession through the EOT. The NKK-1 borehole is dominated by 60–70 m of homogeneous volcanic-rich claystones and silty clays and terminates in 12 m of interbedded tuffaceous, andesitic sandstone and clay.

Our combined planktonic foraminifera and nannofossil biostratigraphy provides mutually consistent constraints, although there is apparent diachronuity in some calcareous nannofossil events. Top *R. reticulata* occurs later in Java than in other mid- and high-latitude localities (e.g. the Massignano Section; Agnini et al. 2014), whereas the Top *R. umbilicus* is earlier than expected (Fig. 13). The Top *R. reticulata* event has previously been reported to be strongly diachronous across latitudes (Berggren et al. 1995), as noted above, but *R. umbilicus* has low abundance in the NKK-1 core, limiting the confidence we can place in this constraint. We thus rely more heavily on planktonic foraminifera for age control in the late Eocene. Therefore, although there is not a unique age model solution we have confidence in Option 2 on the grounds that it utilizes a constant sedimentation rate through Unit VA and, as an independent test, produces an age calibration for our stable isotope curve at least in relative timing. Comparison of the absolute δ18O and δ13C with Site 1218 shows that NKK-1 benthic data were systematically lower than the Pacific record in the late Eocene, becoming more similar in the early Oligocene (Fig. 14).

Depleted δ13C, are excluded from the interpretation. The matching δ13C is also rejected. The same approach is used for benthic foraminifera, which, although probably less sensitive to the infilling issue because of the less globular tests, are also affected in the basal Oligocene.

Ignoring the diagenetically influenced basal Oligocene interval (Fig. 15), our planktonic and benthic foraminifera records show an overall increase in δ18O between c. 34 and 33.3 Ma of c. 1.5% for planktonic foraminifera and 2.5% for benthic foraminifera associated with the EOT. Data are sparse and noisy, probably obscuring orbital-scale variability. However, comparison with benthic records from Equatorial Pacific Ocean Drilling Program Site 1218 (Figs 14 and 15) shows that the timing of increase is consistent with the global pattern, although the absolute values and magnitude of change are larger in Java. The majority of the δ18O increase post-dates the E/O boundary (extinction of *Hantkenina* spp.) by c. 200 kyr and maximum benthic δ18O occurring from 33.65 to 33.15 Ma, and a little later for planktonic foraminifera at 33.4–33.05 Ma. Benthic and planktonic δ13C is noisy in the late Eocene, varying by as much as ±1.5% between samples and leaning towards isotopically lower values between 34.5 and 34 Ma. Like δ18O, although interrupted by the basal Oligocene alteration event, overall there is a relative increase in planktonic and benthic δ13C of c. 1.5%. The largest part of the shift appears to occur earlier in benthic foraminifera (c. 33.65 Ma) compared with planktonic foraminifera (33.45 Ma), although the lack of reliable data points between 34 and 33.8 Ma conceals the true pattern. From 33.15 to 33.05 Ma, δ18O and δ13C of benthic foraminifera decrease by c. 1‰, matching the Pacific reference curve at least in relative timing. Comparison of the absolute δ18O and δ13C with Site 1218 shows that NKK-1 benthic data were systematically lower than the Pacific record in the late Eocene, becoming more similar in the early Oligocene (Fig. 14).
our samples that the family Hantkeninidae became scarce before its extinction: the last sample recorded to have hantkeninids contains only rare *H. alabamensis*, whereas the previous sample (10 cm below), contains five species, including *Cribrohantkenina inflata* and *H. nanggulanensis*. The NKK-1 Nanggulan borehole is one of the few EOT sections world-wide having both a stable isotope stratigraphy and the hantkenid extinction, which is necessary for exploring synchronicity of climatic and biotic change. The extinction timing is consistent with previous findings in Tanzania (Pearson *et al.* 2008; Wade and Pearson 2008) and the South Atlantic (Poore 1984; Coxall and Pearson 2006; Hutchinson *et al.* 2021), occurring c. 200 kyr before the major increase in $\delta^{18}$O that records the massive build-up of Antarctic ice in the earliest Oligocene (Early Oligocene Isotope Shift; EOIS). This is apparent despite our coarse isotopic sampling and the complications with diagenetic infilling.

Table 1 compares the stratigraphic units recognized in the Nanggulan region based on previous coring, outcrop sampling and the results of this study, with a broader overview in Figure 16. Importantly, we find in the NKK-1 borehole that the Tegalsari Marls (middle Eocene to upper Oligocene) are 85 m thick. This is significantly thicker than the 10 m thickness reported from outcrop
sampling (Lunt 2013a). We do not find the EOT sequence especially condensed as previously suggested, at least compared with deep-sea standards (e.g. Zachos et al. 1996; Coxall and Wilson 2011). NKK-1 sedimentation rates, which are about 15.2 m Ma\(^{-1}\) below the basalt and 10.1 m Ma\(^{-1}\) above (Fig. 13), are lower than the contemporaneous hemipelagic sequence offshore Tanzania (24–164 m Ma\(^{-1}\); Pearson et al. 2008), implying lower clay inputs in the Nanggulan basin. The dominantly olive green–grey colour of the silty clays of Unit VA, and the common occurrence of pyrite (FeS) is suggestive of somewhat limited bottom water oxygen. The iron might be detrital, but could also be the result of local hydrothermal activity. The combination of temporary up-core colour change into Unit VB from olive–grey silty clay to rusty brownish grey silty clay in Core 56 preceded in Core 57 by a horizon with cemented beds, including P. sp., indicative of shallow water, or at least shallow water debris inputs. The Watu Puru facies is not seen in the NKK-1 core.

At several locations around Nanggulan thin (>1 m, sometimes >0.5 m) limestone debris beds have been found just above the Jetis Beds (Rothpletz 1944; Hartono 1969; Lunt 2013a). These contain abundant shallow marine bioclasts rich in Letter Stage Tb larger benthic foraminifera Discoyclina indicative of shallow water, or at least shallow water debris inputs. The Watu Puru facies is seen in the NKK-1 core.

We identify a clear age gap of c. 1.5–2 myr (Fig. 13) between Unit V and the interbedded tuffaceous sands and clays with high radiolaria content (Unit VI) occurring below 89 mbs, and also seen in outcrop at the drill site (Fig. 4). This unconformity appears to be a regional tectonostratigraphic sequence boundary. The hiatus covers the middle–late Eocene boundary in a relatively deep marine setting, indicating a pause in sediment input. The arkosic or volcaniclastic sands and interbedded arkosic–silty clays below the hiatus are the Jetis Beds of Lunt (2013a) but in the NKK-1 core they are much more fossiliferous than in outcrop samples. This is a character of meteoric weathering of permeable tuffaceous outcrops in the tropics noted by Lunt (Lunt 2013a). The older Watu Puru Beds have units richer in silt-free clays with abundant and extremely well-preserved calcareous microfossils. They also have local concentrations of the larger benthic foraminifera Discoyclina indicative of shallow water, or at least shallow water debris inputs. The Watu Puru facies is not seen in the NKK-1 core.
sedimentary systems very close to the middle to late Eocene boundary. A correlative unconformity is known over a large part of SE Sundaland (Lunt and van Gorsel 2013). The new perspectives from NKK-1 add to the general understanding of the regional stratigraphy in the Kulonprogo–Nangguulan area (Fig. 16).

Palaeoenvironment and palaeoceanography

Pelagic microfossils in the NKK-1 samples imply strong influence of open ocean surface waters during the late Eocene to early Oligocene on the southern Java margin. The high planktonic: benthic foraminifera ratio and general benthic foraminifera fauna suggest a water depth >400 m (at least upper bathyal depths); that is, below the seasonal thermocline (Keij 1964; van Gorsel et al. 2014).

Given the regional tectonic activity and heterogeneity of early Cenozoic Eocene marine sequences on and around Java (Lunt 2013, 2019), this probably reflects the depth of a shelf sub-basin. The often greenish colour of the sediments and high pyrite content are consistent with limited oxygen, reducing conditions in the sediment pile and thus some degree of basin restriction at the seafloor, although persistent bioturbation implies that bottom waters were never anoxic. Common radiolaria in NKK samples imply abundant silica supply to surface waters, possibly connected to local upwelling (see below), fuelling high surface production and export to the seafloor.

Planktonic foraminifera assemblages are diverse and contain typical tropical elements with variability in the test size and dominance of key genera between samples that are suggestive of shifts in water mass structure, possibly on orbital timescales unresolved by our sampling. For example, dentoglobigerinids appear to show variability in test size through the Oligocene, T. cerroazulensis fluctuates and Hantkenina spp. become common only in the upper part of Zone E16. Dentoglobigerina galavisi and Subbotina spp. are the more consistent elements. Turborotalia ampliapertura is common in the upper Eocene and basal Oligocene, providing a useful surface ocean tracer for palaeoceanographic studies. In general, diversity appears to be lower in ‘glassy’ samples, possibly because of high clay dilution, although other environmental factors might be at play. For instance, foraminiferal preservation tends to be poorer (more carbonate-rich ‘frosty’ preservation) where there are more hantkeninids, whereas relatively few hantkeninids occur in glassy more clay-rich samples.

Our stable isotope records are of insufficient resolution to offer much direct insight into the cause of the planktonic foraminifera variability but consideration of modern Java oceanography provides some clues. Hydrological conditions along the modern Java coast are complex, with high interannual variability driven by an interplay of (1) remote oceanic forcing owing to annually reversing monsoon winds in the Northern Indian Ocean, (2) local coastal winds on the Java margin, also seasonally reversing with the monsoon and driving upwelling, and (3) Indonesian Throughflow (Clarke and Liu 1993; Valsala and Maksyutov 2010; Utari et al. 2019) (see Fig. 1). Asian monsoon climates have been simulated back to Eocene time (Huber and Goldner 2012), which, with a then expanded Western Pacific–Indonesian Warm Pool, with high sensitivity to El Niño
forcing (Carlson and Caballero 2017; Hutchinson et al. 2018), would perhaps drive significant variability, on orbital timescales, in stratification, mixed layer depth and precipitation offshore Java, strongly affecting plankton and sedimentation in the Nanggulan basin. Future comparisons between Tanzania and Java records will be useful for testing hypotheses about late Eocene monsoon forcing in the Indo-Pacific. Similarly, quantitative data on NKK-1 planktonic foraminifera are needed to compare with evidence for systematic evolutionary turnover among calcareous nanofossils between these two sides of the Indian Ocean (Wade and Pearson 2008; Jones et al. 2019).

**NKK-1 δ¹⁸O and δ³¹C EOT stable isotope stratigraphy: challenges and initial insights into EOT ocean–climate change on the Java margin**

Generally, in the EOT part of the NKK succession there is glassy or near glassy preservation of foraminifera; however, for the stable isotope analysis, effort was often needed to separate ‘good’ calcite fragments from sparite infill. The finding of distinct offsets between analyses of whole filled shells, separated fragments and isolated sparite infill showed that in much of the core this approach worked, and we believe that near original δ¹⁸O signatures have been captured for the late Eocene up to 33.9 Ma, and from 33.4 Ma and younger (Figs 14 and 15). Our new records, therefore, provide the first documentation of EOT δ¹⁸O and δ³¹C shifts in Java.

In the basal part of Zone O1, just above the E/O boundary, strong negative excursions in δ¹⁸O and δ³¹C complicate the picture. Here even isolated fragments are affected. This is interesting, as there is no obvious change in preservation quality of the shell material; that is, no clear evidence of neomorphic crystallites replacing original shell calcite. Why this would be restricted to basal Zone O1 is not entirely clear. We cannot entirely rule out that the negative isotopic excursions are contemporaneous with the early Oligocene deposition of the sediments. However, a more probable explanation is that this interval was affected by porewater with an isotopically depleted composition, probably carbonate-rich meteoric waters that influenced the EOT section post uplift; that is, many millions of years later. No obvious lithology change is apparent. We do, however, find higher abundance of deep-water agglutinated benthic foraminifera, which have relatively large sand-sized tests (0.5–1 mm), suggesting a facies change. It is possible that the associated increase in grain size increased permeability (although data are lacking to substantiate this), providing a conduit for meteoric groundwater flow, a phenomenon recently investigated in the Nanggulan Formation in relation to identifying and utilizing natural springs as water supplies (Winarti et al. 2020).

The δ¹⁸O composition of rainwater in Java today is around −5.7‰, and Holocene and late Pleistocene speleothem carbonate δ¹³C is −4 to −7‰ VPDB and +2 to −12‰ VPDB for δ¹³C (Westaway et al. 2007), which is similar. The isotopic composition of even the sparry calcite becomes highly depleted in two samples from the basal Oligocene, with compositions of c. −10 to −7‰ δ¹⁸O and c. −6 to −5‰ δ¹³C respectively being recorded. In this case sparry calcite perhaps penetrates the pores of the foraminifera and attaches in such a way that the fragmentation approach did not remove all contamination. A simple mixing-relationship calculation (sparry calcite value of −10.3‰ and ‘average original’ test calcite value of 4.6‰) implies that 60–70% of the sample would have to be replaced to shift the sample isotopic values by as much as we see. Based on our few SEM images of shell fragments from this interval this looks plausible considering the highly porous nature of *T. ampliapertura* tests, providing abundant space for crystalline infill. It is also possible that our cleaning process failed to remove stubborn adhering layers of sparite (Fig. 5c). The fact that benthic foraminifera, which are highly unporous compared with *T. ampliapertura*, do not acquire such low δ¹³C values supports this suggestion. This problem would benefit from further study.

Nevertheless, by excluding the values with anomalously low values we present a best-case δ¹⁸O and δ¹³C stratigraphy (Fig. 15) and interpret it in terms of EOT climate and environmental change. Broadly, we find a familiar pattern and timing of increased δ¹⁸O and δ¹³C in both planktonic and benthic foraminifera across the EOT, recording a global increase in oceanic water δ¹⁸O associated with the growth of continental ice on Antarctica around the EOIS, and accompanying feedbacks in the carbon system (Zachos et al. 1996; Coxall et al. 2005; Hutchinson et al. 2021).

The overall lower δ¹⁸O values (by c. 2‰) of NKK-1 benthic foraminifera compared with the equatorial Pacific (Fig. 14) are consistent with the intrashelf sub-basin environment for Nanggulan sediments, involving some degree of restriction, and thus warmer seafloor temperatures. Moreover, the Nanggulan seafloor may have been bathed in an intermediate depth Indian Ocean water mass (400–1000 m), compared with the much deeper Site 1218 (c. 3500 m palaeodepth); potentially, warm saline Tethyan deep water as for Indian Ocean drill-holes on Ninety East Ridge (Zachos et al. 1992). The magnitude of δ¹⁸O change across the NKK-1 EOT is also larger than typical; that is, c. 2.5‰ change compared with 1–1.7‰ increase elsewhere (Coxall and Wilson 2011; Hutchinson et al. 2021). This suggests local changes in temperature, salinity and/or water masses in Java, although it is possible that diagenetic biases also play a role. That NNK-1 benthic δ¹⁸O and δ¹³C are systematically lower in the late Eocene and more similar in the early Oligocene suggests some difference either in the degree of basin connectivity (better connected to the open ocean) or influence of a different water mass in the Nanggulan basin in the early Oligocene (Fig. 14).

Planktonic foraminifera δ¹⁸O is especially low, even compared with Tanzania, with values between −4 to −5‰ in the late Eocene, compared with −3.7 to −3.3‰ in Tanzania at correlative levels (Pearson et al. 2008). An isotopic study of middle and late Miocene foraminifera from similarly preserved material from the nearby Sambipitu and Oyo Formations of central Java found planktonic foraminiferal δ¹⁸O of −3 to −4.6‰, occasionally falling to −5‰, implying high equatorial temperatures and thus a western position of the Warm Pool well into the Miocene (Akmaluddin et al. 2010). Reconstruction of sea surface temperatures is beyond the scope of this study. However, a general estimate based on typical planktonic foraminifera δ¹⁸O–temperature relationships (Bemis et al. 1998) would imply SSTs 2–6°C warmer in Java than in Tanzania. This initial estimate is consistent with sparse existing data for the middle Eocene (39 Ma) based on clumped isotopes in larger benthic foraminifera from Nanggulan and Mg/Ca in planktonic foraminifera from Tanzania (reconstructed SSTs of 35–36°C and 32°C respectively), as well as model predictions of a late Eocene Indo-Pacific ‘super-Warm Pool’ (Pearson et al. 2001; Evans et al. 2018; Hutchinson et al. 2018) (Fig. 2). Future studies will be important for testing this idea using a variety of SST proxy approaches, including further planktonic foraminiferal δ¹⁸O analysis carefully constrained for local δ¹⁸O seawater effects, clumped isotopes and Mg/Ca in planktonic foraminifera, as well as organic biomarkers, including TEX86, to be compared with existing tropical EOT constraints from Tanzania and elsewhere (Leat et al. 2008; Cramwick et al. 2018; Evans et al. 2018). Invasive analytical techniques, including laser ablation and the ion microprobe (secondary ion mass spectrometry) could be important for overcoming problems associated with foraminifera diagenetic infills.

Comparison of our EOT isotopic records with benthic foraminifera and nanofossil assemblage data provide interesting insights. Assemblage data show major disruption in phytoplankton communities from the uppermost Eocene of NKK-1 (Jones et al. 2019). Importantly, a major decline in the abundance of key groups resulting in a so-called *Discoaster* Extinction Event was identified.
‘Group 1A’ and ‘Group 1B’, replotted in Figure 15, represent ecologically meaningful nannofossil species clusters identified by principal component analysis; Group 1A consists of Reticulofenestra reticulata, R. isabellae, R. nanggulanensis, Discoaster saipanensis, D. barbadiensis and Umbilicosphaera protoannula; Group 1B consists of Reticulofenestra bisecta, Sphenolithus akropodas and S. tribulosus (Jones and Dunkley Jones 2020). The replacement of Group 1A by Group 1B is interpreted as indicating a shift from strongly oligotrophic to more mesotrophic conditions centred around 34.7–34.5 Ma (Jones et al. 2019). The change was interpreted to signal increased nutrient supply to the tropical surface ocean 0.5–1 myr prior to the EOT. The parallel spikes in deep-water agglutinated benthic foraminifera found here (Fig. 15) suggest a contemporaneous response at the seafloor, possibly related to changes in surface sediment export or food quality, reduced bottom water oxygenation, or taphonomy more generally (Kaminski and Ortiz 2014).

Weakening of ocean stratification allowing increased mixing of subsurface nutrients into the surface ocean could drive such a change in surface and deep-water conditions; however, late Eocene
$\delta^{18}$O shows rather stable $\delta^{18}$O of mixed-layer living $T$. amphiaptura, with a constant offset from benthic foraminifera, implying stable thermal stratification on the southern Java margin through the EOT. Benthic $\delta^{13}$C shows some changes that may be significant. Figure 15 identifies a c. 500 kyr phase of globally negative $\delta^{13}$C between 34.4 and 33.9 Ma in the NKK-1 record, which has been found in many deep-sea EOT records (Coxall et al. 2018). This excursion has been linked to a global reorganization of overturning ocean circulation and associated changes in ocean nutrient distributions in the 1 myr preceding the EOT. That this phenomenon coincides with the Discosteer extinction event and nannofossil turnover suggests that the two are related in some way, consistent with the interpretation of previous studies of nannofossil assemblages from the Tanzanian margin (Dunkley Jones et al. 2008). One possibility is that phytoplankton evolution was driven by changes in the chemistry and nutrient content of global subsurface waters advected and upwelled to the photic zone on the southern Java margin. ODP Site 1218 benthic stable isotope records are relevant in this respect as the then-open Indonesian Seaway would allow unhindered equatorial flow, including flow of upwelling equatorial waters that could affect surface waters of the Indonesian archipelago. That the biotic changes preceded the largest phase of Antarctic ice sheet growth (EOIS) provides further support for the proposal that changes in the ocean carbon system played a role in priming the Earth for the shift into the icehouse.

**Volcanic rocks**

Identification of a basalt flow and ash beds at c. 39 mbs in the NKK-1 borehole is novel to studies of the Nanggulan section as far as we are aware. Figure 17 attempts to summarize the volcanic history in west central Java, before, during and after deposition of sediment sampled by the NKK-1 core. Intrusive dykes of middle Oligocene age have been reported in central Nanggulan, but these were based on K/Ar ages whereas more recent U/Pb measurements indicate a latest Oligocene age (25.0 Ma (Smyth et al. 2003); Sample KS in Fig. 17; 28.31 ± 3.46 Ma (Soeria-Atmadja et al. 1994)). The planktonic foraminifera and calcareous nannofossil biozonation at the basalt flow and associated tuffaceous sandstone in NKK-1 constrains the age of this eruption to early Oligocene Zones O1–O2 and Zone CNO3 (NP23–NP22 below, NP23–24 above) respectively, with an implied age range of 30.5–33 Ma. The Menoreh Hills are a series of three ‘Old Andesite’ volcanic stubs (early Oligocene to early Miocene) eroded down to their crystalline cores and surrounded by the remnants of proximal volcaniclastic deposits. The recognized volcanoes are Gajah (29.6 ± 1.48 Ma) in the centre, Ijo (25.4 ± 1.27 Ma) in the south, and youngest, Menoreh (22.6 ± 1.13 Ma), with radiometric ages given by Bellon et al. (1989) and Soeria-Atmadja et al. (1994) (Fig. 17). The NNK-1 basalt is thus probably part of the Gajah lavas complex (Fig. 13), and may be the earliest evidence for its onset, recording early Oligocene arc volcanism on the Sunda margin. Small outcrops of mid-Oligocene pillow lavas are also known 45 km to the east of the borehole in Bayat area, in the Kebo-Butak Beds. These are as old as 29.3 ± 1.4 Ma(Smyth et al. 2003) and are dated from interbedded foraminifera-bearing clays as Zone P22–O6 to early O7 (Lunt 2013a). Radiometric dating and major element chemistry are needed to constrain how the NKK-1 lava compares with other contemporary volcanic emplacements (Smyth et al. 2005). In west Java the early Oligocene is characterized by tholeiitic basalts and volcaniclastic rocks, proposed to have been associated with early rifting, compared with calc-alkaline deposits in the Miocene (Soeria-Atmadja et al. 1994; Sukarma et al. 1994; Lunt 2013a). Such distinctly dark-coloured rocks have not been recognized on field evidence in central Java, although detailed geochemistry is lacking. In both west and central Java volcanic rocks are intermittent features in the early Cenozoic sedimentary record but in earliest Miocene times they become dominant (‘First Breccia’ formation, Fig. 17).

The middle Eocene interbedded tuffaceous sands and clays in Cores 68 (88 mbs) and below (Unit VI, Jetis Beds of Lunt 2013a), identify an earlier acme of volcanism (Fig. 17). This unit correlates to a deep marine ash-tuff, c. 10 m thick as observed in the Gamping Barat core drilled 16 km SE of NKK-1 (Lunt 2013a). Our biostatigraphic suggests that these beds in NKK-1 have an age of roughly 39–41 Ma, which is consistent with the oldest known magmatic activity of the ‘Old Andesite’ volcanic phase, dated radiometrically in pillow basalts from the Pacitan area (130 km ESE of NKK-1) to 42 Ma, although with a very large error (±9.78 Ma; annotated in Fig. 17; Soeria-Atmadja et al. 1994). Minor tufts are observed in the Watu Puru Beds of about the same age (Oppenoorth and Gerth 1929; Lunt 2013a). These all correspond to the less well-documented ‘Pre-Gajah’ volcanic rocks of Widagdo et al. (2018). In Nanggulan outcrops and the NKK-1 core the sediments show increasing volumes of volcaniclastic rocks eventually drowning the abundant pelagic supply found in the Watu Puru Beds, resulting in the Jetis Beds and equivalents. This first acme of middle Eocene volcanism ended at the time of the hiatus identified in this study. Although tuffaceous beds are not observed again until the early Oligocene, there is minor volcanic material in most of our foraminiferal preparations from the Tegalsari Marl. Previous X-ray diffraction analyses revealed that plagioclase predominates over quartz in silts from these sediments(Lunt 2013a, fig. 96a), implying a proximal volcanic source and minor eruptions on Java throughout the middle and late Eocene. It is important to emphasize that Eocene and early Oligocene volcanic rocks are minor compared with the subsequent Neogene episodes, and are dominated by distal tufts. In some regions tholeiitic eruptions are evident and unrelated to normal arc processes. Only in the mid-Oligocene did significant volcanism begin, with dozens of active volcanoes producing extremely thick proximal volcanic breccias and pyroclastic deposits as part of the Old Andesite system. From 20 to 12 Ma there was a decline in volcanic activity (Oppenoorth and Gerth 1929; Bellon et al. 1989; Lunt 2013a). The Old Andesite phase ended, followed by marine transgression producing renewed limestone deposition across the length of Java over the slowly subsiding and eroding volcanic stubs. Around 12 Ma influx of fresh volcaniclastic deposits across the north of Java signals a jump in the volcanic axis and onset of the modern phase of arc volcanism.

**Conclusions**

The 2006 Nanggulan drilling campaign successfully recovered marine sedimentary and igneous rock samples useful for reconstructing climate and environmental changes on the Java margin during the Eocene–Oligocene climate transition and for providing insights into Paleogene volcanic activity along the Sunda subduction zone. From lithological, micropalaeontological and geochemical analyses we reach the following conclusions.

1. (1) The 100 m long NKK-1 borehole recovered late Oligocene (planktonic foraminifera Zone O5, nannofossil Zone CNO4–O5) to middle Eocene (Zone E13, CNE15) marine sediments rich in, at times, well-preserved (often glassy) calcareous microfossils with variable volcanic inputs. The contained microfossils provide important biostratigraphic age constraints, which allow us to recognize a complete sequence across the EOT.

2. (2) Our studies confirm the perspectives of Lunt (2013a) showing that the Nanggulan Formation extends into the Late Oligocene, which is younger than in many previous studies.
Using the lithostratigraphic nomenclature presented by Lunt (2013a) we recognize at least 82 m of chalky marls, claystones and silty clays referred to as the Tegalasari Marls. This is separated by a hiatus from alternating volcanic sands and silty claystones, in which the hole terminates (Jetis Beds). This unconformity correlates to other outcrops in central Java, and perhaps a wider area across southern Sundaland.

Marine microfossils in the recovered section are fully pelagic. Benthic foraminifera assemblages and counts imply a relatively deep (>400 m water depth) intrashelf basin throughout EOT time.

The previously found ‘Pellatispira’ limestone does not occur in the new borehole, implying that this is a locally occurring phenomenon. It was probably a thin and inconsistent debris flow occurring above an unconformity.

Despite glassy preservation of planktonic foraminifera, sparry calcite infilling is common at some levels, especially in the basal Oligocene, and requires special preparation techniques for geochemical methods. Separation of test calcite and infill, as we tested here, is one option, and laser ablation or ion microprobe-style analyses will be required to extract data from the intervals found here to be most problematic.

Our new planktonic and benthic stable isotope records provide the first documentation of EOT δ18O and δ13C shifts in the Indo-Pacific. The low-resolution records provide coherent signals of EOT climate change comparable with global records, although interrupted by a phase of anomalously depleted δ18O and δ13C in the basin Oligocene where diagenetic calcite biases measurements.

Absolute δ18O values of NKK-1 benthic foraminifera are c. 2‰ VPDB lower than for the equatorial Pacific and other global deep-sea records (−1‰ compared with +1‰ at Site 1218), reflecting local conditions or an intermediate Indian Ocean water mass affecting the Java margin. The magnitude of change across the EOT is also larger than typical; that is, c. 2.5‰ change compared with 1–1.7‰ increase elsewhere. This suggests local changes in temperature, salinity and/or water masses on the SE Java margin.

The NKK-1 Nanggulan borehole represents one of the few EOT sections world-wide having both a stable isotope stratigraphy and the epoch-defining hantkeninid planktonic foraminifera extinction. This is necessary for exploring synchronicity of climatic and biotic change both in Java and on the other side of the Indian Ocean in Tanzania, a gradient across which even monsoon signals might be detectable. Higher resolution sampling for foraminiferal and geochemical analysis is under way to address these questions.

Late Eocene turnovers in calcareous nannofossils, coincident with a spike in deep-water agglutinated foraminifera, suggest increased nutrient availability to the surface ocean on the Java margin. Although local influences cannot be ruled out, the coincidence of the Javanese nannofossil changes with contemporaneous nannofossil turnover on the Tanzanian Indian Ocean margin, as well as globally synchronous adjustments in deep water δ13C, suggest global changes in circulation and carbon cycling at the end of the Eocene that may have played a role in bringing down the Eocene greenhouse.

A c. 10 m thick sequence of basaltic lava and volcanic ash recovered in the borehole is constrained by biostratigraphy to planktonic foraminifera Zones O1–O2 (nannofossil Zone CNO3), with an age of 30.5–33 Ma. This overlaps with published radiometric ages for some of the earliest volcanic activity related to Oligo-Miocene Sunda margin subduction and demonstrates Old Andesite volcanic activity from the early Oligocene rather than the late Oligocene as commonly stated.

Future work will focus on using the Java planktonic foraminifera for reconstructing surface temperatures and other properties, including, for example, the use of foraminiferal Mg/Ca for constraining the mixed signals of temperature and salinity, and potential contributions from diagenetic calcite. Future investigations of organic-walled microfossils and derived organic biomarker SST reconstructions will also be important in this respect. Initial insights into Eocene–Oligocene palaeoceanography suggest that the Indonesian Warm Pool was warmer than today and was then the warmest oceanic region of the Earth.

Acknowledgements We thank Gadjah Mada University for supporting a reconnaissance survey, B. Seubert of Perteppe for logistical support in Jakarta and Bandung, and P. Markwick for providing rotated palaeogeographical coordinates for the borehole location.

Author contributions HKC: conceptualization (supporting), data cur- ation (lead), formal analysis (lead), investigation (lead), methodology (equal), visualization (lead), writing – original draft (lead), writing – review & editing (lead); TDJ: formal analysis (supporting), investigation (supporting), writing – review & editing (supporting); APJ: formal analysis (supporting), investigation (supporting), visualization (supporting), writing – review & editing (supporting); PL: conceptualization (supporting), methodology (supporting), resources (supporting), validation (lead), visualization (supporting), writing – original draft (supporting), writing – review & editing (supporting); IM: formal analysis (supporting), investigation (supporting), visualization (supporting), writing – original draft (supporting); CJN: formal analysis (supporting), investigation (equal), visualization (supporting), writing – review & editing (supporting); AO: investigation (supporting); EP: investigation (supporting), writing – review & editing (supporting); PS: investigation (supporting), methodology (supporting), resources (supporting); WR: resources (supporting), validation (supporting); PNP: conceptualization (lead), formal analysis (supporting), funding acquisition (lead), investigation (equal), writing – original draft (supporting), writing – review & editing (supporting).

Funding This study was funded primarily by UK Natural Environment Research Council (NERC) Small Grant NE/C514523 ‘Eocene–Oligocene climate change in Indonesia’ to P.N.P. H.K.C.’s participation and the foraminifera stable isotope and SEM analysis were supported by a Royal Society University Research Fellowship awarded to H.K.C. Calcaneous nannofossil research by A.P.J. was funded by the NERC CENTA Doctoral Training Programme. E. Piga was supported by an NERC GW4+ Ph.D. Research Grant (NE/L002434/1).

Data availability All data generated or analysed during this study are included in this published article (and its supplementary information files). The geochemical data are also available at the Bolin Centre for Climate Research Database: https://doi.org/10.17043/coxall-2021-java-1

Scientific editing by Rosalie Tostevin

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