

1 **Southern high-latitude terrestrial climate change during the Paleocene–**
2 **Eocene derived from a marine pollen record (ODP Site 1172, East**
3 **Tasman Plateau)**

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26 **ABSTRACT**

27 Reconstructing the early Paleogene climate dynamics of terrestrial settings in the high
28 southern latitudes is important to assess the role of high-latitude physical and
29 biogeochemical processes in the global climate system. However, whereas a number
30 of high-quality Paleogene climate records has become available for the marine realm
31 of the high southern latitudes over the recent past, the long-term evolution of coeval
32 terrestrial climates and ecosystems is yet poorly known. We here explore the climate
33 and vegetation dynamics on Tasmania from the middle Paleocene to the early Eocene
34 (60.7 – 54.2 Ma) based on a sporomorph record from Ocean Drilling Program (ODP)
35 Site 1172 on the East Tasman Plateau. Our results show that three distinctly different
36 vegetation types thrived on Tasmania under a high-precipitation regime during the
37 middle Paleocene to early Eocene, with each type representing different temperature
38 conditions: (i) warm-temperate forests dominated by gymnosperms that were
39 dominant during the middle and late Paleocene; (ii) cool-temperate forests dominated
40 by southern beech (*Nothofagus*) and araucarians across the middle/late Paleocene
41 transition interval (~59.5 to ~59.0 Ma); and (iii) paratropical forests rich in ferns that
42 were established during and in the wake of the Paleocene–Eocene Thermal Maximum
43 (PETM). The transient establishment of cool-temperate forests lacking any frost-
44 sensitive elements (i.e., palms and cycads) across the middle/late Paleocene transition
45 interval indicates markedly cooler conditions, with the occurrence of frosts in winter,
46 on Tasmania during that time. The integration of our sporomorph data with previously
47 published TEX₈₆-based sea-surface temperatures from ODP Site 1172 documents that
48 the vegetation dynamics on Tasmania were closely linked with the temperature
49 evolution in the Tasman sector of the Southwest Pacific region. Moreover, the
50 comparison of our season-specific climate estimates for the sporomorph assemblages

51 from ODP Site 1172 with the $\text{TEX}_{86}^{\text{L}}$ - and $\text{TEX}_{86}^{\text{H}}$ -based temperature data suggests a
52 warm-season bias of both calibrations for the early Paleogene of the high southern
53 latitudes.

54

55 **1 Introduction**

56 The Southern Ocean is an important region for early Cenozoic (65–34 Ma) climates,
57 being the dominant region for deep-water formation during that time (Thomas et al.,
58 2003; Sijp et al., 2011; Hollis et al., 2012). South Pacific sea-surface and global
59 intermediate water temperatures increased from the late Paleocene to the early
60 Eocene, with maximum warmth during the Early Eocene Climatic Optimum (EECO;
61 53–51 Ma), followed by a cooling trend during the middle and late Eocene (Zachos et
62 al., 2001; 2008; Bijl et al., 2009; Hollis et al., 2012). This cooling trend ultimately
63 culminated in the establishment of a continental-scale ice shield on Antarctica during
64 the earliest Oligocene (e.g., Zachos et al., 1994; 2008; Barrett, 1996), which
65 represented a decisive step in the Earth’s transition from a “greenhouse” into an
66 “icehouse” world.

67 Organic geochemical surface-water temperature proxy records from the high-latitude
68 Southwest Pacific Ocean (notably TEX_{86} ; Bijl et al., 2009) closely mirror trends in the
69 benthic foraminiferal oxygen isotope data from the late Paleocene to the early
70 Oligocene (Zachos et al., 2001; 2008), which lends further support to the suggestions
71 that the southern ocean was the main region for deep-water formation (Thomas et al.,
72 2003). Irrespective of the calibration used, Southwest Pacific TEX_{86} -derived sea-
73 surface temperatures (SSTs) were relatively cool during the early and middle
74 Paleocene. During the late Paleocene and early Eocene, SSTs gradually rose to
75 tropical values ($>26^{\circ}\text{C}$), with maxima being reached during the Paleocene-Eocene

76 Thermal Maximum (PETM; Sluijs et al., 2011) and the EECO (Bijl et al., 2009;
77 Hollis et al., 2009; 2012). Towards the end of the early Eocene (49–50 Ma), a
78 pronounced SST cooling of $\sim 4^{\circ}\text{C}$ occurred on the Australo-Antarctic margin; this
79 cooling has been attributed to the onset of westbound surface-water throughflow
80 across the Tasmanian Gateway (Bijl et al., 2013a). A similar cooling trend is
81 registered for surface waters off New Zealand (Hollis et al., 2009; 2012) and in the
82 Tasman sector of the Southwest Pacific Ocean (Bijl et al., 2009; Hollis et al., 2012).
83 Strikingly low SSTs are recorded for the high-latitude Southwest Pacific Ocean
84 during the interval spanning the middle/late Paleocene transition (59.5–59.0 Ma);
85 TEX₈₆-derived SST decreased by $\sim 3^{\circ}\text{C}$ during that time (Bijl et al., 2009; Hollis et al.,
86 2012; 2014). Along with this SST drop, lowered sea level and marked bathyal erosion
87 suggest that a transient growth of an Antarctic ice sheet may have occurred (Hollis et
88 al., 2014).

89 While an increasing amount of data has become available on the marine climate
90 evolution in the southern high latitudes during the early Paleogene, the coeval
91 terrestrial climate dynamics of that region are yet poorly documented (e.g., Passchier
92 et al., 2013). For the Paleocene, paleobotanical records reflect the thriving of
93 temperate, gymnosperm-rich forests dominated by podocarps in Southeast Australia
94 (e.g., Macphail et al., 1994; Greenwood et al., 2003; Greenwood and Christophel,
95 2005), on Seymour Island (Antarctica; Askin, 1990), and in New Zealand
96 (Mildenhall, 1980; Raine et al., 2009).

97 In contrast, highly diverse angiosperm forests containing taxa that today are restricted
98 to tropical environments characterize the early Eocene vegetation in the higher-
99 latitude Southwest Pacific region. This vegetation is widely known from Southeast
100 Australia (Macphail et al., 1994; Greenwood et al., 2003; Greenwood and

101 Christophel, 2005), Tasmania (Truswell, 1997; Carpenter et al., 2012) and New
102 Zealand (Raine et al., 2009; Handley et al., 2011); notably, it also thrived on the
103 Wilkes Land margin of the Antarctic continent, i.e., at paleo-latitudes of $\sim 70^\circ$ South
104 (Pross et al., 2012; Contreras et al., 2013).

105 The pronounced vegetation turnover from temperate forests during the Paleocene to
106 near-tropical forests during the early Eocene suggests a marked, climatically driven
107 change in terrestrial environments in the high southern latitudes. However, the
108 transitional process between these two vegetation types and the underlying change in
109 terrestrial climate conditions of the Southwest Pacific region have remained poorly
110 understood. The gaps in the documentation of terrestrial climate dynamics during the
111 early Paleogene appear particularly pronounced considering the coeval datasets for
112 the marine realm from the same region (Bijl et al., 2009; 2013a; Hollis et al., 2009;
113 2012; 2014). This is due to the fact that prior paleobotanical studies are mainly based
114 on stratigraphically discontinuous outcrops that provided only limited insights into the
115 Paleocene and early Eocene vegetation of the region (Greenwood et al., 2003). In
116 addition, available studies on sporomorphs from the early Paleogene of the Southwest
117 Pacific region have predominantly focussed on the taxonomical characterization of
118 the assemblages and the generation of biostratigraphic schemes (e.g., Stover and
119 Evans, 1973; Stover and Partridge, 1973; Truswell, 1997; MacPhail, 1999). In any
120 case, estimates of terrestrial temperatures in the high southern latitudes and a
121 comparison with the currently available SST estimates from the marine realm can
122 strongly enhance the understanding of the climate evolution in the high southern
123 latitudes during the early Paleogene.

124 In light of the above, we here explore the vegetation response to high-southern-
125 latitude climate forcing from the Paleocene to the “hothouse” conditions of the early

126 Eocene based on a new, chronostratigraphically well-calibrated (Bijl et al., 2013b)
127 sporomorph record from Ocean Drilling Program (ODP) Site 1172 off eastern
128 Tasmania. We quantitatively evaluate the compositional variations of the sporomorph
129 assemblages and carry out quantitative sporomorph-based paleoclimatic
130 reconstructions. For a further assessment of the terrestrial climatic conditions in the
131 Southwest Pacific region, we apply the same paleoclimate reconstruction approach to
132 previously published sporomorph records from Southeast Australia (Bass Basin,
133 Gippsland Basin, Southeast Highlands) and New Zealand. Finally, we compare our
134 terrestrial paleoclimate estimates with previously published $\text{TEX}_{86}^{\text{L}}$ - and $\text{TEX}_{86}^{\text{H}}$ -
135 based SST reconstructions from the same site (Bijl et al., 2009; 2013b; Hollis et al.,
136 2014) in order to contribute to a better understanding of the early Paleogene climate
137 dynamics in the high southern latitudes.

138

139 **2 Material and methods**

140 The middle Paleocene to early Eocene strata studied here were recovered at ODP Site
141 1172, which is located ~100 km east of Tasmania on the western side of the East
142 Tasman Plateau (ETP; 43°57.6' S, 149°55.7' E; Fig. 1) (Shipboard Scientific Party,
143 2001b).

144

145 **2.1 Regional setting and paleoceanography**

146 During the early Paleogene, Tasmania and the ETP were located at ~65°S, much
147 closer to Antarctica than today (Exon et al., 2004b). Paleoceanographic patterns as
148 determined by winds and gateway configuration were likely vital for regional climates
149 on land (Sijp et al., 2011). The study site was located close to the Tasmanian
150 promontory, which hampered deep ocean exchange between the Southwest Pacific

151 and the Australo-Antarctic Gulf for most of the early Paleogene (Shipboard Scientific
152 Party, 2001b; Stickley et al., 2004; Fig. 1). During the Paleocene and early Eocene,
153 the Tasman region was under the persistent influence of the Antarctic-derived Tasman
154 Current; in contrast, the Australo-Antarctic Gulf (west of Tasmania) was bathed by
155 the low-latitude-derived Proto-Leeuwin Current (Huber et al., 2004; Sijp et al., 2011;
156 see Fig. 1). This paleoceanographic configuration determined marine biogeographical
157 patterns in the region (Huber et al., 2004; Bijl et al., 2011; 2013a). The onset of the
158 deepening of the Tasmanian Gateway at ~49–50 Ma initiated a westbound Antarctic
159 Counter Current flowing along the Antarctic margin from the Pacific into the
160 Australo-Antarctic Gulf (Bijl et al., 2013a). Continued rifting through the Eocene and
161 accelerated deepening of the Tasmanian Gateway (~35.5 Ma) led to the inflow of
162 Australo-Antarctic Gulf waters through the Tasmanian Gateway into the southern
163 Pacific during the early Oligocene (Stickley et al., 2004; Sijp et al., 2011). **These**
164 **paleoceanographic reorganizations had important effects on terrestrial climates in the**
165 **Australo-Antarctic region (Bijl et al., 2013a).**

166

167 **2.2 Age model, lithology and depositional environment**

168 We here follow the corrected sample depths for ODP Hole 1172D as published in
169 Sluijs et al. (2011) based on detailed correlation of the X-ray fractionation core
170 scanning to the γ -ray downhole log. The age model of the studied sequence is based
171 on the magnetostratigraphy, chemostratigraphy and dinoflagellate cyst (dinocyst)
172 biostratigraphy as presented in Bijl et al. (2013b). For the interval studied (60.7–54.2
173 Ma), the age model is based on three magnetic reversals confidently correlated to the
174 Geomagnetic Polarity Time Scale (GPTS) of Vandenberghe et al. (2012), the PETM
175 (~56 Ma), which exhibits a negative carbon isotope excursion of ~3‰ between

176 611.89 and 611.86 mbsf (revised meters below sea floor) (Sluijs et al., 2011), and
177 several dinocyst first and last occurrence data that have been calibrated regionally
178 following Crouch et al. (2014). The magnetostratigraphic age model for the Paleocene
179 section (Röhl et al., 2004) was adjusted by Bijl et al. (2010) on the basis of the
180 recognition of the PETM in Core 1172D-15R (611.8 mbsf; Sluijs et al., 2011). The
181 missing interval between Cores 16R and 17R (~620 mbsf) represents a ~1.3-Ma-long
182 hiatus that spans the time interval correlative to between infra-Subchrons C26n and
183 C25n (57.7–59 Ma; Bijl et al., 2010; 2013b).

184 The upper 60 cm of section in Core 17R (i.e., below the hiatus) are heavily disturbed
185 and have many sediment characteristics more consistent with Core 16R than with the
186 underlying sediment (Röhl et al., 2004). We therefore follow Röhl et al. (2004) in
187 their suggestion that this interval represents caved material of late Paleocene age.

188 The Middle Paleocene to lower Eocene succession of ODP Site 1172 consists mainly
189 of gray to grayish brown clay- and siltstones with low abundances of calcareous and
190 siliceous microfossils (Shipboard Scientific Party, 2001a; Röhl et al., 2004).

191 Environmentally, the succession is interpreted to reflect very shallow to restricted
192 marine conditions, with marked runoff from the nearby shores (Röhl et al., 2004).

193 Any study on sporomorphs from marine sediments critically relies on the
194 identification of the source region in order to provide meaningful paleoclimatic
195 information. An in-depth discussion of this issue is provided in Section 3.2
196 (“Constraints on sporomorph source region”).

197

198 **2.3 Sample processing and data analysis**

199 Eighty-nine samples from the Paleocene and Eocene of ODP Site 1172, originally
200 processed at the Laboratory of Palaeobotany and Palynology, Utrecht University (Bijl

201 et al., 2011), were here reanalysed for terrestrial palynomorphs. Sample processing
202 followed standard palynological techniques, including treatment with HCl (10%) and
203 HF (38%) and sieving through a 15 µm nylon mesh (e.g., Pross, 2001). The
204 microscope slides are stored in the collection of the Laboratory of Palaeobotany and
205 Palynology at Utrecht University. Whenever possible, 300 sporomorphs (excluding
206 reworked specimens) were analysed per sample and determined to the species level;
207 this required the analysis of up to six slides per sample. The analyses were performed
208 using a light microscope at 200x magnification; morphological details were studied
209 with a magnification of 1000x. Sporomorph identifications are mainly based on
210 Couper (1960), Harris (1965), Stover and Partridge (1973), Truswell (1983), Raine et
211 al. (2008), and Truswell and Macphail (2009). The botanical affinities are given
212 following Macphail et al. (1994), Raine (1998), and Truswell and Macphail (2009).
213 All sporomorph data (including photomicrographs of key taxa) are provided in the
214 Supplementary Information.

215 Rarefaction was applied to evaluate sporomorph diversity; this allows to estimate the
216 number of sporomorph species at a constant sample size (Raup, 1975). Detrended
217 Correspondence Analysis (DCA) is an ordination technique that was used to analyse
218 floral composition change through time.

219 To constrain the source region of the sporomorphs from ODP Site 1172, we compared
220 the floristic composition of our record with coeval sporomorph records from
221 Southeast Australia (Bass Basin – data from Blevin 2003; Gippsland Basin –
222 Department of Primary Industries, 1999; Southeast Highland sections – Taylor et al.,
223 1990) and New Zealand (Middle Waipara and Otaio River sections – Pancost et al.,
224 2013) using DCA (see Fig. 1 for locations and Table 1 for further details and
225 references). From all records derived from drillholes, only data from core or sidewall

226 core samples were evaluated to avoid potential contamination by caving. To exclude a
227 bias as it could have been introduced into our comparison through different
228 taxonomic criteria, we have standardized all datasets using broader groups of taxa
229 (e.g., *Gleicheniidites* spp./*Clavifera* spp., *Phyllocladidites* spp., *Nothofagidites* spp.
230 [*fusca* group]) when necessary. Due to the differences in sample sizes, sporomorph
231 percentages were used as input for the DCA, and only samples with counts ≥ 100
232 individuals were evaluated.

233

234 **2.4 Sporomorph-based climate reconstructions**

235 Quantitative sporomorph-based climate estimates were carried out following the
236 bioclimatic analysis approach of Greenwood et al. (2005). For each taxon with a
237 known nearest living relative (NRL; Table 2), climatic profiles were generated with
238 regard to mean annual temperature (MAT), coldest month mean temperature
239 (CMMT), warmest month mean temperature (WMMT), and mean annual
240 precipitation (MAP). The climate profiles are derived from (i) the dataset of Pross et
241 al. (2012), which is mainly based on distribution data from the Australian National
242 Herbarium online database (Australian National Herbarium, 2011) and the
243 mathematical climate surface software ANUCLIM 5.156 (Houlder et al., 1999), and
244 (ii) the PALAEOFLORA database, which contains climatic information for a plant
245 taxon based on its global distribution (Utescher and Mosbrugger, 2013). Following
246 Greenwood et al. (2005), the climatic values for each assemblage were calculated
247 based on the zone of overlap of the majority of taxa from that assemblage with respect
248 to a given climate parameter. This overlap interval was calculated using the 10th
249 percentile (as lower limit) and 90th percentile (as upper limit) of the total range of the
250 NLRs recorded in that assemblage. The climate estimate is given as the midpoint

251 between the lower and upper limits, with the error spanning from the lower to the
252 upper limit. Only samples with counts ≥ 100 individuals were used in the climate
253 reconstructions.

254 Paleoclimate estimates based on the NLR concept may be influenced by a number of
255 factors that need to be considered prior to the application of NLR-based
256 reconstruction methods (e.g., Mosbrugger and Utescher, 1997; Mosbrugger, 1999;
257 Pross et al., 2000; Utescher et al., 2000). These factors include (i) the potential
258 misidentification of the fossil taxa and/or NLRs; (ii) the potentially incomplete
259 coverage of the climatic tolerances of the NLRs; (iii) potentially unidentified
260 differences between the climatic tolerances of fossil taxa and their NLRs; and (iv) a
261 weakening of the connection between fossil taxa and NLRs the further one goes back
262 in time. Generally, these issues become increasingly important with the age of the
263 floras analyzed and may diminish the significance of the results (e.g., Poole et al.,
264 2005). They can, however, be identified and corrected via the application of multi-
265 proxy approaches. In particular, the NLR concept has been successfully applied to
266 both macrofloral and sporomorph assemblages from the early Paleogene of the higher
267 southern latitudes (Greenwood et al., 2003; Carpenter et al., 2012; Pross et al., 2012;
268 Contreras et al., 2013); the validity of the NLR-based results has been demonstrated
269 through the comparison with data from other, independent proxies.

270

271 **2.5 Statistical examination of the connection between floristic** 272 **composition and temperature**

273

274 To examine the correlation between the floristic composition of our sporomorph
275 record from ODP Site 1172 (as represented by DCA Axis 1 sample scores; Figs. 2, 4)

276 and $\text{TEX}_{86}^{\text{L}}$ - and $\text{TEX}_{86}^{\text{H}}$ -derived SST values from the same site (Bijl et al., 2013b;
 277 Hollis et al., 2014), we applied a state space model. State space models or dynamic
 278 linear models allow data distributed along time to be interpreted as the combination of
 279 several components, such as trends, or seasonal or regressive components (Petris et
 280 al., 2009; see West and Harrison, 1997, for further details on the advantages and
 281 development of the method). In essence, we modelled the dynamics of the 'true' (but
 282 unknown) SST and DCA Axis 1 sample scores in time, and modelled the observed
 283 values as deviations from these true values.
 284 The model was fitted with a Bayesian approach. The likelihood is defined by
 285 Equations (1) and (2), with both SST and DCA for each time point where either one
 286 or both was measured. We compared 60 SST data points with our 40 values of the
 287 DCA Axis 1 scores. Because only six of the SST data points and DCA Axis 1 sample
 288 scores are from the exact same depths, we treated the missing data with multiple
 289 imputation; in essence, we estimated them as extra parameters to be estimated (e.g.,
 290 Gelman et al., 2003). The cross-covariance between two points was calculated
 291 following Equation (3) and the final correlation using Equation (4).

$$293 \quad SST_i \sim N(x_{\text{SST}}(t(i)), t_{\text{SST}}^2) \quad \text{Eq. (1)}$$

$$294 \quad DCA_i \sim N(x_{\text{DCA}}(d(i)), t_{\text{DCA}}^2) \quad \text{Eq. (2)}$$

$$295 \quad \Delta d \begin{pmatrix} \sigma_{11}^2 & \rho\sigma_{11}\sigma_{22} \\ \rho\sigma_{11} & \sigma_{22} \end{pmatrix} + \begin{pmatrix} \sigma_{\text{SST}}^2 & 0 \\ 0 & \sigma_{\text{DCA}}^2 \end{pmatrix} \quad \text{Eq. (3)}$$

$$296 \quad \text{corr}(\text{SST}, \text{DCA}) = \frac{\Delta t \rho \sigma_{\text{DCA}} \sigma_{\text{SST}}}{\sqrt{(\Delta dt + \tau_{\text{DCA}}^2)(\Delta t \sigma_{\text{SST}}^2 + \tau_{\text{SST}}^2)}} \quad \text{Eq. (4)}$$

297

298 The model was fitted using OpenBUGS run through the BRugs package (Thomas et
 299 al., 2006) of the R software for statistical computing (R Development Core Team,

300 2011) fitted in a Bayesian framework using the BRugs package. The R script and
301 further explanation of the analysis is given in the Supplementary Information.

302

303 **3 Results**

304 **3.1 Sporomorph results from ODP Site 1172**

305 Of the 89 palynological samples processed from the Middle Paleocene to Lower
306 Eocene of ODP Site 1172, 40 samples yielded sporomorph counts ≥ 100 individuals
307 and were further used in our analyses. The preservation of sporomorphs is generally
308 good. A total of 197 sporomorph types were identified. A range chart with the relative
309 abundances of key taxa is presented in Fig. 2; a full account of the identified taxa and
310 their abundance data along with photomicrographs of key taxa is provided in the
311 Supplementary Information. Based on rarefied values, the entire study interval is
312 characterized by rich sporomorph assemblages (mean \pm s.d. = 29.2 ± 3.4 taxa/sample
313 at 100 individuals, $n = 40$). Remarkably low sporomorph species numbers are
314 recorded for the samples corresponding to the PETM (22.8 and 24.5 species/sample at
315 100 individuals; Fig. 2). The DCA results of our sporomorph record yield distinctly
316 different values for the Axis 1 and Axis 2 sample scores (Fig. 3a). They allow to
317 define three sample groups, with each sample group being characteristic for specific
318 time intervals of our record (Fig. 3a). These intervals are: (i) the middle and late
319 Paleocene; (ii) the middle/late Paleocene transition; and (iii) the early Eocene
320 including the PETM. All three sample groups comprise characteristic sporomorph
321 assemblages that are portrayed in the following.

322

323 **3.1.1 Middle Paleocene (60.7 – 59.5 Ma) and late Paleocene (59.0 –** 324 **55.6 Ma) intervals**

325 The sporomorph assemblages from the middle and late Paleocene intervals (excluding
326 the
327 middle/late Paleocene transition, see below) are represented by 20 samples. They are
328 dominated by gymnosperm pollen, which on average accounts for 45% of all
329 sporomorphs. The gymnosperm pollen is represented mainly, in the order of
330 decreasing abundances, by *Podocarpidites* spp. (botanical affinity: Podocarpaceae;
331 podocarps), *Dilwynites granulatus* (*Wollemia* [Wollemi pine]/*Agathis* [Kauri];
332 Macphail et al., 2013), *Phyllocladidites mawsonii* (*Lagarostrobos franklinii*; Huon
333 pine), and *Araucariacites* spp. (*Araucariaceae*; *Agathis* [Kauri] and *Araucaria*) (Fig.
334 2). Other abundant sporomorphs are, in the order of decreasing abundances,
335 *Cyathidites* spp. (probably *Cyatheaceae*), *Nothofagidites* spp. (*Nothofagus*; southern
336 beech), *Gambierina* spp., *Gleicheniidites* spp. (*Gleicheniaceae*), and *Ceratosporites*
337 spp. (*Lycopodiaceae*, *Selaginellaceae*).

338 Although the sporomorph assemblages from the middle to late Paleocene have very
339 similar floristic compositions, differences exist with regard to the percentages of some
340 taxa (Fig. 2). For instance, the early part of the middle Paleocene exhibits relatively
341 high (~30%) percentage of podocarpaceous pollen (mainly *Podocarpidites* spp.
342 [*Podocarpaceae*; podocarps] and *Phyllocladidites mawsonii* [*Lagarostrobos franklinii*;
343 Huon pine]). In contrast, the latest Paleocene is characterized by lower abundances of
344 these taxa (~14%), but exhibits higher percentages of *Cyathidites* spp. (~18%) and the
345 constant presence of *Myricipites harrisii* (probably *Casuarinaceae*; sheoak).
346 Remarkably, both the middle and late Paleocene are characterized by the presence of
347 *Arecipites* spp. (*Arecaceae*; palms) and *Cycadopites* spp. (*Cycadales*).

348

349 **3.1.2 Middle/late Paleocene transition interval (59.5 – 59.0 Ma)**

350 The sporomorph assemblages of the middle/late Paleocene transition interval are
351 represented by ten samples. They exhibit higher percentages of *Araucariacites* spp.
352 (*Araucariaceae*; ~19%), *Nothofagidites* spp. (*Nothofagus*; ~13%), and *Ceratosporites*
353 spp. (*Lycopodiaceae*, *Selaginellaceae*; ~6%) than the assemblages from the under-
354 and overlying Paleocene strata described in Section 3.1.1 above (Fig. 2). Other
355 abundant taxa are, in the order of decreasing abundances, *Cyathidites* spp.
356 (*Cyatheaceae*), *Dilwynites granulatus* (*Wollemia/Agathis*) and *Gleicheniidites* spp.
357 (*Gleicheniaceae*). The percentages of podocarpaceous types (*Podocarpidites* spp.
358 [*Podocarpaceae*] and *Phyllocladidites mawsonii* [*Lagarostrobos franklinii*]) are lower
359 (~10%) than in assemblages from the under- and overlying middle and late Paleocene.
360 Notably, the assemblages from the middle/late Paleocene transition interval are
361 devoid of *Arecipites* spp. (*Arecaceae*; palms) and *Cycadopites* spp. (*Cycadales*)
362 pollen.

363

364 3.1.3 PETM and early Eocene interval (55.6 – 54.2 Ma)

365 Assemblages from the PETM are documented in two samples with counts ≥ 100
366 individuals. They are characterized by high percentages (up to 39%) of *Dilwynites*
367 *granulatus* (*Wollemia/Agathis*). Other abundant taxa are, in the order of decreasing
368 abundances, *Cyathidites* spp. (*Cyatheaceae*), *Araucariacites* spp. (*Araucariaceae*), and
369 *Gleicheniidites* spp. (*Gleicheniaceae*). The assemblages of the PETM interval are
370 further characterized by the presence of *Malvacipollis* spp. (*Euphorbiaceae*, probably
371 *Eumalvoideae*) and *Myricipites harrisii* (probably *Casuarinaceae*). Findings of
372 *Spinizonocolpites prominatus* (*Nypa* palm) are restricted to the PETM and the earliest
373 Eocene (Fig. 2).

374 Sporomorph assemblages of the early Eocene interval are documented in 8 samples.

375 They show in general very high percentages (mean: 20%) of *Cyathidites* spp.
376 (probably Cyatheaceae; Fig. 2). Other abundant taxa are, in the order of decreasing
377 abundances, *Dilwynites granulatus* (*Wollemia/Agathis*), *Araucariacites* spp.
378 (*Araucariaceae*), *Podocarpidites* spp. (*Podocarpaceae*), and *Phyllocladidites mawsonii*
379 (*Lagarostrobos franklinii*). The early Eocene interval is also characterized by
380 relatively high percentages of *Proteacidites* spp. (~7%) and the constant presence of
381 *Malvacipollis* spp. (*Euphorbiaceae*, probably *Eumalvoideae*), *Myricipites harrisii*
382 (probably *Casuarinaceae*), *Arecipites* spp. (*Arecaceae*; palms), and *Cycadopites* spp.
383 (*Cycadales*).

384

385 **3.2 Constraints on sporomorph source region**

386 Several lines of evidence suggest that eastern Tasmania was the main source of the
387 sporomorphs encountered in the Middle Paleocene to Lower Eocene of ODP Site
388 1172. The distance of this site to the paleo-shoreline of eastern Tasmania during the
389 Paleocene–early Eocene was on the order of ~100 km, whereas the minimum distance
390 to George V Land (Antarctica) amounted to ~500 km (Fig. 1). A Tasmanian source is
391 further suggested based on the distribution pattern of reworked Permian and Triassic
392 sporomorphs. The assemblages from ODP Site 1172 are characterized by the constant
393 presence of elements reworked from Permian and Triassic strata (e.g.,
394 *Cannanoropollis* spp., *Protohaploxypinus* spp., *Alisporites* spp.; see Supplementary
395 Information, Plate II); the percentages of reworked sporomorphs reach up to 16% of
396 the total assemblages. A similar input of reworked Permian and Triassic material is
397 known for sporomorph assemblages from Paleocene–Eocene strata along the
398 Australo-Antarctic Gulf (Otway Basin; Harris, 1965) and from the Eocene of the
399 Wilkes Land margin (Contreras et al., 2013). In the Tasmania region, reworked

400 Permian and Triassic sporomorphs are recorded in the Paleocene–Eocene of the Bass
401 Basin (Partridge et al., 2003) as well as in Eocene strata off western Tasmania and on
402 the South Tasman Rise (Truswell, 1997). This pattern is consistent with the fact that
403 sporomorph-bearing sediments of Permian and Triassic age occur in several regions
404 of Tasmania (e.g., Playford, 1965; Truswell, 1978; Calver et al., 1984). In contrast,
405 reworked sporomorphs on the continental shelf off George V Land (Antarctica; Fig.
406 1) comprise only taxa with Cretaceous and Cenozoic ages (Truswell, 1983). A similar
407 picture emerges for the Cenozoic of the Gippsland Basin where Permian and Triassic
408 sporomorphs occur only sporadically and in low numbers (see reports in Department
409 of Primary Industries, 1999).

410 A further constraint on the source of the sporomorphs at ODP Site 1172 comes from
411 the DCA-based comparison of the floristic composition of the Site 1172 record with
412 other coeval Paleocene–Eocene sporomorph records from Southeast Australia (Bass
413 and Gippsland Basins, Southeast Highlands) and New Zealand (Middle Waipara and
414 Otaio River sections) (Fig. 3b; see Fig. 1 and Table 1 for site locations and details on
415 records). It suggests marked differences in floristic compositions between most
416 records, a result that can be further corroborated for the Bass and Gippsland Basins
417 based on carbon-isotope data from both basins. The Paleocene–Eocene successions of
418 these basins exhibit distinctly different bulk carbon isotope values, which is
419 interpreted to reflect the signal from different plant communities contributing to the
420 organic carbon input into the basins during that time (Boreham et al., 2003).
421 However, despite the floristic differences between the records analysed, there is a
422 remarkable similarity in DCA scores between the records from ODP Site 1172 and
423 the Bass Basin (Fig. 3b), which is located mainly on the shelf off northern Tasmania,
424 but also extends into Northeast Tasmania (Moore et al., 1984). This similarity further

425 corroborates the scenario of a Tasmanian source for the ODP Site 1172 sporomorph
426 assemblages.

427

428 **3.3 Sporomorph-based paleoclimate estimates**

429 The paleoclimate estimates derived from the sporomorph assemblages from ODP Site
430 1172 are presented in Fig. 4 and summarized in Table 3. For the middle Paleocene
431 (60.7–59.5 Ma), all reconstructed temperature parameters yield relatively cool values
432 (MAT = ~12°C; CMMT = ~7°C; WMMT = ~18°C). A pronounced further cooling is
433 documented in all temperature parameters for the middle/late Paleocene transition
434 interval (59.5–59.0 Ma). The reconstructed values represent the lowest temperatures
435 of the entire record; typical values are ~10°C for MAT, ~5°C for CMMT and ~16°C
436 for WMMT (Fig. 4). Markedly higher temperatures prevailed during the late
437 Paleocene (59.0–55.6 Ma) and early Eocene (55.6–54.2 Ma), with estimates for that
438 interval being on the order of ~15°C for MAT, ~8°C for CMMT and ~20°C for
439 WMMT. The highest temperatures of the entire study interval are recorded for the
440 Paleocene/Eocene transition interval (including the PETM); MAT, CMMT and
441 WMMT reached values of ~22°C, ~18°C and ~26°C, respectively, during that time
442 (Fig. 4). The sporomorph-based MAP estimates yield high, near-constant values
443 throughout the sequence studied, with an average value of ~180 cm/yr (Fig. 4).

444

445 **4 Interpretation**

446 **4.1 Floristic and climatic evolution**

447 Based on our qualitative and quantitative results from the sporomorph record from
448 ODP Site 1172, three main vegetation types prevailed on Tasmania from the middle
449 Paleocene to early Eocene. These vegetation types (i) exhibit different floristic

450 compositions based on the DCA results (Figs. 2 and 3a), (ii) show similar diversities
451 based on the rarefaction results (Fig. 2), and (iii) represent specific climatic conditions
452 based on our sporomorph-derived paleoclimatic reconstructions (Fig. 4). The floristic
453 characteristics and climatic requirements of all three vegetation groups are discussed
454 in the following sections.

455

456 **4.1.1 Middle Paleocene (60.7 – 59.5 Ma) and late Paleocene (59.0 –** 457 **55.6 Ma) intervals**

458 During both time intervals, the flora of Tasmania was characterized by gymnosperm-
459 rich forests dominated by podocarps; Araucariaceae, ferns, protealeans and
460 *Nothofagus* were further important components of the vegetation. Areaceae (palms)
461 and Cycadales were present during both time intervals. Although the middle
462 Paleocene and late Paleocene forests exhibited a strongly similar composition,
463 important differences existed. Podocarps reached a markedly stronger dominance
464 during the middle Paleocene than during the late Paleocene, whereas the late
465 Paleocene was characterized by higher abundances of ferns (mainly Cyatheaceae) and
466 the presence of Casuarinaceae (Fig. 2).

467 With regard to modern analogues, these forests bear resemblance to the extant warm-
468 temperate, Podocarpaceae-dominated forests of Southeast Australia and New Zealand
469 (e.g., Kershaw, 1988). Based on the structure of extant temperate forests from the
470 southern latitudes (Kershaw, 1988; Enright and Hill, 1995; Veblen et al., 1996; Reid
471 et al., 1999), the overstorey of Tasmanian forests during the middle and late
472 Paleocene was dominated by Podocarpaceae and, to a somewhat lesser extent,
473 Araucariaceae and *Nothofagus*. The understorey, in turn, was likely dominated by
474 ferns, with Cyatheaceae being particularly abundant during the late Paleocene.

475 Considering the habitats of their extant representatives, Arecaceae (palms),
476 Cycadales, Proteaceae, and Casuarinaceae formed parts of both the over- and
477 understory (compare Johnson and Wilson, 1993; Hill, 1994; Morley, 2000; Jones,
478 2002).

479 Climatically, the presence of tree ferns (i.e., Cyatheaceae), Arecaceae and Cycadales
480 implies mild climates with no or merely rare frost events. Owing to physiological
481 constraints (including manoxylic wood, large, unprotected buds, soft, water-rich
482 tissues, and a near-absence of frost-“hardening” mechanisms), all these plants are
483 unable to cope with sustained freezing (Sakai and Larcher, 1987; Wing and
484 Greenwood, 1993); today, they only occur in settings with CMMT $\geq 5.5^{\circ}\text{C}$
485 (Greenwood and Wing, 1995; Utescher and Mosbrugger, 2013).

486

487 **4.1.2 Middle/late Paleocene transition interval (59.5 – ~59.0 Ma)**

488 Across the middle/late Paleocene boundary, the warm-temperate forests
489 characterizing the vegetation on Tasmania during most of the middle and late
490 Paleocene as described above were transiently replaced by cool-temperate forests
491 dominated by *Nothofagus* (mainly *N. fusca* type) and Araucariaceae. Ferns,
492 podocarps and protealeans were further prominent components of this vegetation.
493 Palms (Arecaceae) and Cycadales, as they occurred both during the preceding part of
494 the middle (60.7–59.5 Ma) and the subsequent part of the late Paleocene (59.0–55.6
495 Ma; see above), were absent (Fig. 2). In light of the age control and the temporal
496 resolution of our record, these cool-temperate forests prevailed on Tasmania for ~0.5
497 Ma (based on the duration of elevated *Nothofagus* percentages) respectively ~0.3 Ma
498 (based on the absence of tropical to subtropical indicators such as palms and
499 Cycadales). The 1.3-Myr-long hiatus precludes us from firmly determining the

500 complete duration of the cold interval in the mid-Paleocene (Fig. 2), however from
501 marine records from New Zealand we deduce a duration of 0.5 Ma (Waipara section
502 and ODP Site 1121; Hollis et al., 2014).

503 Today, vegetation dominated by *Nothofagus* (*N. fusca* type) is typical of cool-
504 temperate forests from southern Australia and New Zealand (Kershaw, 1988).
505 Considering the structure of such forests (e.g., McGlone et al., 1996; Reid et al.,
506 1999), the vegetation thriving during the middle/late Paleocene transition interval was
507 characterized by a canopy dominated by *Nothofagus* and open understories dominated
508 by ferns (mainly Cyatheaceae and the parent plants of *Ceratosporites* spp.
509 [Selaginellaceae, Lycopodiaceae]). Araucariaceae were also present. Because extant
510 members of Araucariaceae are tall trees generally confined to the lower mid-latitudes
511 (Kershaw and Wagstaff, 2001), their presence appears at first sight incompatible with
512 an occurrence of cool-temperate forests. However, members of the genus *Araucaria*
513 also thrive in cold temperate forests in mountainous areas of Chile (Veblen, 1982) and
514 can withstand frost events as cold as -15°C (Prentice et al., 1992), which supports our
515 observation that Araucariaceae were a component of the cool-temperate forests of
516 Tasmania across the middle/late Paleocene transition.

517 Based on the overall floristic evidence, Tasmania witnessed a transient period of
518 cooler conditions lasting from ~ 59.5 to ~ 59.0 Ma. In light of the frost sensitivity of
519 Areaceae, Cycadales and Cyatheaceae (compare Section 4.1.1), the decline or total
520 absence of these taxa across the middle/late Paleocene transition (Fig. 2) suggests
521 harsher winters (with particularly frequent and/or cold frost events) during that time.
522 Such lower temperatures are corroborated by the comparison with coeval $\text{TEX}_{86}^{\text{L}}$ -
523 and $\text{TEX}_{86}^{\text{H}}$ -based SST data from ODP Site 1172 (Bijl et al., 2009; 2013b; Hollis et
524 al., 2014). For the interval from ~ 59.4 to ~ 59.0 Ma, they show the lowest values of

525 the entire Paleocene–Eocene SST record (Fig. 4, Table 3).

526

527 **4.1.3 PETM and Early Eocene interval (55.6 – 54.2 Ma)**

528 During the early Eocene, the composition of the forests on Tasmania was distinctly
529 different from that of the temperate forests thriving during the Paleocene (Figs. 2, 3a);
530 the underlying floristic turnover coincides with the onset of the PETM (see DCA Axis
531 1 sample scores in Fig. 2). Based on our sporomorph data, the early Eocene
532 vegetation was dominated by ferns and different angiosperms (mainly Proteaceae,
533 Casuarinaceae and Euphorbiaceae/Eumalvoideae). Remarkably, taxa that were
534 common during the Paleocene (e.g., *Ceratosporites* spp. [Lycopodiaceae,
535 Selaginellaceae] and *Gambierina* spp. [extinct clade]) declined dramatically in
536 abundance or disappeared completely during that time (Fig. 2). Because extant
537 Selaginellaceae and Lycopodiaceae are cosmopolitan families (Jermy, 1990;
538 Øllgaard, 1990), it is difficult to connect the disappearance of these taxa on Tasmania
539 during the earliest Eocene with specific ecological and climatic conditions.

540 The coexistence of frost-tolerant (e.g., Araucariaceae, Podocarpaceae) and
541 thermophilous taxa (e.g., Casuarinaceae, Arecaceae [palms]) suggests the presence of
542 paratropical forests *sensu* Morley (2000). In particular, the occurrence of *Nypa* from
543 the PETM onward into the early Eocene suggests the presence of tropical mangrove
544 vegetation along the coast of Tasmania. A similar vegetation, also containing
545 thermophilous taxa such as *Nypa* and *Gymnostoma* (Casuarinaceae), is documented
546 in early Eocene macrofloras from western Tasmania (Pole, 2007; Carpenter et al.,
547 2012). Hence, forests on Tasmania during the early Eocene consisted temporarily of
548 at least two vegetation associations: (i) A mangrove association characterized by
549 *Nypa*, which is only recognized during the PETM and the earliest Eocene, and (ii) a

550 paratropical association characterized by the coexistence of frost-tolerant taxa (i.e.,
551 Araucariaceae, Podocarpaceae and *Nothofagus*) and thermophilous elements such as
552 palms and Casuarinaceae.

553 Considering the ecology of the nearest living relatives of the plants represented by the
554 encountered sporomorphs, the overstorey vegetation during the early Eocene
555 comprised taxa such as Araucariaceae, Podocarpaceae and *Nothofagus*, whereas the
556 understorey was probably dominated by ferns (mainly Cyatheaceae). Members of the
557 Proteaceae, Casuarinaceae, Cycadales, Arecaceae (palms), and
558 Euphorbiaceae/Eumalvoideae may have been both components of the over- and the
559 understorey (Johnson and Wilson, 1993; Hill, 1994; Morley, 2000; Jones, 2002).

560 The number of sporomorph species registered at ODP Site 1172 remained relatively
561 constant from the middle Paleocene to the early Eocene (Fig. 2). This observation is
562 in contrast to Southeast Australia, where sporomorph assemblages from non-marine
563 and marginal marine settings (Partridge, 1976) exhibit a considerably higher diversity
564 during the early Eocene than during the Paleocene (Macphail et al., 1994; see Section
565 4.2.2. below). The reasons behind this discrepancy may be sought in the particularly
566 high sea level during the early Eocene as it is recorded regionally based on
567 sedimentological and paleontological data from ODP Site 1172 (Exon et al., 2004a)
568 and globally (Miller et al., 2005; Cramer et al., 2011). Owing to the selective nature
569 of marine sporomorph transport as a function of transport distance (e.g., Moss et al.,
570 2005), the higher sea level during the early Eocene than during the Paleocene
571 potentially caused a diversity decrease of the sporomorph assemblages at ODP Site
572 1172 (see also below).

573 Our data suggest that the floristic change connected to the PETM is similar to that
574 registered for the early Eocene. However, unravelling the exact anatomy of vegetation

575 change across the PETM at ODP Site 1172 is difficult due to the low sporomorph
576 yields in the respective sediments at that site. In addition, the interpretation of the
577 available data is hampered by the sea-level rise during the PETM (Sluijs et al., 2011);
578 the transgression-induced change in depositional setting towards more distal
579 conditions may have caused a bias in the composition and diversity patterns of
580 sporomorph assemblages, with the resulting assemblages being skewed towards a
581 dominance of easily transported sporomorphs (compare Traverse, 1994; 2008). In
582 light of this bias, the high abundances (up to 39%; Fig. 2) of *Dilwynites granulatus*
583 (*Wollemia/Agathis*) and the remarkably low diversities (Fig. 2) in the PETM samples
584 from ODP Site 1172 likely represent a change in depositional setting rather than a
585 true paleoecological signal. This interpretation is supported by the higher abundances
586 of the same species (~35%) in early Paleogene sediments deposited in distal
587 environments of the Bass Basin when compared to the markedly lower abundances
588 (~10%) in coeval sediments from nearshore settings in the same basin (see reports in
589 Partridge et al., 2003). Nevertheless, our data show that the environmental
590 perturbations connected to the PETM had a profound impact on the vegetation on
591 Tasmania. They lead to the extirpation of various ferns (e.g., *Perinomonoletes* spp.,
592 *Ceratosporites* spp. [Lycopodiaceae, Selaginellaceae]) and angiosperms (e.g.,
593 *Gambierina rudata*, *Nothofagidites* sp. 1 [*Nothofagus*]), and the appearance of new
594 angiosperms mainly within the Proteaceae family (e.g., *Proteacidites grandis*).

595

596 **4.2 Integration with other terrestrial vegetation records and** 597 **temperature estimates from the southern high latitudes**

598 Our results from ODP Site 1172 yield a ~6.5-Ma-long vegetation record for the
599 Tasman sector of the SW Pacific region spanning from the middle Paleocene to the

600 early Eocene. To obtain insights into the potential regional differentiation of
601 terrestrial ecosystems and climates in the high southern latitudes during that time, we
602 have integrated our data from ODP Site 1172 with the available information on
603 terrestrial ecosystems and temperatures from other parts of the SW Pacific domain.
604 Our integration is augmented by newly generated temperature estimates for
605 previously published sporomorph records from Southeast Australia and New Zealand
606 (see Tables 1 and 3 for further information on records evaluated and results). Because
607 other continuous, stratigraphically well-calibrated vegetation records across the
608 middle/late Paleocene transition interval are not yet available for the SW Pacific
609 region, we focus our comparison on the middle Paleocene, late Paleocene and early
610 Eocene.

611

612 **4.2.1 Middle and late Paleocene**

613 Paleobotanical records for the middle and late Paleocene are well known from
614 Southeast Australia (e.g., Bass Basin – Macphail et al., 1994; Blevin, 2003; Gippsland
615 Basin – Stover and Partridge, 1973; Macphail et al., 1994; Department of Primary
616 Industries, 1999). In summary, the middle and late Paleocene vegetation in this region
617 consisted predominantly of warm temperate forests characterized by podocarps,
618 Araucariaceae and ferns; angiosperms were represented mainly by Proteaceae (e.g.,
619 Taylor et al., 1990; Macphail et al., 1994; Greenwood et al., 2003; Greenwood and
620 Christophel, 2005). Similar warm temperate forests dominated by podocarps and
621 Araucariaceae, and with a strong contribution of Proteaceae, thrived on New Zealand
622 (Mildenhall, 1980; Raine et al., 2009).

623 Based on our results, Podocarpaceae together with Araucariaceae, Cyatheaceae and
624 Proteaceae were also the prevailing group of plants during the middle and late

625 Paleocene on Tasmania (Fig. 2; compare also Section 4.1.1). However, important
626 floristic differences existed between Southeast Australia, New Zealand and Tasmania
627 (Fig. 3b); they are mainly based on the restriction of certain sporomorph taxa to
628 specific regions (e.g., *Liliacidites* spp., *Cibotioidites tuberculiformis* – New Zealand;
629 *Ilexpollenites* spp. – Southeast Australia [Bass and Gippsland Basins, Southeast
630 Highlands]; *Tripunctisporis maastrichtiensis* – New Zealand, Bass Basin and
631 Tasmania). Despite the differences in floristic composition, the remarkable
632 dominance of Podocarpaceae and Araucariaceae in Southeast Australia, Tasmania and
633 New Zealand suggests that warm-temperate forests dominated by gymnosperms were
634 the prevalent vegetation type in the Southwest Pacific region during the middle and
635 late Paleocene. Moreover, thermophilous taxa such as palms, Olacaceae (*Anacolosa*)
636 and Cupanieae first appeared and/or increased significantly in abundance and
637 diversity during the latest Paleocene in Southeast Australia (Kemp, 1978; Macphail et
638 al., 1994). On New Zealand, typical tropical taxa (e.g., Cupanieae, *Austrobuxus*
639 [Euphorbiaceae], *Nypa*) also begin to appear during the latest Paleocene prior to the
640 PETM (Crouch and Brinkhuis, 2005; Raine et al., 2009). Hence, the arrival of
641 thermophilous elements indicates the onset of warmer conditions in the Southwest
642 Pacific region during the latest Paleocene. A scenario of warm conditions is further
643 corroborated by MAT estimates reaching ~18°C as derived from latest Paleocene
644 macrofloras in Southeast Australia (Greenwood et al., 2003).

645 Considering our sporomorph-based climate estimates for Southeast Australia and
646 Tasmania (Table 3), temperatures were higher during the late Paleocene than during
647 the middle Paleocene. Hence, the overall climatic and vegetation signal suggests that
648 terrestrial settings across the Southwest Pacific region consistently experienced a
649 pronounced warming during the late Paleocene.

650

651 **4.2.2 PETM and Early Eocene**

652 The effects of the PETM on terrestrial ecosystems in the high southern latitudes are
653 yet poorly constrained. Available records from Southeast Australia (Bass and
654 Gippsland Basins) covering the PETM and the earliest Eocene suggest the widespread
655 presence of *Nypa* during that time (Partridge, 1976). Climatically, this indicates a
656 MAT >21.7°C (Utescher and Mosbrugger, 2013). For the South Island of New
657 Zealand, sporomorph data from nearshore marine sediments document the
658 development of *Nypa* mangrove swamps and the appearance of pollen from the
659 thermophilous subfamily Cupanieae connected to the PETM interval (Handley et al.,
660 2011); moreover, the PETM is characterized by a percentage increase of fern spores
661 as well as of Euphorbiaceae/Eumalvoideae and Myrtaceae pollen at the expense of
662 gymnosperm pollen percentages (Handley et al., 2011). Similarly, sporomorph data
663 for the North Island of New Zealand as available from the Tawanui section show the
664 presence of *Nypa* pollen connected to the PETM (Crouch and Visscher, 2003).
665 However, besides a marked increase of *Dilwynites granulatus* (*Wollemia/Agathis*)
666 pollen, no other significant changes in floristic composition occur (Crouch and
667 Visscher, 2003). Considering that the PETM sediments of the Tawanui section are
668 part of a transgressive systems tract (Crouch and Brinkhuis, 2005; Sluijs et al., 2008),
669 the high abundances of *Dilwynites granulatus* (*Wollemia/Agathis*) may represent a
670 taphonomic rather than a paleoecological signal as suggested for the sporomorph
671 record of the PETM from ODP Site 1172 (see Section 4.1.3).

672 With regard to the early Eocene, the majority of vegetation records in the Southwest
673 Pacific region come from Southeast Australia; they suggest that by early Eocene
674 times the warm temperate, conifer-dominated forests of the late Paleocene had been

675 replaced by more diverse, meso- to megathermal angiosperm forests (Macphail et al.,
676 1994; Greenwood et al., 2003; Greenwood and Christophel, 2005). Although
677 sporomorph percentages are extremely variable within the available records from
678 Southeast Australia, Araucariaceae, Casuarinaceae, Euphorbiaceae/Eumalvoideae,
679 Proteaceae, and ferns are generally the dominant taxa; typical tropical elements such
680 as *Nypa*, *Anacolosa* and Cupanieae are also recorded (Kemp, 1978; Macphail et al.,
681 1994). This trend in vegetation development during the early Eocene is also
682 documented for Tasmania.

683 On New Zealand, early Eocene sporomorph assemblages exhibit a mixed Paleocene-
684 Eocene character, with a continued high abundance of conifer pollen (Crouch and
685 Visscher, 2003; Raine et al., 2009). However, thermophilous taxa such as Cupanieae,
686 Casuarinaceae and Euphorbiaceae (*Austrobuxus*) are constantly present in these
687 records (Pocknall, 1990; Raine et al., 2009), and Casuarinaceae pollen abruptly
688 started to dominate the sporomorph assemblages from ~54.5 Ma onwards (Raine et
689 al., 2009). On the Wilkes Land margin (Antarctica), paratropical vegetation has been
690 recorded during the early Eocene (53.9–51.9 Ma) with the notable presence of
691 thermophilous taxa such as Arecaceae (palms) and Bombacoideae (Pross et al., 2012;
692 Contreras et al., 2013),

693 With regard to temperature conditions, early Eocene macrofloras from Southeast
694 Australia suggest a MAT of ~19°C from ~56 to ~53 Ma (Greenwood et al., 2003),
695 which is very similar to our MAT estimates for coeval sporomorph records (~55.8–
696 54.3 Ma; lower *Malvacipollis diversus* zone of Partridge (2006; Table 1) from the
697 Bass and Gippsland Basins (~18°C; Table 3). On the Wilkes Land margin, climatic
698 estimates for the early Eocene (53.9–51.9 Ma) suggest a MAT of ~16°C for the
699 lowland regions (Pross et al., 2012). Although occasionally MAT values as high as

700 ~23°C are recorded during the earliest Eocene at ODP Site 1172, the mean MAT for
701 this time interval is ~14°C (Fig. 4, Table 3). This is markedly lower than those from
702 Southeast Australia, and even lower than those from the Wilkes Land margin.
703 Considering that our sporomorph-derived climate data from ODP Site 1172 mainly
704 reflect climate conditions along the coast of eastern Tasmania (compare discussion on
705 sporomorph source region in Section 3.2), these relatively low values may suggest
706 that the eastern part of Tasmania was influenced by the relatively cool Tasman
707 Current (Fig. 1). However, this argument is not supported by the TEX₈₆-derived SSTs
708 from ODP Site 1172 for the early Eocene, which are much higher (mean: 23°C -
709 TEX₈₆^L, 28°C - TEX₈₆^H; Bijl et al., 2013b). Terrestrial, macroflorally derived
710 temperatures on the order of 24°C from western Tasmania (Carpenter et al., 2012)
711 suggest that this region was significantly warmer than the eastern part of Tasmania.
712 Alternatively, another potential explanation for this discrepancy is that the sea-level
713 rise during the early Eocene biased the composition of the sporomorph assemblages at
714 ODP Site 1172 (compare Section 4.1.3) towards a dominance of easily transported
715 and/or particularly abundant sporomorphs indicative of cool conditions at the expense
716 of rarer sporomorphs indicative of warmer conditions. This scenario is supported by
717 the fact that many thermophilous plants from the Lower Eocene of the Southwest
718 Pacific region (e.g., Arecaceae [palms], Cupanieae, *Ilex*, *Nypa*) are mainly insect-
719 pollinated (Bush and Rivera, 1998; Barfod et al., 2011). Hence, these taxa reach only
720 low abundances in pollen spectra when compared to wind-pollinated taxa, and they
721 are not likely to be transported over larger distances before they settle (Jackson,
722 1994).

723 Despite of the potential bias on the early Eocene sporomorph assemblages at Site
724 1172, the supraregional replacement of temperate forests by paratropical forests

725 during the early Eocene on Southeast Australia, New Zealand and Tasmania and the
726 widespread occurrence of *Nypa* palms during the PETM in the same regions
727 consistently indicate a pronounced reorganization of the vegetation during the early
728 Eocene in the high southern latitudes connected to the PETM.

729

730 **4.3 Integration with other precipitation records from the southern high** 731 **latitudes**

732 Based on our paleoclimatic results from the sporomorph record of ODP Site 1172,
733 MAP on Tasmania was nearly constant (~180 cm/yr) from the middle Paleocene to
734 the early Eocene (Fig. 4). These values are comparable to the present-day
735 precipitation received by rainforests in western Tasmania at ~42 °S (185 cm/yr,
736 Corinna; Bureau of Metereology, 2012), along the west coast of the South Island of
737 New Zealand (212 cm/yr; Westport; NIWA, 2012), northern Australia (180 cm/year,
738 Darwin Botanic Gardens; Bureau of Metereology, 2012) and on the east coast of New
739 Caledonia (197 cm/year, Puloébo; Pesin et al., 1995). The high-precipitation regime on
740 Tasmania during the early Paleogene as deduced from the sporomorphs is further
741 corroborated by the dominance of the dinocyst genus *Senegalinium* at ODP Site 1172
742 during this time interval (Sluijs et al., 2011); this genus is characterized by many
743 freshwater-tolerant species (Sluijs et al., 2005; Brinkhuis et al., 2006). Hence, the
744 observed *Senegalinium* dominance, which is best explained by substantial freshwater
745 input, is well compatible with high precipitation on Tasmania and a resulting strong
746 freshwater influx on the Tasmanian continental shelf during the early Paleogene.
747 Because of the general lack of precipitation data from other sites at high southern
748 latitudes we can mainly compare our estimates from Tasmania with values deduced
749 from other coeval paleobotanical records. Based on our results from sporomorph

750 assemblages from Southeast Australia (Table 3), this region experienced similarly
751 high precipitation (MAP mean: ~200 cm/yr) during the early Paleogene; this is
752 consistent with MAP mean estimates (186–240 cm/yr) as derived from macrofloral
753 records from the Upper Paleocene to Lower Eocene of the same region (Greenwood
754 et al., 2003). On the Wilkes Land margin (Antarctica), high precipitation values
755 (MAP mean: ~132 cm/yr) are also suggested for the early Eocene (Pross et al., 2012).
756 These high precipitation values (>100 cm/yr) have been corroborated recently by
757 alkaline major element geochemistry for Eocene sediments from Antarctica
758 (Passchier et al., 2013).

759 Based on the overall precipitation data, Tasmania and the Australia-Antarctic region
760 experienced high rainfall conditions during the early Paleogene, comparable with
761 present-day rainforests from southern latitudes. This lends support to modeling
762 studies that include high atmospheric humidity as an important warming mechanism
763 for the higher latitudes (e.g., Abbot et al., 2009).

764

765 **4.4 Comparison with marine temperature evolution**

766 Based on our sporomorph data (as evidenced in the Axis 1 sample scores of the DCA
767 results, which represent the variation in floristic composition along the studied
768 interval; Fig. 4) and the $\text{TEX}_{86}^{\text{L}}$ and $\text{TEX}_{86}^{\text{H}}$ data of Bijl et al. (2009; 2013b) and
769 Hollis et al. (2014), there is a strong correlation between the vegetation composition
770 on eastern Tasmania and SST at ODP Site 1172 (Fig. 4). This connection is clearly
771 borne out by our results from the state space model, where there is a very strong
772 correlation of 0.997 when DCA Axis 1 sample scores are compared with $\text{TEX}_{86}^{\text{L}}$
773 (95% highest posterior density: 0.633–0.999 based on Equation [4]; compare Section
774 2.5) and 0.978 when DCA Axis 1 sample scores are compared with $\text{TEX}_{86}^{\text{H}}$ (95%

775 highest posterior density: 0.879–0.997 based on Equation [4]; compare Section 2.5).
776 Hence, the strong correlation between the temperature variability derived from TEX_{86}
777 and the floristic composition recorded at ODP Site 1172 demonstrates the impact of
778 temperature on the vegetation dynamics in the Southwest Pacific region during the
779 early Paleogene.

780 A close coupling between the temperature evolution in the marine and the terrestrial
781 realms is also evident through the comparison of our sporomorph-based temperature
782 estimates (notably WMMTs) with the TEX_{86} -derived SSTs (Fig. 4 and Table 3); it is
783 only during the early Eocene that the pronounced warming trend recorded by $\text{TEX}_{86}^{\text{L}}$
784 and $\text{TEX}_{86}^{\text{H}}$ is not clearly reflected in the sporomorph-based temperature estimates,
785 likely due to the sea-level increase during the early Eocene (compare Sections 4.1.3
786 and 4.2.2).

787 The terrestrial, sporomorph-derived MATs are markedly cooler than the SSTs derived
788 from $\text{TEX}_{86}^{\text{L}}$ and $\text{TEX}_{86}^{\text{H}}$ (Fig. 4), which based on the traditional perception of the
789 TEX_{86} proxy are supposed to represent surface-water MAT (e.g., Schouten et al.,
790 2002). At the same time, the TEX_{86} -derived SSTs are closely related to the
791 sporomorph-derived WMMTs (Fig. 4, Table 3). These observations suggest that
792 TEX_{86} -based temperatures **might be** biased towards summer conditions when applied
793 to early Paleogene records from the high southern latitudes. Such a seasonal bias has
794 also been suggested for other early Paleogene records from the Southwest Pacific
795 Ocean based on the comparison of TEX_{86} -derived SSTs with $\delta^{18}\text{O}$ and Mg/Ca values,
796 and with SSTs derived from climate model simulations (Hollis et al., 2012).

797

798 **5 Conclusions**

799 The middle Paleocene to early Eocene vegetation on Tasmania as reconstructed from

800 the sporomorph record of ODP Site 1172 was characterized by three different forest
801 types that thrived in high-precipitation regimes under different temperature
802 conditions. These forest types were: (i) warm-temperate rainforests dominated by
803 Podocarpaceae during the middle and late Paleocene; (ii) cool-temperate rainforests
804 dominated by *Nothofagus* and Araucariaceae that transiently prevailed across the
805 middle/late Paleocene transition interval (iii) paratropical rainforests dominated by
806 Cyatheaceae during the early Eocene with the remarkable presence of the mangrove
807 palm *Nypa* during the PETM and the earliest Eocene. The comparison with other,
808 previously published floral records from the Southwest Pacific region (including
809 Southeast Australia and New Zealand) supports the validity of our data for Tasmania.
810 It shows that temperate forests were replaced by paratropical forests during the early
811 Eocene throughout the Southwest Pacific region. This reorganisation in vegetation
812 composition included an increase in fern (mainly Cyatheaceae) and angiosperm
813 abundances (e.g., Proteaceae, Euphorbiaceae/Eumalvoideae, Casuarinaceae) at the
814 expense of gymnosperms (mainly podocarps).

815 The integration of terrestrial (i.e., floristic) and previously published marine (i.e.,
816 TEX₈₆-based SST) climate information from ODP Site 1172 shows that the surface-
817 water cooling of ~3°C across the middle/late Paleocene transition interval (~59.5 to
818 ~59.0 Ma) was paralleled by a transient demise of frost-sensitive plants (i.e., palms
819 and cycads) and the establishment of cool-temperate forests dominated by *Nothofagus*
820 and Araucariaceae on Tasmania. This suggests that cooler conditions (and notably
821 harsher winters with strong and/or frequent frosts) prevailed on Tasmania during that
822 time.

823 In light of the statistically robust connection between the floristic composition of the
824 sporomorph record from ODP Site 1172 and the previously published TEX₈₆- based

825 SST record from the same site, the vegetation dynamics on Tasmania during the
826 middle Paleocene to early Eocene were mainly driven by temperature; precipitation
827 remained high (with a MAP mean of ~180 cm/yr) throughout that time. Based on the
828 comparison of our sporomorph-derived temperatures with the TEX₈₆-based SSTs, we
829 conclude that TEX₈₆^L- and TEX₈₆^H-derived temperatures for the high southern
830 latitudes of the early Paleogene are likely biased towards summer conditions.

831

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845

846 **TABLE AND FIGURE LEGENDS**

847

848 **Table 1.** Sporomorph datasets evaluated in this study from the Southeast Australia
849 (Bass Basin, Gippsland Basin, Southeast Highlands) and New Zealand (Middle

850 Waipara and Otaio River sections). Asterisks indicate data derived from palynological
851 reports in Department of Primary Industries (1999). Plus signs denote data derived
852 from the Appendix C of Blevin (2003). Biozones and ages are based on Stover and
853 Evans (1973), Stover and Partridge (1973) and Partridge (2006).

854

855 **Table 2.** List of fossil sporomorph taxa from the Middle Paleocene to Lower Eocene
856 of ODP Site 1172 with known botanical affinities and literature source, the nearest
857 living relative (NLR) used in the climate reconstruction, and database where climate
858 profiles of the NLRs are derived from. Database (1) = PALAEOFLORA (Utescher
859 and Mosbrugger, 2013), Database (2) = Pross et al. (2012). Taxa used in the climatic
860 evaluation are printed in boldface.

861

862 **Table 3.** Sporomorph-based climate estimates for Paleocene to early Eocene records
863 from Southeast Australia (Bass Basin, Gippsland Basin, Southeast Highlands), New
864 Zealand (Middle Waipara and Otaio River sections) and ODP Site 1172. SST values
865 given for each stratigraphic interval represent average values of the data of Bijl et al.
866 (2013b) and Hollis et al. (2014), with (H) and (L) denoting TEX_{86}^H respectively
867 TEX_{86}^L . Sporomorph-derived climate estimates are based on the methodology of
868 Greenwood et al. (2005) and are presented with the average and the standard
869 deviation. MAT = Mean Annual Temperature, CMMT = Coldest Month Mean
870 Temperature, WMMT= Warmest Month Mean Temperature, MAP = Mean Annual
871 Precipitation.

872

873 **Figure 1.** Map of the Southwest Pacific Ocean showing the early Eocene (~53 Ma)
874 continental configuration, illustrating modern continents (gray), areas shallower than

875 300 m (blue) and locations of ODP Site 1172 and localities listed in Table 1. SM =
876 Southern Monaro Sections (Southeast Highlands), B = Burong-1, K = Kingfish-8, Ko
877 = Konkon-1, MW-OS = Middle Waipara and Otaio River sections, P = Poonboon-1,
878 R = Roundhead-1, S = Sweetlips-1, T = Turrum-4, ETP = East Tasman Plateau.
879 Modified after Cande and Stock (2004) and Sluijs et al. (2011).

880

881 **Figure 2.** Relative abundances of selected sporomorph taxa ([A] angiosperms, [G]
882 gymnosperms, [P] pteridophytes) representative of the middle Paleocene to early
883 Eocene assemblages from ODP Site 1172. DCA Axis 1 scores represent the
884 fluctuations in floristic composition between samples. Relative abundances and DCA
885 results are based on samples with counts ≥ 100 individuals only. Numbers of
886 sporomorph species are rarefied at 100 individuals. The intervals corresponding to the
887 middle/late Paleocene transition and PETM are marked by horizontal blue and red
888 bars, respectively. Age model and dinocyst zonation after Bijl et al. (2013b).

889

890 **Figure 3.** Comparison of the floristic composition based on the DCA sample scores
891 for (a) middle Paleocene to early Eocene sporomorph assemblages from ODP Site
892 1172; (b) Paleocene/early Eocene sporomorph assemblages from Southeast Australia
893 (Bass Basin, Gippsland Basin, Southeast Highlands), New Zealand (Middle Waipara
894 and Otaio River sections) and ODP Site 1172. Results are based on samples with
895 counts ≥ 100 individuals only.

896

897 **Figure 4.** Comparison of the sporomorph-derived climate estimates, SST values
898 based on TEX_{86}^L and TEX_{86}^H , and DCA Axis 1 sample scores from the Middle
899 Paleocene to Lower Eocene of ODP Site 1172. Sporomorph-derived climate estimates

900 are based on the methodology of Greenwood et al. (2005). Error bars represent the
901 minimum and maximum estimates determined using that method. SST data are from
902 Bijl et al. (2009; 2013b) and Hollis et al. (2014). Sporomorph-derived climate
903 estimates and DCA results are based on samples with counts ≥ 100 individuals only.
904 The intervals corresponding to the middle/late Paleocene transition and the PETM are
905 marked by horizontal blue and red bars, respectively. Age model and dinocyst
906 zonation after Bijl et al. (2013b).

907

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