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1 Vegetation and climate change during the Medieval
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4 Langvlei

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10 brian.chase@umontpellier.fr

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13 Nadia du Plessis¹, Brian M. Chase^{1,2}, Lynne J. Quick³, Torsten Haberzettl⁴, Thomas Kasper⁵, Michael E.
14 Meadows^{1,6}

15
16 ¹*Department of Environmental and Geographical Science, University of Cape Town, South Lane, Upper Campus, 7701*
17 *Rondebosch, South Africa*

18 ²*Institut des Sciences de l'Evolution-Montpellier (ISEM), University of Montpellier, Centre National de la Recherche*
19 *Scientifique (CNRS), EPHE, IRD, Montpellier, France*

20 ³*African Centre for Coastal Palaeoscience, Nelson Mandela University, Port Elizabeth, Eastern Cape 6031, South Africa*

21 ⁴*Physical Geography, Institute for Geography and Geology, University of Greifswald, Germany*

22 ⁵*Physical Geography, Institute of Geography, Friedrich-Schiller-University Jena, Germany*

23 ⁶*School of Geographic Sciences, East China Normal University, 500 Dongchuan Road, Shanghai 200241, PR China*
24
25

26 **Corresponding author:** Nadia du Plessis; dup.nadia@gmail.com

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28 **Abstract**

29 This paper presents continuous, high resolution fossil pollen and microcharcoal records from Bo
30 Langvlei, a lake in the Wilderness Embayment on South Africa's southern Cape coast. Spanning the
31 past ~1300 years and encompassing the Medieval Climate Anomaly (MCA; c. AD 950 – 1250) and the
32 Little Ice Age (LIA; c. AD 1300 – 1850), these records provide a rare southern African perspective on
33 past temperature, moisture and vegetation change during these much debated periods of the recent
34 geological past. Considered together with other records from the Wilderness Embayment, we
35 conclude that conditions in the region during the MCA chronozone were – in the context of the last
36 1300 years – likely relatively dry (reduced levels of Afrotropical forest pollen) and perhaps slightly
37 cooler (increased percentages of *Stoebe*-type pollen) than present. The most significant phase of
38 forest expansion, and more humid conditions, occurred during the transition between the MCA and
39 the most prominent cooling phase of the LIA. The LIA is clearly identified at this locality as a period
40 of cool, dry conditions between c. AD 1600 to c. AD 1850.

41 The mechanisms driving the changes observed in the Bo Langvlei pollen record appear to be
42 generally linked to changes in temperature, and changes in the influence of tropical circulation
43 systems. During warmer periods, moisture availability was higher at Bo Langvlei, and rainfall was
44 perhaps less seasonal. During colder periods, precipitation resulting from tropical disturbances was
45 more restricted, resulting in drier conditions. While increased precipitation has been reported during
46 the LIA from Verlorenvlei in the Western Cape as a result of an equatorward displacement of the
47 westerly storm-track at this time, the opposing response at Bo Langvlei suggests that any increased
48 influence of westerlies was insufficient to compensate for the concurrent reduction in tropical/local
49 rainfall in the region.

50 **Introduction**

51 To understand the nature and challenges of climate change in southern Africa it is important to
52 establish an understanding of natural variability in the recent geological past. Two key periods
53 characterize the climate since AD 650: the Medieval Climate Anomaly (MCA; c. AD 950 – 1250) and
54 the Little Ice Age (LIA; c. AD 1300 – 1850) (Jones et al., 2001; Matthews and Briffa, 2005). While
55 recognized as not being robust features in proxy records from all regions of the globe (e.g. PAGES2k
56 2013; Neukom et al. 2019), the study of these periods provides an important context for modern
57 and predicted future climate states.

58 It is well documented that the South African palaeoenvironmental record is relatively limited
59 in terms of both the quantity and quality of records compared with the mid-latitudes of the
60 Northern Hemisphere (see Chase and Meadows 2007), where the MCA and LIA have been defined.
61 This is mainly due to the region's highly seasonal rainfall regimes and generally arid to semi-arid
62 environments, which are not conducive to the preservation of sedimentary sequences and
63 associated proxy records. Recent reviews by Nicholson et al. (2013), Nash et al. (2016) and Lüning et
64 al. (2017, 2018) survey and synthesize the available data, but conclude that more records are still
65 needed in order to resolve inconsistencies and account for the complexity of spatial and temporal
66 variations.

67 Recent research initiatives on the southern Cape coast have been directed at addressing this
68 knowledge gap, focusing on sea level, climate and vegetation dynamics during the Holocene (e.g.
69 Haberzettl et al., 2019; Kirsten et al., 2018; Quick et al., 2018; Reinwarth et al., 2013; Strobel et al.,
70 2019; Wündsche et al., 2016a, 2016b, 2018). The climate along the southern Cape coast is influenced
71 by both tropical and temperate climate systems, and the region hosts a highly diverse vegetation
72 including fynbos and thicket elements and includes the Knysna Afrotemperate Region – the most
73 extensive forest complex in southern Africa (Geldenhuys, 1993; Midgley et al., 1997). Studies of the
74 available palaeoenvironmental records have concluded that regional climates have likely been highly
75 dynamic over time (Chase and Meadows, 2007), as the mechanisms controlling tropical and

76 temperate systems have responded to changing global boundary conditions (Chase et al., 2017;
77 Chase and Quick, 2018), and that these changes have significantly impacted the regional vegetation
78 mosaic (Martin, 1968; Quick et al., 2016, 2018). This apparently highly variable nature of the region's
79 climate and vegetation suggests it is particularly sensitive to climate change, making it an ideal area
80 to evaluate changes in these systems and how they interact over time.

81 To study vegetation and climate change on the southern Cape coast, we analyzed a ~1300-
82 year record of fossil pollen and microcharcoal from Bo Langvlei, one of several coastal lakes in the
83 Wilderness Embayment (Figure 1). We present here the first record to come from the region that is
84 sufficiently highly resolved to address questions relating to climate and vegetation change during the
85 last 1000 years (and therefore encompassing the MCA and LIA), and it contributes to the expanding
86 body of work defining a baseline for natural environmental variability along the southern Cape coast.

87 **Regional Setting**

88 A notable feature of the southern Cape coast is the Wilderness Embayment, which includes a series
89 of lakes separated from the coastline by shore parallel dune ridges of Pleistocene age (Bateman et
90 al., 2011; Illenberger, 1996) (Figure 1). The embayment is underlain by quartzites of the Table
91 Mountain Group as well as Palaeozoic (Ordovician – Silurian) Peninsula Formation sandstones
92 (Bateman et al., 2011; Marker and Holmes, 2002, 2010).

93 Today the largest of the three lakes of the Wilderness Lakes system, Bo Langvlei (2.14 km²;
94 Watling 1977) is connected by a short channel to Rondevlei to the East, and Eilandvlei to the West.
95 Langvlei Spruit, with a catchment size of 8.2 km² (Fijen and Kapp, 1995), is the main source of
96 freshwater to Bo Langvlei. Additional contributions are received through occasional overflow from
97 Rondevlei and groundwater from non-lithified sandy sediments that border and underlie parts of the
98 Wilderness Embayment (Fijen and Kapp, 1995).

99 *Climate*

100 Located in the modern year round, or aseasonal, rainfall zone (YRZ/ARZ) (*sensu* Chase and Meadows,
101 2007), the contemporary climate of the Wilderness region can be described as temperate oceanic
102 (“Cfb”; Peel et al., 2007) with 800 to 1000 mm of mean annual rainfall distributed largely uniformly
103 throughout the year (Allanson and Whitfield, 1983; Russel et al., 2012). Winter rainfall along the
104 coast is brought about by frontal depressions embedded in the mid-latitude westerlies, with ridging
105 anticyclones and cut off lows responsible for the majority of the rainfall events (Engelbrecht et al.,
106 2015; Engelbrecht and Landman, 2016; Tyson and Preston-Whyte, 2000). Summer rainfall results
107 from the association of warmer sea surface temperatures in the Indian Ocean and enhanced easterly
108 flow, as well as tropical-temperate trough cloud bands (Engelbrecht et al., 2015; Tyson and Preston-
109 Whyte, 2000). The Agulhas Current also has a major influence on local climatic conditions along the
110 coast, as its warm waters provide a source of moisture, generating increased humidity (Jury et al.,
111 1993).

112

113

[Insert Figure 1]

114 Figure 1: **Panel A** Map of southern Africa showing seasonality of rainfall and sharp climatic gradients dictated
115 by the zones of summer/tropical (red) and winter/temperate (blue) rainfall dominance. Major atmospheric
116 (white arrows) and oceanic (blue arrows) circulation systems are indicated. Locations of palaeoenvironmental
117 records discussed in this paper are numbered: 1; Verlorenvlei (Stager et al., 2012), 2; Seweweekspoort (Chase et
118 al., 2013, 2017), 3; Cango Cave (Talma and Vogel, 1992), 4; Groenvlei (Martin 1968; Wündsche et al. 2016a) 5;
119 Bo Langvlei (this paper), 6; Cold Air Cave (Holmgren et al., 1999, 2003; Lee-Thorp et al., 2001; Sundqvist et al.,
120 2013), 7; Pafuri (Woodborne et al., 2015) and 8; Lake Sibaya (Neumann et al., 2008; Stager et al., 2013). **Panel**
121 **B** Aridity index map of the southwestern Cape (black box in A). Locations of palaeoenvironmental records are
122 numbered as in Panel A. **Panel C** The Wilderness lakes region, indicating the location of Bo Langvlei and the
123 current distribution of dominant vegetation types (Mucina and Rutherford, 2006).

124

125 *Contemporary Vegetation*

126 Vegetation in the Wilderness region consists of a mosaic of fynbos and thicket elements with
127 pockets of Knysna Afrotropical forest also present (Cowling and Heijmans, 2001). In the immediate
128 vicinity of the lakes, including the interconnecting channels, semi-aquatic vegetation is represented
129 by bulrushes (*Typha latifolia*), reeds (*Phragmites australis*) and sedges (*Scirpus littoralis*) (Allanson
130 and Whitfield, 1983; Russel et al., 2012). The area inundated during extremely high water levels
131 represents an intermediate zone between the semi-aquatic and terrestrial environments and is
132 occupied by rushes such as *Juncus kraussi* (Allanson and Whitfield, 1983; Russel et al., 2012). Azonal
133 vegetation specifically along the estuaries is represented by halophytic taxa with *Sarcocornia*
134 *capensis* and *S. pillansii* (Amaranthaceae), *Chenolea diffusa* (Amaranthaceae) and *Plantago*
135 *crassifolia* being most prominent (Mucina and Rutherford, 2006; Quick et al., 2018).

136 North of the lakes, the coastal plains are occupied by Knysna Sand Fynbos which is
137 characterised by dense, moderately tall microphyllous shrublands (predominantly *Erica curvifolia*,
138 *Metalasia densa* (Asteraceae) and *Passerina rigida*) (Mucina and Rutherford, 2006; Quick et al.,
139 2018). Southern Cape Dune Fynbos (mainly *Olea exasperata*, *Phyllica littoralis* and a variety of *Searsia*
140 species) is found on the seaward barrier dune, as termed by Illenberger (1996), separating Bo
141 Langvlei from the coastline (Mucina and Rutherford, 2006; Quick et al., 2018). Patches of Southern
142 Afrotropical Forest are typically found on the south-facing slopes of the adjacent river catchments
143 and valleys, generally comprising *Afrocarpus falcatus* and *Podocarpus latifolius* (in the pollen record
144 we cannot differentiate between these species, as such these are all labelled *Podocarpus* for the
145 purpose of this paper), *Ocotea bullata* (Lauraceae) and *Olea capensis ssp. marcocarpa* (Oleaceae)
146 (Midgley et al., 2004; Quick et al., 2018). Garden Route Shale Fynbos occupies the boundary
147 between afrotropical forest and fynbos, dispersed along the coastal plateau, defined by
148 ericaceous and tall dense proteoid fynbos (Mucina and Rutherford, 2006; Quick et al., 2018).

149 **Material and Methods**

150 *Core extraction and sampling*

151 Coring was undertaken at Bo Langvlei during October 2013 as part of the first fieldwork campaign for
152 the RAIN project (Haberzettl et al., 2014). The core, BoLa 13.2 (33°59'12.54''S, 22°40'44.46''E), was
153 retrieved using a modified ETH-gravity corer (Kelts et al., 1986), and measures 178.5 cm in length.
154 BoLa 13.2 was opened at the Institute of Geography at Friedrich Schiller University Jena, Germany
155 where it was photographed, lithologically described and sampled. The core was continuously sub-
156 sampled in a closed laboratory setting at a resolution of 1 cm for both pollen and microcharcoal
157 analysis.

158 *Chronology*

159 The BoLa 13.2 age-depth model was established using both ²¹⁰Pb and ¹⁴C ages. 24 bulk samples from
160 the top 12 cm of the core (at 0.5 cm intervals) were sent to the Radiochronology Laboratory (Centre
161 for Northern Studies, Laval University, Quebec, Canada) for ²¹⁰Pb-dating. For AMS-¹⁴C dating, five
162 organic sediment samples were sent to Beta Analytic Inc. (Miami, Florida, USA). The ¹⁴C age for the
163 sample at 45.5 cm was calibrated using the SHcal13 curve (Hogg et al., 2013) as the radiocarbon
164 content of the uppermost sample showed 105.5±0.3 pMC indicating the absence of a reservoir
165 effect in the lacustrine facies of the sediment. The Marine13 data set (Reimer et al., 2013) was
166 applied for the samples at 98.5 cm, 137.5 cm and 178.5 cm, respectively as the lithology indicated
167 that sediment in these depths is of marine origin. A marine reservoir correction of $\Delta R = 148 \pm 27$ was
168 applied here in reference to recent studies by Wündsche et al. (2016a) and Haberzettl et al. (2019).
169 The CRS (constant rate of ²¹⁰Pb supply) model (Appleby, 2008; Appleby and Oldfield, 1978) was
170 applied to the ²¹⁰Pb results to obtain the ages. The age-depth model was subsequently developed
171 using the R software package Bacon (v2.2) (Blaauw and Christen, 2011).

172

173 **[Insert Table 1]**

174 Table 1: Radiocarbon ages and calibration details. Ages are presented as both cal BP and AD where relevant.

175

176 **[Insert Table 2]**

177 Table 2: ^{210}Pb data and age estimates

178

179

[Insert Figure 2]

180 Figure 2: The BoLa 13.2 lithology and age-depth model. The age-depth model was developed using the R
181 software package Bacon (V2.2) (Blaauw and Christen 2011). The 2σ probability distribution of calibrated ^{14}C
182 ages is presented in blue and the 95% confidence intervals are represented by the grey dotted line. ^{210}Pb ages
183 are represented by the turquoise area and the red line represents the best model according to the weighted
184 mean age at each depth. BoLa 13.2 is divided into four lithological units. Unit I (178 – 100 cm) predominantly
185 consists of silty sand and is characterized by an abundance of marine shell fragments. A gradual coarsening of
186 sediment is observed in Unit II-a (100 – 94 cm), changing to finer silty clay in Unit II-b (94 – 86 cm). Marine
187 shell fragments appear less frequently upwards of 94 cm. Unit III (86 – 27 cm) is mainly composed of clayey
188 silt. Units IV (27 – 6 cm) and V (6 – 0 cm) consists of the similar fine-grained sediment as Unit III. Unit IV is
189 characterized by the presence of filamentous plant material, while marine shell fragments are present again in
190 Unit V.

191

192 *Pollen and microcharcoal analysis*

193 In total, 84 samples were processed for pollen and microscopic charcoal analyses. Samples from the
194 basal 78.5 cm of the core yielded insufficient pollen concentrations for analysis. Standard
195 palynological methods were employed as per Faegri & Iversen (1989) and Moore et al. (1991) with
196 particular adaptations for dense media separation (Nakagawa et al., 1998) in order to extract
197 palynomorphs. These adaptations included the removal of carbonates and humic acids using 30%
198 HCl and 10% KOH, respectively. This was followed by heavy liquid separation, through the use of
199 ZnCl_2 , in order to separate pollen grains from the non-pollen matrix (Faegri and Iversen, 1989;
200 Moore et al., 1991; Nakagawa et al., 1998). Acetolysis was further applied to remove any cellulose
201 and/or organic matter still present in the sample, after which the samples were mounted on slides
202 using the aqueous mounting agent Aquatex. Three slides were produced per sample and 0.5 ml of

203 LacCore's polystaene microsphere pollen spike was added to each sample to determine pollen
204 concentrations and absolute counts.

205 Counts of 300 pollen grains (Birks and Birks, 1980) were performed using a Zeiss Axiostar
206 Plus microscope with magnifications from 400x to 1000x. Three samples exhibited lower pollen
207 concentrations, and from these only 222 (AD 1965), 280 (AD 740) and 299 (AD 760) grains were
208 observed. To aid in identification, the pollen reference collection from the Environmental and
209 Geographical Science department at the University of Cape Town was used, as was reference
210 material from Van Zinderen Bakker (1953, 1976), Van Zinderen Bakker and Coetzee (1959), Welman
211 and Kuhn, (1970) and Scott (1982). The counting process was aided by the use of the software
212 program Polycounter, version 2.5.3 (Nakagawa, 2007).

213 Charcoal particles were counted in conjunction with pollen grains. Particle number was
214 recorded according to the guidelines of Tinner & Hu (2003) with only black, opaque and angular
215 fragments $>75 \mu\text{m}^2$, or $\sim 10 \mu\text{m}$ in length, being counted. These fragments were classified into two
216 groups according to size: 10 – 100 μm and $>100 \mu\text{m}$ (Tinner and Hu, 2003). Charcoal concentrations
217 were calculated using the microsphere spike, as with the pollen concentrations.

218 The results obtained through pollen and charcoal analysis are presented using the software
219 package Tilia (version 1.7.16) (Grimm, 2011). Zonation of the diagrams was achieved using the
220 CONISS (Constrained Incremental Sum of Squares) module of the same software using
221 stratigraphically constrained analysis and square root transformation of the data (Grimm, 1987).

222 *TraCE21ka climate model*

223 To compare our results with general circulation model simulations, we use data from the TraCE21ka
224 experiment (He et al., 2013; Liu et al., 2009; Otto-Bliesner et al., 2014). TraCE21ka used the
225 Community Climate System Model ver. 3; (Collins, 2006) a global coupled atmosphere – ocean – sea
226 ice – land general circulation model that has a latitude – longitude resolution of $\sim 3.75^\circ$ in the
227 atmosphere and $\sim 3^\circ$ in the ocean and includes a dynamic global vegetation module. The simulation
228 includes transient orbitally forced insolation changes and changes in the atmospheric concentrations

229 of carbon dioxide, methane and nitrous oxide, as well as the evolution of ice sheets and their
230 meltwater contributions to the ocean. The climate data have been regridded using bilinear
231 interpolation to a spatial resolution of 2.5° x 2.5° (latitude/longitude) (Fordham et al., 2017).

232 **Results**

233 *Chronology*

234 The ¹⁴C ages (Table 1) indicate that BoLa13.2 has a basal age of 4140 ⁺¹⁸⁵/₋₂₂₀ cal BP (Figure 3).
235 The results from the CRS-model of ²¹⁰Pb activity (Table 2) provide ages for the top 12 cm of the core
236 ranging from AD 1825 to present. The age-depth model suggests continuous deposition and an
237 average sedimentation rate of 0.07 cm y⁻¹.

238 *Pollen and microcharcoal analyses*

239 Due to the sandy nature of the sediment in Unit I, as well as the presence of marine shells, it was
240 inferred that this unit represents marine conditions within the lake. Accordingly, it is suggested that
241 this marine environment was not conducive to the preservation of organic matter hence the
242 inadequate pollen preservation below 100 cm. As such, the pollen record encompasses the time
243 period from AD 680 ⁺⁵³⁰/₋₁₁₀ to present.

244 The BoLa13.2 pollen assemblage are summarized by ecological affinity based on the primary
245 vegetation types present in the region today, of which the main contributor is the fynbos vegetation
246 group, with *Ericaceae*, *Restionaceae* and *Stoebe*-type pollen being most prevalent. Succulent and/or
247 drought resistant taxa are mainly represented by *Euphorbia* with lesser contributions from
248 *Aizoaceae* and *Crassula*, while *Olea* is the main constituent of the coastal thicket group. *Podocarpus*
249 dominates the Afrotropical forest group with the other taxa in this group making negligible
250 contributions. *Cyperaceae*, *Juncaceae* and *Typha* were deemed to form part of the local wetland
251 taxa, and *Amaranthaceae* is also included in this group as it is an indicator of salt marsh conditions.
252 These local wetland taxa were excluded from the total pollen sum. From the cluster analysis results,

253 the pollen and charcoal diagrams were divided into six pollen assemblage zones – BoLa13.2-A to
254 BoLa13.2-F.

255 Zone BoLa13.2-A represents the oldest part of the pollen sequence, c. AD 650 to 750. As for
256 much of the record, fynbos taxa dominate this zone, exhibiting increasing percentages towards the
257 top of the zone, with a concomitant peak in the charcoal concentration. Relative to mean values for
258 the whole of the record, succulent/drought resistant taxa are more prevalent in this zone, most
259 notably *Euphorbia*. Coastal thicket is present throughout this period, since taxa from this group are
260 relatively abundant. Afrotropical forest (predominantly *Podocarpus*) pollen is relatively low during
261 this period.

262 Both the succulent/drought resistant and coastal thicket groups reach maximum levels in
263 zone BoLa13.2-B (c. AD 750 to 1000). A prominent increase in coastal thicket (*Olea*, *Euclea* and
264 *Celastraceae*) is noted ~AD 820 after which values remain consistently elevated throughout the rest
265 of the zone. Afrotropical forest pollen is notably more abundant during this period. There is,
266 however, a sharp decline ~AD 900, a point where the fynbos elements *Ericaceae* and *Passerina* reach
267 maximum percentages. Additionally, (non-local) aquatic vegetation is completely absent, while the
268 pollen concentration is at a minimum (1.4×10^3 grains g^{-1}). Towards the end of the zone, a marked
269 peak in charcoal concentration is evident ~AD 980 along with slightly elevated values noted for
270 fynbos (mostly *Stoebe*-type) and *Euphorbia*.

271 BoLa13.2-C (c. AD 1000 to 1100) is the shortest of the pollen assemblage zones, including
272 only three samples. *Podocarpus* is present at a minimum ~AD 1000, with *Icacinaeae* the only other
273 forest element present here. Simultaneously, *Ericaceae* representation is notably elevated (17%) and
274 *Cyperaceae*, representing local wetland vegetation, occurs minimally. At the top of the zone
275 significantly higher levels of *Amaranthaceae* are noted, accompanied by increased representation of
276 the fynbos elements *Passerina* and *Protea*. Coastal thicket is present at lower values than in the
277 previous zones with succulent/drought resistant taxa remain present at higher levels throughout the
278 zone.

279 The longest of the pollen assemblage zones, BoLa13.2-D (c. AD 1100 to 1600), is dominated
280 by the halophytic element *Amaranthaceae* which achieves maximum values ~AD 1150. The fynbos
281 taxa *Passerina* and *Stoebe*-type exhibit relatively high percentages at the beginning of the period,
282 and then decline between ~AD 1230 and AD 1410. The inverse of this pattern is observed in
283 *Podocarpus*, which increases from the beginning of the period, and maintains high levels until ~AD
284 1550. Succulent/drought resistant taxa remain generally constant throughout the zone. Towards the
285 top of the zone, ~AD 1550, both pollen- (13.89×10^3 grains g^{-1}) and charcoal (5.27×10^4 fragments g^{-1})
286 concentrations are at their highest.

287 The beginning of zone BoLa13.2-E (c. AD 1600 to 1900) is marked by a sharp decline in both
288 *Amaranthaceae* and *Podocarpus*, and an increase in fynbos taxa, particularly *Stoebe*-type. This latter
289 remains abundant throughout the zone, reaching maximum representation ~AD 1800. Charcoal
290 concentrations are close to their lowest values at this point. *Pinaceae* also appears for the first time
291 towards the top of the zone, ~AD 1850. *Podocarpus* percentages are variable but notably lower than
292 in zone BoLa13.2-D. Succulent/drought resistant taxa show a similar, but subtler pattern, declining
293 after ~AD 1750 towards the top of the zone.

294 Zone BoLa13.2-F represents the top of the core, extending from ~AD 1900 to the top of the
295 record. Fynbos generally declines across this zone apart from a peak between ~AD 1950 and 1980.
296 Both coastal thicket and succulent/drought resistant taxa follow a similar pattern in this zone with
297 percentages increasing until ~AD 1960, after which both decline. *Podocarpus* continues to exhibit an
298 inverse relationship to *Stoebe*-type pollen during this period, with a peak ~AD 1920, a sharp decline
299 ~AD 1960, followed by a period of increasing values until ~AD 1990, when it reaches its highest
300 levels. Charcoal concentrations are at their lowest for the whole assemblage at this point.

301

302

[Insert Figure 3]

303

304

Figure 3: Relative pollen percentage diagram for BoLa13.2 organized according to ecological affinity. Pollen taxa occurring at less than 1% are shown only as presence points. Exaggeration curves are 4x for taxa present

305 between 5 and 2% and 2x for those above 5%. Zonation of the diagram is based on cluster analysis results from
306 CONISS.

307

308

[Insert Figure 4]

309 Figure 4: Charcoal and pollen concentrations for BoLa13.2. Only charcoal fragments smaller than 100 μm were
310 present. Charcoal and pollen concentrations were calculated in the same manner using microsphere counts.
311 The zonation of the diagram is the same as for the pollen diagram.

312 Discussion

313 *Climate and environmental change during the last ~1300 years at Bo Langvlei*

314 Despite the diversity of vegetation types present in the Wilderness Embayment region, and
315 observed in the Bo Langvlei record, there are several primary patterns in the fossil record that can
316 be used to infer past environmental changes. Foremost among these is the variability observed in
317 the 1) Afrotropical forest pollen record (dominated by *Podocarpus*) and 2) *Stoebe*-type pollen,
318 which is one of the most prevalent fynbos taxa. Afrotropical forests, and *Podocarpus* in particular,
319 are sensitive to drought, and in the fossil record related pollen types have been shown to be
320 valuable indicators of aridity/humidity in the region (e.g. Quick et al. 2018) (with aridity/humidity
321 being distinct from rainfall amount per se [see Chevalier & Chase 2016]). *Stoebe*-type pollen is most
322 clearly associated with cooler temperatures, making it an important indicator of past temperature
323 change at many sites in South Africa (Quick et al., 2016; Scott, 1982; Scott et al., 2012). Considering
324 these elements of the pollen record, it is possible to infer general changes in – and relationships
325 between - moisture and temperature at Bo Langvlei.

326 In this paper, we limit the contextualization of these results to records within the southern
327 Cape, and select keystone records from more distant regions that can be used to infer mechanistic
328 relationships with southern Cape climates. This selection was made on the basis of a record's
329 resolution, age control, and the extent to which the proxy can be reasonably interpreted in terms of
330 changes in temperature and/or hydroclimate. Other records spanning the last 1300 years do exist in

331 southern Africa (e.g. Norström et al., 2018; Scott, 1996; Stager et al., 2013), but we defer detailed
332 inter-regional consideration of these records for a fuller synthetic study.

333

334 **[Insert Figure 5]**

335 Figure 5: Comparison of Northern Hemisphere temperature reconstruction (Moberg et al., 2005) with
336 percentages of *Stoebe*-type (cold indicator) and Afrotropical forest (humidity indicator) pollen and key
337 palaeoclimate records southern Africa (location of sites indicated in Figure 1). The Medieval Climate Anomaly
338 (MCA) and Little Ice Age (LIA) are indicated by red and blue shading, respectively, with the degree of shading
339 within these periods indicating the strength of the reconstructed Northern Hemisphere temperature anomaly
340 (Moberg et al., 2005) relative to the average temperatures between AD 500 and 1979.

341

342 From c. AD 650 to 900 the landscape around Bo Langvlei was generally characterized by
343 open scrub vegetation, dominated by ericoid/restioid fynbos (Figure 4). Despite the prevalence of
344 fynbos vegetation, relatively low levels of *Stoebe*-type and *Passerina* pollen imply that conditions
345 were not significantly cooler than present. Levels of Afrotropical forest pollen are low during this
346 period, while succulent/drought-resistant taxa and coastal thicket vegetation is relatively abundant,
347 indicating drier conditions and a diverse mosaic of vegetation in the Wilderness Embayment during
348 this period.

349 The period from AD 900 to AD 1300 – broadly consistent with the Medieval Climate Anomaly
350 (MCA; AD 950 – 1250, Jones et al. 2001) – begins with a peak in Afrotropical forest pollen from c.
351 AD 900 to AD 950, followed by a decline to minimum values ~AD 980, indicating dry conditions, and
352 a subsequent progressive increase until ~AD 1300 (Figure 4, 5). At Groenvlei, 20 km to the east of Bo
353 Langvlei in the Wilderness Embayment, changes in terrigenous sediment fluxes (e.g., grain size
354 percentages, Ti and Fe counts) have been interpreted as – in part – indicators of rainfall
355 amount/intensity (Wündsche et al. 2016a). These records show similar patterns to the Bo Langvlei
356 Afrotropical forest pollen record at multi-centennial timescales (Figure 5) and indicate that the
357 MCA was a relatively arid phase in the context of the last 1300 years. This may also be reflected in

358 the strongly variable nature of Amaranthaceae pollen frequencies. Quick et al. (2018), noted that in
359 this setting Amaranthaceae most likely represents halophytic species like *Salicornia* (Slenzka et al.,
360 2013); however, the presence of Amaranthaceae could also be related to drier and more
361 evaporative conditions. The strong negative correlation between Amaranthaceae and
362 Afrotemperate forest taxa during this period suggests that it may have been a time of greater
363 climatic variability and lower moisture availability.

364 In terms of temperature, *Stoebe*-type pollen is present at minimum values from c. AD 900 to
365 AD 950, increases slightly until ~AD 1180, and then declines as Afrotemperate forest pollen increases
366 in abundance. This potential indication of slightly cooler conditions around this time runs counter to
367 some other studies (Lüning et al., 2017; Tyson et al., 2000; Tyson and Lindesay, 1992), but the
368 existing evidence from southern Africa does not unequivocally indicate warmer conditions during
369 the MCA (see Nicholson et al. 2013; Figure 5). The Congo Cave speleothem palaeotemperature
370 record – the closest temperature record to Bo Langvlei – indicates only a slight warming during the
371 MCA, while the Cold Air Cave speleothem records indicate cooler conditions (Holmgren et al., 2003;
372 Lee-Thorp et al., 2001) according to the $\delta^{18}\text{O}$ interpretation of Sundqvist et al. (2013) (lower values
373 indicating cooler conditions). When considered with the $\delta^{13}\text{C}$ (lower values indicating less shallow-
374 rooting grass and thus drier conditions) and grey-scale records from Cold Air Cave (Holmgren et al.,
375 2003; Lee-Thorp et al., 2001), this cooling may be associated with weakened tropical influence in
376 eastern South Africa and the southern Cape. Concurrent indications of increased precipitation at
377 Verlorenvlei in the Western Cape (Stager et al., 2012) and more humid conditions at nearby
378 Seweweekspoort (Chase et al., 2013, 2017) may suggest the increased influence of temperate
379 circulation systems at this time, supporting the model of an coeval inverse relationship between
380 temperate and tropical circulation systems in South Africa (Cockroft et al., 1987; Van Zinderen
381 Bakker, 1976). While simulations of seasonal precipitation in the Bo Langvlei region are complex
382 during the MCA, precipitation seasonality is generally higher (Figure 6) (He et al., 2013; Liu et al.,

383 2009). This is consistent with a decrease in Afrotemperate forest pollen, which favours regular
384 rainfall and low seasonality.

385

386 **[Insert Figure 7]**

387 Figure 7: Comparison of Northern Hemisphere temperature reconstruction (Moberg et al., 2005) with
388 percentages of *Stoebe*-type (cold indicator) and Afrotemperate forest (humidity indicator) pollen and climatic
389 parameters for the Wilderness region obtained from the TraCE-21k transient climate model simulation (He et
390 al., 2013; Liu et al., 2009). The Medieval Climate Anomaly (MCA) and Little Ice Age (LIA) are indicated by red
391 and blue shading, respectively, with the degree of shading within these periods indicating the strength of the
392 reconstructed Northern Hemisphere temperature anomaly (Moberg et al., 2005) relative to the average
393 temperatures between AD 500 and 1979.

394

395 Along with climate, fire should also be considered as a factor driving vegetation change
396 when interpreting the Bo Langvlei record. Major peaks in charcoal concentration are observed at AD
397 715 ⁺⁵⁰⁰/₋₁₃₀, AD 980 ⁺³³⁰/₋₁₈₀, AD 1265 ⁺¹⁵⁰/₋₉₅ and AD 1550 ⁺¹⁴⁰/₋₁₆₀, indicative of
398 discrete periods of increased fire activity and/or large fire events. These episodes are often followed
399 by declines in forest pollen and increases in fynbos representation (Figure 5), which might be
400 expected considering the fire-adapted nature of fynbos (Cowling et al., 2004). Afrotemperate forest
401 percentages and charcoal concentrations exhibit a generally positive relationship for these major
402 peaks, which may appear contradictory, as fires are associated with more seasonal rainfall or drier
403 conditions. In absence of other charcoal records that may help elucidate the regional significance of
404 these data, we propose that the relationship with forest pollen may relate to increases in biomass,
405 which would result in a larger charcoal flux, and/or the warmer conditions that appear to foster
406 forest expansion. Under warmer conditions, periods of increased seasonality or anomalously dry
407 years or decades may occur that provide favorable conditions for ignition, and the accumulation of
408 forest biomass may result in larger fires. Moreover, the occurrence of periodically more extreme

409 berg wind conditions can have a desiccating effect on vegetation, increasing the flammability of
410 habitually moist vegetation, such as forests (Geldenhuys, 1994). A more detailed study of long-term
411 fire ecology in the Wilderness region and Afrotemperate Forest Biome is clearly warranted.

412 In general, the palaeoenvironmental data described above appear to support climatic factors
413 as being the primary determinant of the Bo Langvlei pollen record. However, as declines in
414 Afrotemperate forest pollen systematically follow the major charcoal peaks in the record, the
415 sensitivity of this vegetation type to fire, and its recovery time follow major burning events/phases
416 may also influence the record, perhaps amplifying declines in Afrotemperate forest pollen at the
417 beginning of more arid phases and contributing to the pattern of progressive increases in this pollen
418 type during these periods.

419 Towards the end of the MCA interval the landscape appears to have become progressively
420 more forested, as Afrotemperate forest percentages increase significantly after ~AD 1200. Although
421 of lower resolution, a similar expansion of forest is noted in the Groenvlei (Martin, 1968) and
422 Eilandvlei (Quick et al., 2018) pollen records, with wetter than present conditions inferred for the
423 region from c. AD 1250 to 1350 (Wündsche et al. 2016a). The Bo Langvlei record indicates that the
424 period from c. AD 1200 to 1400, spanning the transition from the MCA to the LIA is characterized by
425 relatively warm (low levels of *Stoebe*-type pollen) and humid (increased Afrotemperate forest
426 pollen) conditions (Figure 5). These findings are supported by the palaeotemperature records from
427 Cango Cave (Talma and Vogel, 1992) and Cold Air Cave (Holmgren et al., 2003; Lee-Thorp et al.,
428 2001), records of terrestrial sediment flux at nearby Groenvlei (Wündsche et al. 2016a), and
429 reductions in simulated precipitation seasonality (He et al., 2013; Liu et al., 2009).

430 After ~AD 1400, broadly coincident with the onset of Northern Hemisphere Little Ice Age
431 cooling, percentages of *Stoebe*-type pollen begin to increase significantly, indicating cooling in the
432 Wilderness Embayment region (Figure 5). This cooling coincides with a period of forest retreat and
433 drier conditions (declining Afrotemperate forest percentages and reduced terrestrial sediment flux
434 at Groenvlei (Wündsche et al. 2016a)), with the period from c. AD 1400 to 1600 exhibiting marked

435 environmental variability, with strong shifts between warm-wet and cool-dry conditions. The Cold
436 Air Cave speleothem records indicate a cooling and drying trend in the northern summer rainfall
437 zone at this time (Holmgren et al., 2003; Lee-Thorp et al., 2001), while the lower resolution Cango
438 Cave speleothem records a strong increase in temperature (Talma and Vogel, 1992). The
439 Verlorenvlei diatom record has been interpreted as indicating increased – but variable –
440 precipitation in the winter rainfall zone (Stager et al., 2012). Based on simulations of winter rainfall
441 amount in the region, these findings cannot be easily extrapolated to the Wilderness Embayment,
442 but patterns in simulated rainfall seasonality are consistent with expectations based on
443 Afrotropical forest pollen (Figure 6). The challenges in determining the influence of tropical versus
444 temperate circulation systems on winter and summer rainfall regimes in the southern Cape has
445 become increasingly recognized in recent years (Chase et al., 2017, 2020; Chase and Quick, 2018),
446 and further data from both terrestrial and marine archives is required to adequately address these
447 questions.

448 From around c. AD 1600 to 1850 – the coldest portion of the LIA (Matthews & Briffa 2005) –
449 fynbos elements dominate the landscape, most notably *Stoebe*-type and the cryophilic element
450 *Passerina*, clearly indicating cooler conditions at Bo Langvlei at this time. A strong decline in
451 Afrotropical forest pollen is also observed, signaling substantially drier conditions (Figure 5). This
452 is further supported by increased Asteraceae pollen, suggestive of drier, more asteraceous fynbos.
453 These findings are consistent with Cango Cave palaeotemperature reconstructions (Talma and Vogel,
454 1992) and inferences of drier conditions at Groenvlei (Wündsche et al. 2016a). The Cold Air Cave
455 record also indicates cool, dry conditions in the northern summer rainfall zone during this period
456 (Holmgren et al., 2003; Lee-Thorp et al., 2001), as does the Parfuri baobab tree-ring record
457 (Woodborne et al., 2015). In contrast, the Seweweekspoort (Chase et al., 2013, 2017) and
458 Verlorenvlei (Stager et al., 2012) records suggest increased temperate influence/winter rain,
459 respectively. Considering the opposing trends between Verlorenvlei and Bo Langvlei, it can be
460 hypothesised that the influence of the frontal systems associated with the westerly storm-track was

461 not sufficient along the southern Cape coast to compensate for the concurrent reduction in
462 tropical/local rainfall in the region.

463 After ~AD 1850, fynbos pollen – and *Stoebe*-type pollen in particular – decreases sharply,
464 and there are marked increases in *Podocarpus* pollen, consistent with warmer, more humid
465 conditions at Groenvlei (Wündsche et al. 2016a). It seems that the LIA was followed by a period of
466 increasing temperatures and moisture availability and the establishment of the current aseasonal
467 rainfall pattern in the Wilderness area. This trend is also observed at Eilandvlei (Quick et al., 2018)
468 and Cango Cave (Talma and Vogel, 1992), and thus seems to be a well-defined response in at least
469 this portion of the southern Cape. It should be noted that this period is one of increasing colonial
470 human impact in the region, with *Pinus* appearing in the Bo Langvlei record from ~AD 1850. As
471 *Stoebe*-type pollen has in some cases been related to disturbance (Meadows et al., 1996) it may be
472 that some of the variability observed in the most recent portion of the record may be related to non-
473 climatic influences.

474 **Conclusions**

475 The Bo Langvlei pollen record provides valuable information regarding past temperature, moisture
476 and vegetation change from the southern Cape Coast during the past 1300 years. Consideration of
477 this period is often framed in terms of the Medieval Climate Anomaly (MCA; c. AD 950 – 1250) and
478 the Little Ice Age (LIA; c. AD 1300 – 1850) (Jones et al. 2001; Matthews & Briffa 2005; PAGES2k
479 2013), which have largely been defined based on evidence from the Northern Hemisphere.
480 Syntheses of palaeoenvironmental records from southern Africa have sometimes concluded that
481 regional patterns have been coherent with these periods of warmer and cooler climates (e.g. Lüning
482 et al., 2017; Tyson et al., 2000; Tyson and Lindesay, 1992), but other works have highlighted a more
483 complex spatio-temporal patterning of climate anomalies (e.g., Nicholson et al. 2013; Nash et al.
484 2016).

485 The data presented in this paper – together with other records from the Wilderness
486 Embayment – indicate that conditions in the region during the MCA chronozone were relatively dry

487 and perhaps slightly cooler than present. The most durable phase of forest expansion, and likely
488 more humid conditions, occurred during the transition between the MCA and core cooling of the
489 LIA. Data from the TraCE-21ka transient simulation (He et al., 2013; Liu et al., 2009) indicate that this
490 was a period of reduced rainfall seasonality, which would be consistent with the expansion of
491 drought-sensitive forest taxa like *Podocarpus*. The strongest signal preserved in the Bo Langvlei
492 pollen record is the period of cool, dry conditions that occurred during the LIA, most notably from c.
493 AD 1600 to 1850. Post-LIA warming occurred rapidly after ~AD 1850, but some of the subsequent
494 variability observed may be related to non-climatic factors such as intensifying colonial land use.

495 The mechanisms driving the changes observed in the Bo Langvlei pollen record appear to be
496 generally linked to changes in temperature, and changes in the influence of tropical systems,
497 perhaps transmitted at least in part via the Agulhas Current and the development of localized
498 precipitation systems (cf. Chase and Quick, 2018). During relatively warmer periods, moisture
499 availability was apparently higher, and rainfall was perhaps less seasonal, fostering the development
500 and expansion of Afrotropical forests. During colder periods, precipitation resulting from tropical
501 disturbances or relating to higher Agulhas Current sea-surface temperatures was more restricted,
502 resulting in drier conditions, and possibly greater seasonal contrasts in rainfall. Records from
503 Seweweekspoort (Chase et al., 2013, 2017) and Verlorenvlei (Stager et al., 2012), which have been
504 associated with temperate circulation system controls, indicate wetter conditions during the LIA.
505 This supports to a degree the commonly cited coeval inverse relationship between tropical and
506 temperate moisture-bearing systems, and increased westerly influence under globally cooler
507 conditions (see Chase and Meadows, 2007; Tyson, 1999; Van Zinderen Bakker, 1976). The limited
508 impact of increased frontal systems on the Bo Langvlei record during the LIA reinforces the
509 proposed importance of summer rainfall in regulating moisture availability along the southern Cape
510 coast, as suggested by Quick et al. (2018).

511 Data from the TraCE-21ka transient climate simulation (He et al., 2013; Liu et al., 2009)
512 indicate that both the MCA and LIA were periods of higher rainfall seasonality, which is consistent

513 with trends in Afrotemperate forest pollen. However, the drying trend during the LIA is simulated as
514 resulting from a decline in winter rainfall, in apparent contradiction with the above-mentioned
515 inferences and conceptual models. Whether this is due to insensitivity or inaccuracy in the models at
516 these spatio-temporal scales during the late Holocene, or the need for refinement of interpretive
517 paradigms, as has been recently suggested (e.g., Chase et al. 2017; Chase & Quick 2018) remains
518 unclear. More data along transects encompassing the climate gradients of the southern Cape region
519 are required to develop a more comprehensive understanding of regional climate dynamics, drivers
520 and the impact of climate change on regional environments.

521

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529

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