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1 **The Late Cretaceous Environment of the Arctic:**
2 **A Quantitative Reassessment based on Plant Fossils**

3
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13
14 **Abstract**

15
16 Late Cretaceous megafossil floras from the palaeo-Arctic of northeastern
17 Russia and northern Alaska are reviewed in respect of their age, composition,
18 structure and floral dynamics. Palaeofloral correlations and comparisons are
19 made between the two regions. Nine angiosperm-rich, predominantly
20 Cenomanian to Coniacian, floras from the palaeo-Arctic are re-evaluated
21 using Climate Leaf Analysis Multivariate Program (CLAMP) calibrated using a
22 global gridded (0.5° x 0.5°) climate data set derived from that used in climate
23 modelling. Additional floras from lower palaeolatitudes were used to derive
24 latitudinal temperature gradients: seven from N. America, five from around 30
25 °N palaeolatitude in Europe and one from Kazakhstan. The Arctic climatic

26 determinations, similar to previous estimates, support the existence of a
27 northern Pacific Ocean cold gyre and a warm Arctic Ocean. At palaeolatitudes
28 greater than 80°N floras are insufficiently diverse in woody dicot taxa to use
29 CLAMP, but using CLAMP-derived latitudinal temperature gradients Arctic
30 Ocean coastal environments at 70 Ma and 82°N, and which supported a
31 diverse dinosaur megafauna, are predicted to have experienced a mean
32 annual temperature of $6.3 \pm 2.2^\circ\text{C}$, a warm month mean of $14.5 \pm 3.1^\circ\text{C}$ and a
33 cold month mean no colder than $-2.0 \pm 3.9^\circ\text{C}$. All uncertainties are 2σ . The
34 new estimates are in good agreement with a wide range of non-
35 palaeobotanical climate proxies and render as an outlier warmer temperature
36 estimates for the Arctic Ocean derived from the TEX_{86} proxy. Modelling,
37 however, shows that land to ocean temperature gradients could have been
38 steep. The CLAMP estimates also suggest high values for humidity and
39 precipitation consistent with sedimentological indicators and, coupled with
40 warm temperatures, support the existence of a persistent polar cloud cap that
41 helped maintain high terrestrial air temperatures throughout prolonged periods
42 (up to 5 months) of winter darkness.

43

44

45 **Keywords:** Late Cretaceous; Arctic; climate; vegetation; CLAMP; plant fossils

46

47 **1. Introduction**

48

49 The existence of fossil floras attesting to vigorous forest growth in polar
50 regions (palaeolatitudes $> 66^\circ$) during Mesozoic and early Tertiary times has

51 been long been used as evidence for significant natural global climate
52 change. In 1914 the palaeobotanist Marie Stopes underscored the importance
53 of polar floras in a climatic context by quoting Asa Gray as saying "Fossil
54 plants are the thermometers of the ages by which climatic extremes and
55 climate in general through long periods are best measured" and cites Darwin
56 who, in a letter to Hooker in 1881, wrote "The extreme importance of the
57 Arctic fossil plants is self evident", and went on to urge Hooker to "Take the
58 opportunity of groaning over our ignorance of the lignite plants of Kerguelen
59 Land or any Antarctic land. It might do good." (Stopes, 1914).

60 Since then our understanding of plate tectonics has emphasized the
61 extent of climate change by showing that in most cases these ancient polar
62 plants were growing at even higher palaeolatitudes than originally thought
63 when continents were regarded as fixed. In the Arctic abundant leaf fossils, in
64 situ trees and dinosaur remains are preserved from sites that were at the
65 highest known palaeolatitudes where land existed (Brouwers et al., 1987;
66 Clemens, 1994; Clemens and Nelms, 1993; Fiorillo, 2008; Fiorillo and
67 Gangloff, 2000; Fiorillo et al., 2009; Gangloff, 1998; Gangloff et al., 2005;
68 Spicer and Parrish, 1987; Spicer and Parrish, 1990b) and in some
69 reconstructions close to the North Pole itself (Fig. 1). That said, the tectonics
70 of the Arctic are complex and differences in plate reconstructions result in
71 differences in the palaeo-positions of individual fossils sites. For example
72 Figure 1 shows north polar projections for 100 Ma, 88 Ma and 70 Ma derived
73 data compiled by Hay et al. (1999). Important differences exist between these
74 reconstructions and others in widespread use (e.g. Smith et al., 1981; Ziegler
75 et al., 1983). Uncertainties for Alaskan Campanian-Maastrichtian

76 palaeolatitudes have been quoted as +4 to -9° (Amniot et al., 2004). Unless
77 otherwise indicated the palaeolatitudes quoted here are those provided in
78 original publications that in some cases are based on palaeomagnetic data
79 local to the fossil sites. However, those locations should be considered
80 against those suggested in Figure 1 as an indication of the uncertainties
81 involved. For internal consistency the calculation of equator-to-pole
82 temperature gradients presented later in this paper use the plate positioning
83 programme Point Tracker (Scotese, per comm. to RAS, 2009).

84

85 [Figure 1 near here]

86

87 During the greenhouse climates of the Mesozoic zones of maximum
88 non-marine organic productivity, as evidenced by the mapping of fossil
89 remains and climatically sensitive sediments, were significantly polewards of
90 the equator (Chumakov et al., 1995; Rees et al., 2000; Semikhatov and
91 Chumakov, 2004; Spicer, 2003). In the Arctic widespread Late Cretaceous
92 sediments across Alaska, Canada and Russia contain an exceptionally rich
93 fossil record that provides a window into a time of abundant primary
94 productivity, carbon sequestration, and insights into a climatic regime that
95 provide a potential blueprint for managing future Arctic environmental change.

96 The areas of the Arctic where the most detailed and extensive
97 palaeobotanical and sedimentological records exist at the highest
98 palaeolatitudes are in northeastern Russia, Siberia and Northern Alaska and
99 they form the basis of this review and re-evaluation (Fig. 2). Comparable sites
100 representing such high palaeolatitudes are not currently accessible in the

101 Southern Hemisphere, so the Arctic provides our only insight into truly polar
102 conditions during the Late Cretaceous. In this paper previous palaeoclimate
103 estimates using these floras are reviewed and revised using a new
104 recalibration (Spicer et al., 2009a) of the multivariate foliar physiognomic
105 technique known as CLAMP (Climate Leaf Analysis Multivariate Program)
106 based on the global 0.5° x 0.5° gridded meteorological data of New et al.
107 (1999).

108

109 [Figure 2 near here]

110

111 **2. Cretaceous Floras from Alaska**

112

113 In northern Alaska Late Cretaceous active deltaic sedimentation
114 bequeathed an unparalleled legacy of inter-fingering marine and non-marine
115 fossiliferous sediments that allow high-resolution reconstructions of this warm
116 polar regime for which there is no modern analogue. Sediment shedding from
117 the south into the Colville Basin led to the development, throughout the Late
118 Cretaceous, of alluvial coastal plains incorporating the river-dominated Corwin
119 and Umiat Delta complexes (Huffman et al., 1985; Huffman, 1985). In the
120 west progradation was predominantly north and eastwards, while in the east
121 the coastal plain extended northwards forming what would later become the
122 central Arctic Slope of Alaska. The various marine and non-marine sediments
123 were originally divided into the Nanushuk and Colville Groups. The geology of
124 the Nanushuk Group is discussed in detail in Ahlbrandt (1979) and Huffman
125 (1985), while that of the Colville is presented in Brosgé and Whittington,

126 (1966) and Molenaar et al. (1987). Mull et al. (2003) revised the stratigraphic
127 nomenclature of the Arctic Slope and it is their scheme (Fig. 3) that is followed
128 here. For ease of referring to the older literature now abandoned
129 nomenclature is also used, but enclosed in square brackets. This convention
130 is incorporated in Figures 4, 5 and 7 that summarize the floral occurrences
131 and relationships. Late Cretaceous Arctic Slope sediments, as well as those
132 across N.E. Russia, consist of intertonguing marine and non-marine units
133 allowing for a rigorous biostratigraphic framework correlating both marine and
134 non-marine realms. In addition, abundant bentonites in the Late Cretaceous
135 have yielded some radiometric constraints (e.g. Bergman et al., 2006; Conrad
136 et al. 1992; Lanphere and Tailleux, 1983).

137

138 [Figure 3 near here]

139

140 The alluvial plain sediments of northern Alaska yield exceptionally rich
141 plant megafossil assemblages ranging in age from Albian (possibly Aptian)
142 through to Maastrichtian (Scott and Smiley, 1979; Smiley, 1966; Smiley, 1967;
143 Smiley, 1969a; Smiley, 1969b; Spicer, 1983; Spicer and Parrish, 1986;
144 Spicer, 2003; Spicer and Herman, 2001), the only major hiatus being the
145 result of a regional late Cenomanian/early Turonian transgression. These
146 non-marine Cretaceous successions also contain more than 2.7×10^{12} tonnes
147 of low sulphur, low ash coal (Sable and Stricker, 1987); reportedly one third of
148 all US coal reserves of all ages combined. By studying the assemblages in a
149 sedimentary facies context, the vegetational communities have been
150 reconstructed in detail and allowed the first quantitative estimates to be made

151 of Cretaceous near-polar air temperatures and precipitation (Spicer and
152 Parrish, 1986; Spicer, 1987; Parrish and Spicer, 1988a; Spicer and Parrish,
153 1990a).

154 Palaeosols are abundant in the Late Cretaceous successions of
155 northern Alaska and are characterised by gleyed colours, rare root traces but
156 abundant organic matter, abundant siderite, rare clay and silty clay coatings
157 and infillings, some preservation of sedimentary stratification and a general
158 lack of ferruginous oxides and mottles. They are interpreted as being poorly
159 drained, reducing, hydromorphic soils similar to modern Inceptisols (Ufnar et
160 al., 2004).

161 In the 1960s C. J. Smiley collected plant material from the central and
162 western North Slope of Alaska. However, although these and other collections
163 made by him formed the basis of a regional phyt stratigraphic scheme
164 (Smiley, 1966; Smiley, 1967; Smiley, 1969a; Smiley, 1969b), the floras were
165 never formally described, illustrated, or even listed. The only exception is a
166 summary paper by Scott and Smiley (1979). Moreover, Smiley was not able to
167 compare his material directly to that from North-eastern Russia and Russian
168 workers only had the names of some typical taxa reported by Smiley as their
169 Alaskan reference material. Now at the Smithsonian Institution National
170 Museum of Natural History, Smiley's collections have been studied and
171 photographed by us in the context of museum collections held in Russia and
172 our own extensive collections from both Alaska and N.E. Russia. In due
173 course all of this material, currently amounting to over 10,000 images, will be
174 made available on-line.

175 Based on a combination of Smiley's material, subsequent collections
176 from Alaska, and comparisons with Russian collections, Albian vegetation in
177 northern Alaska appears typically to have been dominated by deciduous
178 conifers such as *Cephalotaxopsis*, *Podozamites*, and *Pityophyllum* with
179 deciduous ginkgophytes (*Ginkgo* and *Sphenobaiera*) forming thickets along
180 river margins and in mires (Spicer and Herman, 2001). Ferns and *Equisetites*
181 were the main herbs with deciduous vine-like cycads, (quite unlike their
182 modern evergreen thermophilic relatives) occurring locally (Herman and
183 Spicer, 1996a).

184 This combination of Alaskan and N.E. Russian plant fossil
185 assemblages (taphofloras) has also allowed the development of
186 phyt stratigraphic correlations throughout the North Pacific region. Herman
187 (2007a, b, c) described a number of regional phyt stratigraphic units; each
188 unit was defined as plant-bearing deposits containing fossil floras of a certain
189 type characterised by their taxonomic composition determined by both
190 evolution and environment (and therefore representing a certain phase of
191 regional floral evolution). Each of these floral units, here termed a "regional
192 palaeoflora", is typified by a reference taphoflora, which is one of several
193 taphofloras with essentially similar characteristics that belong to that regional
194 palaeoflora. Within a given geographic range a regional palaeoflora has a
195 correlative function across formations and their subdivisions, while a
196 palaeofloral 'stage' reflects a phase of evolutionary development. For an
197 extended discussion of these concepts see Herman (2007a).

198 This approach is exemplified by taphofloras recovered from 34 localities
199 representing alluvial floodplain environments along the Kukpowruk River,

200 western North Slope Alaska (Spicer and Herman, 2001), that serve as the
201 type flora of the Kukpowruk floral stage and the Kukpowruk regional
202 palaeoflora. Other taphofloras that are assigned to the same floral stage are
203 the Kokolik, the Corwin Bluff, Pitmegea River, Punak Creek, taphofloras from
204 members 1–4 of the Kuk-Kaolak section, and by the Lower Killik taphoflora
205 from the Umiat-Chandler area (Fig. 4). Taken together these taphofloras point
206 to the Kukpowruk floral stage being taxonomically diverse with over 100 taxa
207 (Spicer and Herman, 2001; Herman and Spicer, 2002) including liverworts (?),
208 sphenophytes, ferns, cycadophytes, czekanowskialean, caytonialean (?),
209 conifers, gymnosperms of unknown affinity, angiosperms and remains that so
210 far elude identification.

211

212 [Figure 4 near here].

213

214 The lower boundary of the Kukpowruk floral stage in the Umiat-
215 Chandler area of the Arctic Slope directly overlies the marine beds of the
216 Nanushuk Formation [Tuktu Formation] containing *Gastroplites*, *Cleoniceras*,
217 *Inoceramus anglicus* Woods and *I. Cadottensis* McLearn (Detterman et al.,
218 1963, Smiley, 1969a), which indicate a basal middle Albian age. In the Kuk-
219 Kaolak area *Entolium* sp. in the upper part of Smiley's (1966) Lithological Unit
220 1 (Fig. 4) underlies the plant-bearing beds and suggests an age no older than
221 middle Albian (Imlay, 1961, cited in Smiley, 1966). Further to the west in the
222 Utukok-Corwin region the middle Albian *Paragastroplites spiekeri* McLearn is
223 found in the upper part of the Lower Nanushuk Formation [Kukpowruk

224 Formation] (Chapman and Sable, 1960). Thus the Kukpowruk floral stage is
225 likely to have begun in middle Albian time.

226 The upper age limit of the Kukpowruk floral stage is in the uppermost
227 part of the Albian because regionally rocks containing the floral stage are
228 conformably overlain by deposits containing taphofloras of the Niakogan floral
229 stage, which in turn is correlative with the Albian – Cenomanian Grebenka
230 flora of the Anadyr-Koryak subregion. The Grebenka flora is well dated by
231 both marine biostratigraphic and isotopic ($^{40}\text{Ar}/^{39}\text{Ar}$) methods (Spicer et al.
232 2002). Herman (2007b) estimated the duration of the Kukpowruk floral stage
233 to be 8-9 m.y.. The ages and duration of other Alaskan and N.E. Russian
234 floral stages and horizons are similarly constrained and the correlations are
235 summarised in Figures 4-7.

236 Detailed recording of repeated plant associations in sedimentary
237 context has not only provided the basis for phytostratigraphic correlation, but
238 also an understanding of ancient plant community succession. For the
239 Kukpowruk flora this typically began with *Equisetites* colonising bare river
240 bank substrate surfaces, quickly joined by the fern *Birisia* which often formed
241 expansive *Birisia* marsh 'heaths' in low-lying interfluvial areas. As seral
242 development continued *Arctopteris* (a fern) and the cycadophyte *Nilssonia*
243 typically would then dominate leaving *Equisetites* and *Birisia* as subordinate
244 elements. River margins were usually dominated by ginkgophytes, but in mire
245 environments stable state communities culminated in a *Podozamites* and
246 *Pityophyllum*-dominated shrub 'woodland' with *Birisia* and *Equisetites*
247 remaining as the main understorey components. In drier settings a more
248 diverse community developed, again dominated by the conifer-like

249 *Podozamites* and *Pityophyllum*, but with *Desmiophyllum* and sometimes
250 *Arctopteris* as typical subordinate elements joining *Birisia* and *Equisetites*
251 (Spicer and Herman, 2001)

252 Overall the Kukpowruk and co-eval taphofloras in northern Alaska (Fig.
253 4) show high levels of similarity with the Buor-Kemyus regional palaeoflora in
254 N.E. Russia (Figs. 5 and 7).

255

256 [Figure 5 near here]

257

258 Flowering plants migrated into Northern Alaska in the late Albian from lower
259 latitudes (Scott and Smiley, 1979; Spicer and Parrish, 1986; Spicer, 1987)
260 and began to oust the ginkgophytes from their riparian habitats. Typical of the
261 earliest angiosperms to appear in northern Alaska are those found in the
262 youngest (latest Albian to early Cenomanian Nanushuk Formation [Corwin
263 Fm.] of the Kukpowruk River assemblages (Spicer and Herman, 2001). As in
264 Russia, the migration pathways of these opportunists appear to have been
265 along coastal margins and along disturbed river habitats (Herman, 2002). In
266 Alaska this is evident both from channel sands in the Nanushuk Formation
267 (later Kukpowruk and Niakogon regional palaeofloras in Fig. 4), as well as in
268 the Cenomanian and Turonian fluvial deposits of the Yukon-Koyukuk Basin
269 (Hollick, 1930; Spicer, 1983). Platanoid (sycamore)-like leaves are the most
270 abundant angiosperm megafossils in channel sands, crevasse splay and
271 overbank settings in latest Albian through the Cenomanian. As the Late
272 Cretaceous progressed evolutionary novelty, combined with a low competitive
273 stress from the non-angiosperms, polyploidy and hybridization (Spicer, 1986)

274 led to a diversity of flowering plant leaf forms occupying the full range of
275 alluvial floodplain environments except mires. However, diversity at the family
276 level was low and no major clades (families or higher taxonomic ranks)
277 appear to have originated in the high Arctic (Spicer et al., 1987; *contra* Hickey
278 et al., 1983).

279 Fossil trees, many in growth position and up to 70 cm in diameter, are
280 abundant in Nanushuk [Niakogon Tongue] non-marine rocks and cell details
281 are often preserved (Spicer, 2003). Productivity evidently was high and tree
282 ring characteristics show that intra-annual environmental variation was small.
283 Seasonality, however, was extreme and typically the rings exhibit sudden
284 transitions from earlywood to latewood indicating a short autumnal period, and
285 small quantities of latewood suggest deep dormancy. Here, as in previous
286 studies of Alaskan North Slope woods (Parrish and Spicer, 1988b; Spicer and
287 Parrish, 1990a), latewood is differentiated from earlywood using the classical
288 definition of these terms as originally proposed by Mork (1928) and clarified
289 by Denne (1989). The concept of earlywood used by Creber and Chaloner
290 (1984) and Falcon-lang (2005) has been criticized in the context of
291 characterizing high latitude light regimes (Taylor and Ryberg, 2007) and is not
292 used here. Evidently the transition from full summer growth to winter
293 dormancy was rapid (< 3 weeks), consistent with a high palaeolatitude and
294 obliquity angle essentially similar to that of the present (Parrish and Spicer,
295 1988b).

296 Palaeogeographic reconstructions (Smith et al., 1981; Ziegler et al.,
297 1983), and palaeomagnetic data from the Nanushuk Formation itself (Witte et
298 al., 1987) indicate the area was at least 75 °N during the Cenomanian

299 although some reconstructions place it closer to the pole (Hay et al., 1999;
300 <http://www.odsn.de/>)(Fig.1). A relative northward drift of Northern Alaska
301 through the Late Cretaceous positioned the Maastrichtian northern coastline
302 of Alaska at approximately 85 °N (Smith et al., 1981; Ziegler et al., 1983).
303 Best fit distribution of coals and other climatically sensitive sediments show
304 that the rotational pole was within 4° of the palaeomagnetic pole in the
305 Maastrichtian (Lottes, 1987). Thus throughout the Late Cretaceous the
306 Northern Alaska winter, assuming a minimum palaeolatitude of 71 °N to a
307 maximum of 89 °N, was characterised by between almost 2 months of
308 continuous twilight to almost 5 months of continuous darkness bounded by a
309 total of three months of bounding twilight (Anonymous, 1978)(Fig. 8). This
310 extended period of darkness is consistent with the small latewood
311 components of the tree rings and rapid transitions from earlywood to latewood
312 cell formation.

313 Wood anatomy (Parrish and Spicer, 1988b), vegetation composition
314 and angiosperm (dicot) leaf margin analysis indicates that the Cenomanian
315 mean annual air temperature on the coastal plain was approximately 10 °C.
316 The mean annual range of temperature could have been over 20 °C, but cold
317 month means were no lower than -11 °C (Spicer and Parrish, 1986; Parrish et
318 al., 1987; Spicer and Parrish, 1990a) and could well have been warmer. Large
319 ranges of leaf size, coupled with lack of thick cuticles and lack of damage to
320 wood cells, suggests evapo-transpirational stress was low at all times during
321 the growing season, including the early spring. Coupled with sedimentological
322 data indicating a lack of periglacial features, this indicates that freezing of the
323 root zones did not occur to any extent.

324 Precipitation was high as evidenced by exceptional quantities of coal
325 accumulation. Contrary to early USGS reports, pollen preservation in the
326 coals and associated clastic facies can be good (Youtcheff et al., 1987) and
327 preparations have yielded significant pollen stratigraphic changes on the
328 millimetre scale within coals (Grant et al., 1988). Many mires, particularly in
329 the Campanian and Maastrichtian, were undoubtedly low lying and
330 maintained by a high regional water table. In this sense they were true
331 (rheotrophic) swamps and coal deposition was associated with frequent
332 influxes of siliciclastics. In late Albian to Coniacian times, however, seams are
333 commonly several metres thick with low ash contents. Chemical and
334 palynological analyses suggest internal nutrient cycling and a gradual
335 reduction in diversity as mire development progressed (Youtcheff et al.,
336 1987). Taken together these signatures are indicative of raised (ombrotrophic)
337 mires maintained by high rainfall. High precipitation, however, did not prevent
338 frequent, low intensity, short duration wildfires as evidenced by numerous
339 millimetre-scale fusain (charcoal) horizons throughout the coals. Peat
340 accumulation was also supported by high summer productivity coupled with
341 temperature-mediated minimal decay of shed organic material during the
342 winters (Spicer et al., 1992).

343 Plant megafossils from the Tuluvak Formation (the [Tuluvak Tongue] of
344 the now revised Prince Creek Formation), central North Slope (Fig. 4), occur
345 in upper alluvial plain bentonitic clays, siltstones and coals and are best
346 preserved in siltstones and siderite nodules within the clays. They are
347 regarded as Coniacian in age based on marine biostratigraphic control
348 (Brosgé and Whittington, 1966; Gryc et al., 1951; Mull et al., 2005) as well as

349 floral composition. The palaeolatitude, based on Nanushuk palaeomagnetic
350 determinations (Witte et al., 1987) and plate reconstruction of (Smith et al.,
351 1981) has been estimated at 75 °N, but could be higher (Fig. 1). Leaf margin
352 analysis (LMA) of Coniacian Alaskan material suggests mean annual
353 temperatures (MATs) may have been as high as 13 °C (Parrish and Spicer,
354 1988a), while an early multivariate foliar physiognomic study suggests a MAT
355 of 12.5 °C with a warm month mean temperature (WMMT) of 20.0 °C and a
356 cold month mean temperature (CMMT) of 5.7 °C (Herman and Spicer, 1997).
357 Coniacian standing forests are known from northern Alaska, although
358 anatomical details of the stumps are poorly preserved (unpublished personal
359 observation by RAS). Angiosperms show a mixture of toothed “platanoid”
360 forms and entire-margined taxa that await formal description but are illustrated
361 in Herman and Spicer (1997).

362 Northern Alaska woody angiosperm floral diversity, like that of the late
363 Cenomanian, was high, but then declined throughout the remainder of the
364 Late Cretaceous as a consequence of increased specialisation and taxonomic
365 canalisation (Spicer, 1986) coupled with climatic deterioration (Spicer and
366 Parrish, 1987; Spicer, 1987). North Slope palaeobotanical studies of the Late
367 Campanian to latest Maastrichtian parts of the Prince Creek Formation
368 [Kogosukruk Tongue] (palaeolatitude 83 - 85 °N) indicate that the MAT was
369 between 2.5 and 5 °C (Spicer, 1987; Parrish and Spicer, 1988a), summer
370 temperatures were lower than in the Cenomanian and Coniacian, and tree
371 ring data suggest that intra- and inter-annual environmental variation was
372 marked (Spicer and Parrish, 1990a). Angiosperm leaf diversity from Alaskan
373 North Slope Maastrichtian localities is too low to attempt a multivariate

374 physiognomic analysis so estimating winter temperatures is problematic, but
375 again there is no sedimentary evidence for periglacial conditions at, or near,
376 sea level and no frost rings have been observed in fossil trees.

377 To the south in Denali National Park Tomsich et al. (this volume) report
378 on a lower Cantwell Formation Maastrichtian floral assemblage that is more
379 diverse consisting, primarily, of taxodiaceous conifers, ferns, *Equisetites*, rare
380 ginkgophytes and cycad fragments together with a woody angiosperm
381 component sufficient for a CLAMP analysis. The composite flora yields a MAT
382 of 7.4 ± 2.4 °C, a WMMT of 17.1 ± 3.2 °C and a CMMT of -2.3 ± 3.8 °C. The
383 growing season was just 4.8 ± 1.4 months (all uncertainties here are 2
384 standard deviations of the residuals about the calibration regressions).

385 Increasingly abundant fossil charcoal and a diminution of coal
386 abundance and quality suggests later Cretaceous vegetation experienced
387 frequent wildfires and, although mud cracks are absent from the sedimentary
388 record (Phillips, 1987, 1988, 2003), the annual distribution of precipitation may
389 have been less uniform than in the mid Cretaceous.

390 Pollen data from the Maastrichtian parts of the Prince Creek Formation
391 reveals a higher diversity of angiosperms than is evident from the megafossils
392 (Frederiksen et al., 1988) and the spatial and stratigraphic heterogeneity are
393 indicative of herbaceous types, probably insect pollinated, with a short life
394 cycle and rapid evolutionary turnover. Trees were small in the latest
395 Cretaceous (Spicer and Parrish, 1987; Spicer and Parrish, 1990b) with forests
396 probably being more open and dominated by deciduous taxodiaceous
397 conifers such as *Parataxodium wigginsii* Arnold and Lowther. This taxon is
398 similar to modern *Metasequoia*, but regarded as sufficiently distinct to warrant

399 a separate genus (Arnold and Lowther, 1955). Ferns and *Equisetites* still were
400 common as ground cover but herbaceous, possibly annual, flowering plants
401 were also abundant. Overall the vegetation structure may have approached
402 that of modern Taiga and been similar in structure to that seen in central
403 Alaska today. Some rare occurrences of the wood type known as *Xenoxylon*
404 *latiporosum* (Cramer) Gothan in the late Maastrichtian of northern Alaska
405 exhibit ring characteristics similar to those of the Cenomanian and may be
406 indicative of short-lived warm episodes just prior to the end of the Cretaceous.
407 Such Maastrichtian warming episodes have been reported from lower
408 latitudes in N. America (Wolfe and Upchurch, 1987).

409

410 **3. Cretaceous Floras from Northeastern Russia**

411

412 As far as can be determined, non-marine floristic successions from
413 N.E. Russia have much in common with those of Alaska, but critical to any
414 North Pacific/Arctic regional studies of vegetation patterns and climate change
415 is reliable correlation facilitated by precise dating. Cretaceous floras from
416 northeastern Russia (Figs. 5 and 6) exhibit much the same vegetational
417 composition as those in Alaska (e.g. Lebedev, 1976; Samylina, 1974;
418 Herman, 1991; Herman, 1993; Herman and Lebedev, 1991; Herman and
419 Shczepetov, 1991; Samylina, 1973; Shczepetov, 1991; Spicer et al., 2002b),
420 although the flowering plants may appear slightly earlier.

421 Pre-angiosperm floras are in the most part conifer-dominated and
422 typified by the Aptian Silyap and the Albian Buor-Kemyus regional
423 palaeofloras (Fig. 5). Most relevant to Alaska is the Buor-Kemyus floral type

424 known from several locations, the richest being the Buor-Kemuss Formation
425 of the Zyrianka Coal Basin (Kolyma River Basin: Prynada, 1938; Samylina,
426 1964; Samylina, 1967), the upper subformation of the Ainakhkurgen
427 Formation and the Chimchememel Formation of the Ainakhkurgen
428 Depression, Bolshoi Anui River basin (Filippova, 1975; Samylina, 1976), the
429 Umkuveem, Marakveem and Irvunei Formations of the Umkuveem
430 Depression, Yeropol River basin (Filippova, 1975; Samylina, 1976) and the
431 middle and upper subformations of the Omsukchan and Toptan Formations of
432 the Omsukchan area near Magadan (Samylina, 1976).

433 The Buor-Kemyus regional palaeoflora is characterised by fern-rich
434 associations (typically with *Osmunda*, *Birisia*, *Coniopteris*, *Arctopteris*,
435 *Cladophlebis*, *Hausmannia*, *Asplenium*, *Onychiopsis*, *Adiantopteris*, etc.)
436 forming the ground cover together with sphenophytes (*Equisetites*) and
437 cycadophytes (*Nilssonia*, *Nilssoniopteris*, *Anomozamites*, *Neozamites*,
438 *Heilungia*, *Butefia* and *Taeniopteris*). Larger stature plants are represented by
439 the czekanowskialeans (leaf taxa *Czekanowskia* and *Phoenicopsis* together
440 with the reproductive structure *Leptostrobus*). Both 'ancient' conifers
441 (*Podozamites* and *Pityophyllum*) together with conifers of a more modern
442 aspect (e.g. *Sequoia*, *Cephalotaxopsis*, *Pityostrobus*, *Pityospermum*) made
443 up the forest canopy in late seral stages. In total over 150 species have been
444 reported as comprising the Albian Buor-Kemyus flora to which the 57 species
445 found in the Kukpowruk flora have been compared (Spicer and Herman,
446 2001).

447 Correlative with floras of the widespread Buor-Kemyus type are those
448 of the Early Ginter regional palaeoflora (Fig. 6) of northeastern Russia. This is

449 typified by the assemblages of the Ginter Formation at Ugol'naya Bay that
450 occur below a 330 m thick marine unit containing the Cenomanian
451 *Inoceramus pennatulus* zone (Herman, 2007a). Early Ginter floras lack
452 angiosperms and have a prevalence of taxa characteristic of the Early
453 Cretaceous such as *Hausmannia* ex gr. *bilobata* Pryn. *Cladophlebis* aff.
454 *williamsonii* (Brongn.) Brongn. *Sagenopteris*, *Neozamites* (?) and
455 *Podozamites*.

456

457 [Figure 6 near here]

458

459 Herman and Spicer (this volume) correlate the K-Ar dated (Kuzmichev
460 et al., 2009) late Albian Balyktakh Flora of Novaya Sibir Island with both the
461 N.E. Russian Buor-Kemyus Flora and the late Albian to Cenomanian
462 Kukpowruk Flora of western North Slope, Alaska (Smiley, 1969b; Spicer and
463 Herman, 2001).

464 Previous non-marine stratigraphy in N.E. Russia, and by extension to
465 some extent other parts of the Russian Federation, was in part based on floral
466 correlation with the Yukon River assemblages of the now abandoned
467 Shaktolik Group (Hollick, 1930). Krassilov (1975) assumed an Albian age for
468 the Yukon Floras and used this to date many eastern USSR floras. However
469 Spicer (1983) showed that the Yukon assemblages were late Cenomanian
470 (cut into by some Turonian (?) channels) and correlative with the Cenomanian
471 [Niakogon Tongue] of the Nanushuk Formation.

472 Also coeval with the Nanushuk is the highly diverse (84 angiosperm
473 taxa) taphoflora of the Yelisseev locality along the banks of the Grebenka

474 River (Filippova, 1979, 1988, 1989; Herman and Spicer, 1999; Spicer et al.,
475 2002a) that typifies the Grebenka regional palaeoflora found throughout N.E.
476 Russia (Figs. 5 & 6). Deposited rapidly at palaeolatitude 72 °N on the
477 floodplains to the south east of the Ochotsk-Chukotka Volcanic Belt (OCVB)
478 highlands, assemblages at the Yelisseev locality have been $^{40}\text{Ar}/^{39}\text{Ar}$ dated as
479 96.8 Ma and are thus Cenomanian in age (Spicer et al., 2002). This age
480 determination is consistent with biostratigraphic dating of marine units above
481 and below the plant bearing succession that lies within the Krivorechenskaya
482 Formation, as well as floristic and palynological studies (summarised in Spicer
483 et al., 2002).

484 Regionally the Grebenka floral type is highly diverse with over 200
485 taxa, of which angiosperms typically make up 35-40%, and conifers and ferns
486 20-25%. Locally fern remains dominate the assemblages. Oversupply of
487 volcanoclastic sediments to the Grebenka (Yelisseev locality) ancient
488 floodplain led to a dynamic sedimentation regime that captured and preserved
489 plant remains deposited in a suite of different settings. This has allowed
490 detailed reconstruction of the seral development of this vegetation, as well as
491 its composition and structure. Material retrieved from the Yelisseev site is
492 dominated by angiosperms, which comprise more than 50% of the species
493 present, followed in order of decreasing diversity by conifers, ferns,
494 sphenophytes etc.. *Menispermites*, *Platanus*, *Araliaephyllum*,
495 *Scheffleraephyllum*, *Dalembia*, and *Grebenkia* are the most abundant
496 angiosperm genera. Less abundant are *Trochodendroides*, *Diospyros*,
497 *Myrtophyllum* and *Cissites*, and several fossil leaves of dicots that are
498 tentatively assigned to the form genus *Dicotylophyllum*.

499 Entire-margined leaves (*Magnoliaephyllum*, *Myrtophyllum*,
500 *Scheffleraephyllum*, etc.) are surprisingly numerous given that these are
501 usually associated with warm climatic regimes (Bailey and Sinnott, 1915,
502 1916; Wilf, 1997; Wolfe, 1979). Lobed (*Platanus*, *Cissites*, *Menispermities*,
503 *Araliaephyllum*) and compound (*Scheffleraephyllum*, *Dalembia*, *Sorbites*,
504 *Sapindophyllum*) forms also occur.

505 The mesic conifer foliage form *Cephalotaxopsis*, especially *C.*
506 *intermedia*, predominates over the more xeromorphic *Araucarites*, *Elatocladus*
507 and *Pityophyllum*. Shoots of *Araucarites anadyrensis* are commonly found
508 associated with elongate female cones and with tree trunks up to 40 cm in
509 diameter. *Florinia*, *Pagiophyllum*, *Tollia*, and *Sequoia* also occur. Like
510 *Araucarites*, *Pagiophyllum triangulare* represents shoots with rigid scale-like
511 leaves and, based on its morphology, is likely to have been evergreen.

512 *Coniopteris* and *Birisia* are the most ubiquitous ferns, with *Gleichenia*,
513 *Gleichenites*, *Hausmannia*, *Arctopteris*, *Asplenium* and *Cladophlebis* also
514 being common.

515 Cycadophytes are numerous and the deciduous shoot-shedding
516 *Nilssonia* is the most diverse among them. Bedding planes occasionally have
517 abundant leaf accumulations of *Nilssonia alaskana* and *Cycadites*
518 *hyperborea*. A short shoot of *Nilssoniocladus chukotensis* Spicer et Herman
519 with three *Nilssonia* leaves attached has been found at the locality (Spicer
520 and Herman, 1996). *Taeniopteris* remains occur but are not common.

521 Remains of the Ginkgoales, i.e., the genera *Ginkgo* and *Sphenobaiera*,
522 are found frequently, whereas *Pseudotorellia* (?) fossils are rare.

523 Czekanowskiales are represented by very rare impressions of *Phoenicopsis*

524 ex gr. *angustifolia*, and the Caytoniales by the polymorphic leaflets of
525 *Sagenopteris*.

526 The pattern of seral development in the Grebenka Flora may be
527 summarized as beginning with the putative bryophytes *Thallites* and
528 *Marchantites* colonizing bare moist sub-aerial sediment, and the fern
529 *Hausmannia* growing in ponds and lakes. As with earlier pre-angiosperm
530 floras *Birisia* and *Equisetites* formed an herbaceous ground cover community.
531 On drier substrates *Birisia* was, typically, then joined by the fern *Coniopteris*
532 and sometimes by *Glechenites*. Significantly this seral stage included woody
533 deciduous flowering plants represented by the leaf genera *Platanus*,
534 *Grebenkia* and more rarely *Sheffleraephyllum*. Deciduous *Ginkgo* and
535 *Nilssonia* formed subordinate components. In wetter sites *Cycadites*
536 dominated this seral stage.

537 *Coniopteris*, *Birisia*, *Nilssonia*, *Grebenkia*, *Sheffleraephyllum* and
538 *Ginkgo* then persisted while being joined by the conifers *Cephalotaxopsis* and
539 *Elatocladus smittiana*, and the angiosperms *Araliaphyllum*, *Menispermities*
540 and *Dalembia*. *Cephalotaxopsis* and *E. smittiana* then went on to dominate
541 the community with *Pityophyllum* as an occasional subordinate component.
542 Loss of diversity at this stage suggests, perhaps, closure of the canopy. In
543 drier sites the xeromorphic conifers *Pagiophyllum*, *Araucarites* and *Sequoia*
544 *reichenbachii* finally dominate the mature forest community. These taxa are
545 not present in wetter environments, which are typified by *Cephalotaxopsis* and
546 *Elatocladus*. Evidence for this seral interpretation and illustrations of the foliar
547 forms are given in Spicer et al. (2002).

548 Many of the Grebenka taxa are common to the Nanushuk and our
549 understanding of the Grebenka vegetation structure and dynamics (seral
550 development) also applies to floras of the Nanushuk Formation. One
551 noticeable difference, however, is that the most mature Grebenka forests
552 contain a high proportion of microphyllous and xeromorphic conifers (e.g.
553 *Pagiophyllum* and *Araucarites*) that may well have been evergreen. These
554 forms are not characteristic of the Nanushuk vegetation where only very rarely
555 microphyllous cupressaceous forms (likely to have been evergreen) are
556 found. This may be indicative of a slightly lower palaeolatitude (72 °N) and/or
557 lower winter temperatures because cold depresses respiratory metabolic
558 rates and makes retention of leaves an energetically viable strategy for over-
559 wintering (Herman and Spicer, 1999).

560 It is unlikely that the Nanushuk floodplain was a more disturbed
561 environment (thereby preventing the establishment of mature forests)
562 because the Grebenka succession shows evidence of rapid deposition and
563 dynamic changes in fluvial regime (Spicer et al., 2002). Edaphic factors were
564 likely also to have been important here in that the highly volcanoclastic,
565 predominantly coarse-grained, Grebenka suite of sediments may, overall,
566 have included plant material derived from sites with a similar coarse substrate
567 that were better drained than on the Nanushuk alluvial plain where ash falls
568 were finer grained and consequently less permeable.

569 Previous palaeoclimatic reconstructions for the Grebenka Flora as a
570 whole suggest a MAT of 13 °C, a WMMT of 20.8 °C and a CMMT of 5.5 °C.
571 Because this is one of the most diverse Late Cretaceous Arctic angiosperm
572 floras (84 morphotypes) and was deposited rapidly (probably <100 kyr) it is

573 regarded as one of the most reliable palaeoclimatic determinations. For a
574 detailed account of the flora, age determination and sedimentary context see
575 Spicer et al. (2002).

576 An intriguing, important, but under-studied Russian flora that post-
577 dates both the Nanushuk and Grebenka Floras is that of the Turonian
578 Derevyannye Gory Section on Novaya Sibir' Island. The sedimentary
579 succession is rich in volcanoclastics, lacustrine deposits and unconsolidated
580 peats and contains an abundant and diverse flora of about 50 species
581 belonging to 33 genera made up of angiosperms, conifers, ferns and
582 ginkgophytes. This, together with the nearby Albian Balyktakh Formation of
583 Kotel'nyi Island, probably represents the warmest, highest latitude, floras of
584 the Cretaceous. Previous palaeoclimatic interpretations of this flora based on
585 20 angiosperm morphotypes suggested a MAT of 9 °C, a WMMT of 18.5 °C
586 and a CMMT of 0 °C. (Herman and Spicer, 1997). A more complete and
587 revised treatment is presented in Herman and Spicer (this volume).

588 As with that of the Grebenka River, another Russian floral assemblage
589 that exhibits a relatively high proportion of xeromorphic elements is the
590 Chauna (sometimes referred to as the Palyavaam) Flora. Originally regarded
591 as 87-100 Ma in age based on Rb-Sr dates and phyt stratigraphy (Belyi,
592 1977; 1997; Samylina, 1988) it is floristically distinct from other floras from
593 N.E. Russia and thus its age and correlation with other floras in the region has
594 proved somewhat problematic. Despite this, Samylina (1974; 1988) proposed
595 a late Albian age based on the presence of Early Cretaceous forms such as
596 *Tchaunia*, *Kolymella*, various species of the fern *Cladophlebis*, *Sphenobaiera*,
597 *Ctenis*, *Heilungia*, *Phoenicopsis*, *Elatocladus* and *Araucarites*, mixed with

598 some younger taxa that appear in the Late Cretaceous such as *Picea*,
599 *Sequoia cf. reichenbachii*, *Menispermites* and *Ziziphoides*. Belyi (1977; 1997)
600 suggested it was correlative with floral zone II of the Chandler River
601 succession (Smiley, 1969a), in Northern Alaska. Shczepetov (1991) proposed
602 a younger Cenomanian to Turonian age, but subsequently revised his
603 estimate to “early Late Cretaceous”. The Chauna volcanic sediments have
604 since been $^{40}\text{Ar}/^{39}\text{Ar}$ dated using plagioclase feldspar, biotite and amphibole
605 (Kelley et al., 1999). All measurements lie within 2_ errors of each other and
606 yield a mean age of 88 Ma, and therefore are indicative of a Coniacian age
607 (Gradstein et al., 2005). This is more consistent with the angiosperm
608 component of the flora that, as well as *Menispermites* and *Zizyphoides*,
609 contains *Cinnamomoides*, *Trochodendroides*, *Quereuxia* and leaves included
610 in the form genus *Dicotylophyllum*, all of which are indicative of a mid Late
611 Cretaceous age. The original dating partly arose because of the highly
612 endemic nature of the flora, the uniqueness of which is possibly due to the
613 highly volcanoclastic nature of the palaeosols and thus is primarily edaphic.
614 There is also the possibility that the flora was deposited at moderate elevation
615 within the Ochotsk-Chukotka volcanic highlands (Kelley et al., 1999).

616 The Arman Flora comes from terrigenous and volcanoclastic deposits of
617 the Arman and Naraula formations of the OCVB in N. E. Russia. According to
618 it’s systematic composition it is Turonian-Coniacian in age (Herman, 2004).

619 $^{40}\text{Ar}/^{39}\text{Ar}$ data provide similar results in that the Naraula Formation has been
620 variously dated as 86 ± 1 , 85.6 ± 1.3 and 85.5 ± 1.1 Ma (Hourigan and Akinin,
621 2004), whereas U-Pb SHRIMP analysis of zircons from the Arman Formation
622 yields an age of 92 ± 2 Ma (Akinin, 2007). The Arman Flora has so far been

623 shown to contain about 80 species of sphenophytes, ferns, ginkgoaleans,
624 czekanowskialeans, conifers and angiosperms, with *Birisia ochotica* (a fern)
625 and *Podozamites*, *Cephalotaxopsis* and *Sequoia* (conifers) being the most
626 abundant. Among angiosperms (26 species), *Araliaephyllum*,
627 *Scheffleraephyllum*, *Dalembia*, *Menispermites*, *Trochodendroides*, *Ziziphoides*
628 and platanoid leaves are numerous. In terms of systematic composition the
629 Arman Flora is similar to the Turonian Kaolak Flora and, to a lesser extent, to
630 the Coniacian Tuluvak Flora of northern Alaska.

631 Recently (Belyi, 2008) has argued that discrepancies between isotopic
632 dates and geological data within the OCVB means that the isotopic systems
633 clock might have been reset in the region, particularly in respect of $^{40}\text{Ar}/^{39}\text{Ar}$
634 and U-Pb (SHRIMP) methodologies, and that it is premature to abandon the
635 older chronostratigraphic framework. He cites, amongst other data, the
636 mismatch between regionally obtained Rb-Sr dates ranging from 104 ± 3 to
637 93 ± 5 Ma (Kotlyar and Rusakova, 2004) and the $^{40}\text{Ar}/^{39}\text{Ar}$ and U-Pb
638 (SHRIMP) results (e.g. Akinin et al., 2006; Akinin and Khanchuk, 2005;
639 Hourigan and Akinin, 2004; Ispolatov et al., 2000; Kelley et al., 1999;
640 Tikhomirov et al., 2006) that tend to yield younger ages. Given the known
641 susceptibility of the Rb-Sr system to chemical alteration this mismatch is not
642 too surprising, but clearly more detailed isotopic work needs to be conducted
643 in the area to resolve the disparities.

644 Additional angiosperm-rich floras on the margin of the OCVB come
645 from the Northern Pekulney Range (Craggs, 2005; Terekhova and Filippova,
646 1983; Terekhova and Filippova, 1984) in the Anadyr-Koryak region of N.E.
647 Russia. Two main assemblages are known: the Tylpegyrgynai and the

648 Poperechnaya, named after the formations in which they occur, and both
649 suggest a predominantly deciduous phenology with few evergreen elements.
650 The compositions of these assemblages are similar, but sufficiently different
651 for them to be treated separately (Craggs, 2005). Both are considered to be
652 early Coniacian in age based on biostratigraphy of the underlying (in the case
653 of the Poperechnaya) and overlying (for both the Poperechnaya and
654 Tylpegyrgynai Formations) marine units (Fig. 6). The Tylpegyrgynai
655 assemblage is dominated by angiosperms (40% of taxa) with less diverse
656 conifer and fern components. Cycadophytes, ginkgophytes, sphenophytes,
657 lycopods and probably liverworts complete the plant spectrum. The
658 Poperechnaya assemblage is even more angiosperm dominated (50% of
659 taxa), but otherwise the composition is similar to that of the Tylpegyrgynai
660 Formation. For a full account of the Northern Pekulney Range floras see
661 (Craggs, 2005). A comprehensive analysis of the angiosperm morphotypes is
662 given on the CLAMP website ([http://www.open.ac.uk/earth-](http://www.open.ac.uk/earth-research/spicer/CLAMP/Clampset1.html)
663 [research/spicer/CLAMP/Clampset1.html](http://www.open.ac.uk/earth-research/spicer/CLAMP/Clampset1.html)).

664 The Vilui Basin in central Siberia accumulated sediments throughout
665 the Late Cretaceous providing a window into continental interior high latitude
666 palaeoenvironments at that time. The abundant flora (Budantsev, 1968;
667 Vakhrameev, 1958; Vakhrameev and Pushczarovskii, 1954) has been
668 collected from the Timerdyakh Formation, the sediments of which are typically
669 channel sandstones interbedded with floodplain silts and clays,
670 autochthonous coals and stacked palaeosols (Spicer et al., 2008). This is an
671 exceptionally diverse flora with over 170 pollen and spore taxa of which 61
672 are spores (at least 33 represent ferns), 14 are gymnosperm pollen, 10 are

673 from monocots (including two probable primitive palms) and approximately 90
674 are pollen from dicot flowering plants. The grains are well preserved (Hoffman
675 and Zetter, 2007) and no more than 10% appear to have been reworked from
676 older sediments. Some of the diversity is a result of the long time interval over
677 which the formation was deposited (latest Albian to Maastrichtian) and the fact
678 that even very small (<10 μ m) grains were recovered by not sieving the
679 preparations. Nevertheless the Cenomanian megafossil floras from the banks
680 of the Tyung River are also diverse and previous analyses of the dicot leaves
681 indicated a MAT of 13.1 °C and WMMT of 21.1°C and CMMT of 5.8 °C. These
682 temperatures are consistent with the overall composition of the flora and the
683 clay mineralogy. Based on weak magnetisation of the Timerdyakh sediments
684 a palaeolatitude of 71.5 - 73.2 °N has been suggested (Spicer et al., 2008) but
685 the Scotese reconstructions estimate a slightly lower palaeolatitude of 68 °N.
686 The length of the growing season (daily temperatures above 10 °C) is
687 estimated to have been 7.4 \pm 1.7 months (Spicer et al., 2008). Despite the
688 continental interior position there is no sedimentological or botanical evidence
689 for freezing.

690 The Valizhgen Formation from northwestern Kamchatka is made up of
691 coarse-grained sandstones together with sandy-clayey and coaly-clayey
692 shales and contains numerous plant fossils that have been divided into a
693 Coniacian Kayvayam Flora and a Turonian Penzhina Flora (Herman, 1991)
694 (Fig. 6). These were combined with co-eval floras from the Valizhgen
695 Formation on the Yelistratov Peninsula to form separate Coniacian and
696 Turonian Kamchatkan floras, both at palaeolatitudes between 70 and 75 °N,
697 that could be used for palaeoclimatic studies. Angiosperm morphotypes used

698 in these analyses are illustrated in Herman and Spicer (1997). Multivariate
699 physiognomic analysis for the Turonian flora indicated a MAT of 6.9 °C with a
700 WMMT of 17.8 °C and a CMMT of -3.8 °C, while the Coniacian flora was
701 interpreted to have grown under a MAT of 9 °C, a WMMT of 18.6 °C and a
702 CMMT of 0 °C (Herman and Spicer, 1997).

703 The latest Cretaceous Kakanaut Flora of the Koryak upland region of
704 N.E. Russia (Fig. 6) is of particular interest because of its association with a
705 latest Cretaceous dinosaur fauna, including remnants of eggs (Godefroit et
706 al., 2009). The plant fossil assemblages are richer than those of the co-eval
707 dinosaur-yielding alaskan Prince Creek Formation (Brouwers et al., 1987;
708 Clemens, 1985; Fiorillo and Gangloff, 2000), but not, as claimed by Godefroit
709 et al. (2009) “the richest Cretaceous palaeoflora in the Arctic region”, as will
710 be evident from the works reviewed here. Nevertheless, over 50 megafossil
711 plant taxa have been reported (Golovneva, 1994a; Golovneva, 1994b)
712 including locally abundant *Ginkgo*, and the cycadophytes *Encephalartopsis*
713 and *Nilssonia* forming monodominant associations. Within the 30 or so
714 angiosperm taxa recognised are *Peculnea*, *Cissites*, “*Corylus*”, *Celastinites*,
715 platanoids, *Trochodendroides* and *Fagopsiphyllum*. Entire-margined forms
716 make up about 15% of the woody flowering plant leaves. Conifers are
717 assigned to the Taxodiaceae, Cupressaceae and Pinaceae. Godefroit et al.
718 (2009) report that of the 10 gymnosperm taxa identified four are regarded as
719 evergreen, but these are not specified.

720 Previous multivariate physiognomic analysis indicated a MAT of 10 °C
721 with a CMMT between 0 and 6 °C and high rainfall between 1500 and 1700
722 mm/yr (Golovneva, 2000). The palaeolatitude of the Kakanaut Flora is of

723 particular importance because the light regime is critical to arguments of
724 dinosaur over-wintering behaviour. Godefroit et al. (2009) cite a position of
725 between 70 and 75 °N based on Howarth (1981) and Smith et al. (1981) and
726 this is supported by the reconstruction shown in Fig. 1. The Smith et al. (1981)
727 reconstruction places the Alaskan Prince Creek localities at a much higher
728 position at around 85 °N, which is consistent with the considerably more
729 depauperate woody flora lacking any discernable evergreen components. In
730 contrast Figure 1 positions the Prince Creek only a few degrees higher than
731 the Kakanaut at about 78 °N. The palaeolatitudinal difference in data supplied
732 by Chris Scotese that form the basis of the latitudinal thermal gradients
733 reported later in this work place the Alaskan Prince Creek locality at 82 °N
734 and the Kakanaut at 76 °N.

735

736 **4. Floral Interchange and Differentiation in the Cretaceous Arctic**

737

738 [Figure 7 near here]

739

740 The relationships between the regional paleofloras of North Alaska and
741 N.E. Russia are summarised in Figure 7. The similarities, even at the species
742 level, between the Early Ginter and the Kukpowruk, and the Grebenka and
743 Niakogon regional paleofloras indicate a high degree of genetic connection
744 and similarities in climatic regime between the two regions. By the Coniacian
745 (and possibly as early as the late Turonian) a strong differentiation had arisen
746 between the Kaivayam and Tuluvak regional paleofloras. This suggests
747 limited genetic connection and/or climatic differentiation. This persists

748 throughout the Coniacian and on to the Maastrichtian. The Koryak floral type
749 spans the end Cretaceous transition to the Paleocene and this continuance is
750 indicative of a notable lack of extinctions at these high palaeolatitudes (Spicer
751 et al. 1994; Herman, 2007c). After the K/T transition warming at high Arctic
752 latitudes led to an early Paleocene poleward migration of many taxa such that
753 the early Sagwon Flora of North Alaska (Selandian-Thonetian) (Spicer et al.,
754 2009b) shows strong similarities with the Maastrichtian Koryak Flora
755 (Moiseeva et al., 2009).

756

757 [Figure 8 near here]

758

759 **5. Lower Latitude Floras**

760

761 As well as the Arctic floras considered here, a small selection of lower
762 latitude floras from the Tethyan margin and N. America have been analysed
763 to derive latitudinal temperature gradients and thus to suggest near polar
764 climatic regimes where plant diversity is otherwise too low for reliable “in situ”
765 analyses.

766

767 *5.1 North America*

768

769 The late Jack A. Wolfe scored several Late Cretaceous floras from
770 North America but as yet they are unpublished. His scores for the Tuscaloosa,
771 Woodbine, Lance, Ripley, Gayhead, Cooper and Edmonton floras (Fig. 1)

772 were used for the analyses reported here. The ages and palaeolatitudes used
773 to derive the latitudinal temperature gradients are given in Figure 9.

774

775 *5.2 Europe*

776

777 The positions of these sites are given in Figure 1. Leaves from two
778 localities within the Peruc Flora, Czech Republic (Vysehorovice and Praha
779 Mala Chuchle), represent florules of the same Cenomanian age, but different
780 environments (Herman et al., 2002). Plant fossils from the Praha Mala Chuchle
781 locality come from the fluvial sediments deposited in a Cenomanian river
782 channel bordered by slopes of Palaeozoic limestone bedrock, whereas those
783 from the Vysehorovice locality come from fluvial sediments of a larger river
784 meandering across an extensive lowland floodplain. CLAMP analyses of these
785 florules are presented separately and in combination (Peruc). Also from the
786 Czech Republic is Senonian flora of Zliv, Klikov and Hluboka (Nemejc, 1962;
787 Nemejc and Kvaek, 1975; Herman et al., 2002; Váchová and Kvaek, 2009),
788 which are here combined and labelled “Czechsen” in Table 1.

789 The Campanian Grünbach Flora comes from the Grünbach Formation
790 of the Gosau Group in the Grünbach-Neue Welt Basin in the Eastern Alps,
791 Austria (Herman and Kvaek, 2007a, b, 2009). The probable early Campanian
792 age of this flora is based on the correlation of plant-bearing deposits with
793 marine biostratigraphy and on palynological data (Summesberger, 1997). The
794 palaeogeographic situation is that of a large island with unknown relief on the
795 northern margins of the Tethys Ocean. Terrestrial freshwater swamps and
796 shallow water sediments indicate a relatively large deltaic plain. Lithological

797 climate indicators, as well the presence of large foraminifers, rudists, corals
798 and a reptile fauna (carnosaurs, iguanodons, pterosaurs, scelidosaur and
799 crocodiles) throughout the Santonian-Maastrichtian of the Grünbach Basin,
800 suggest a 'subtropical' climate considerably warmer than that experienced in
801 the Arctic. As well as a diverse fern, conifer and dicot flora the plant
802 assemblage contains early palms and *Pandanus* (Herman and Kvaček, 2009).
803 An end Cretaceous palaeolatitude of 32 °N for the Gosau Basin has been
804 reported (Preisinger et al., 1986).

805

806 **6. Quantifying Non-marine Circum-Arctic Palaeoclimates**

807

808 Earlier palaeoclimate determinations have been a combination of
809 qualitative and quantitative analyses. Qualitative interpretations have included
810 those of tree rings (Parrish and Spicer, 1988b; Spicer and Parrish, 1990b),
811 vegetational composition (Spicer and Parrish, 1986), and interpretations of
812 shoot and leaf architecture coupled with taphonomic observations (Spicer and
813 Herman, 1996) and floral diversity (Smiley, 1967). These have constrained
814 the light regime and seasonality as well as indications of evaporation /
815 precipitation ratios, overall precipitation, and, to some extent, air
816 temperatures. Quantitative estimates of mean annual temperatures have
817 been derived from leaf margin analysis (Parrish and Spicer, 1988a; Spicer
818 and Parrish, 1990a) but this technique can offer no indication of seasonal
819 variations in temperature, which is a critical parameter at high latitudes. More
820 latterly, multivariate analysis of foliar physiognomy, specifically Climate Leaf
821 Analysis Multivariate Program (CLAMP) (Wolfe, 1993; Craggs, 2005;

822 Golovneva, 2000; Herman and Spicer, 1996b; Herman and Spicer, 1997;
823 Spicer et al., 2002), has been applied because it can offer estimates of both
824 the warm month mean and cold month mean temperatures as well as
825 seasonal variation in rainfall.

826

827 *6.1 CLAMP Analysis*

828

829 Climate Leaf Analysis Multivariate Program (CLAMP) (Wolfe, 1993)
830 (<http://www.open.ac.uk/earth-research/spicer/CLAMP/Clampset1.html>) is a
831 multivariate palaeoclimate proxy based on the observation that leaf
832 architecture is highly responsive to, and reflective of, climate. Leaf
833 architecture (physiognomy) is the result of adaptation within the context of the
834 capabilities of the genome honed by natural selection, and the evolutionary
835 driver is increasing fitness gained by the maximization of photosynthetic
836 capability in the energetically most efficient way. Non-adapted physiognomies
837 do not compete effectively, and over time there is a degree of convergence of
838 form largely independent of taxonomy (Spicer, 2000, 2007, 2008). No single
839 architectural feature of a leaf, or whole plant, determines adaptive success
840 (Lande and Arnold, 1983), and thus no single feature can be expected to
841 correlate with a single climatic variable. It is becoming increasingly clear that
842 numerous interacting traits influence fitness (Ackerly et al., 2000) and
843 architectures of conducting tissues in roots, stems and leaves, overall canopy
844 architecture (Hellicker and Richter, 2008) and foliar physiognomy are all
845 important in producing a photosynthetically efficient structure. However,

846 because plants rarely fossilise as complete entities, analysis of dispersed
847 organs such as isolated woody stems and leaves is all that is possible.

848 CLAMP is a non-taxonomic statistical method embedded in well
849 defined calibration and numerically descriptive protocols that attempt an
850 objective decoding of foliar physiognomy in terms of environmental
851 constraints, particularly climate. It has a particular advantage over the so-
852 called Nearest Living Relative methods such as the Co-existence Approach
853 (Mosbrugger and Utescher, 1997) for the Late Cretaceous in that it is
854 grounded in the time-stable laws of the physics of fluid conduction, gas
855 diffusion and radiation balances and therefore can be used in 'deep time'
856 studies where taxon-based evolutionary change is likely. A significant
857 advantage of CLAMP, and most palaeobotanical proxies for that matter, is
858 that unlike isotopic techniques it is not subject to uncertainties arising from
859 diagenetic alteration.

860 Although non-angiosperms do code for climate in broad qualitative
861 terms (Rees et al., 2000; Spicer et al., 1994), to date quantification is
862 restricted to the physiognomy of woody dicot leaves and can only be applied
863 reliably to floral assemblages where there is a diversity (>20 morphotypes) of
864 such leaves. However such floras provide climatic constraints and insights to
865 Arctic conditions where angiosperms, sufficient in diversity to provide reliable
866 climate estimates, are lacking.

867 CLAMP is a dynamic methodology that has been periodically improved
868 by, for example, the addition of new modern vegetation calibration (training)
869 sites and the use of different sets of meteorological observations for
870 calibration purposes. Wolfe's (1993) original meteorological data were

871 gleaned from observations made in, or close to, the vegetation sites used to
872 calibrate the technique and define the structure of multidimensional
873 “physiognomic space”. As such, the observations were to some extent
874 influenced by the presence of the vegetation itself and therefore reflective of
875 the local microclimate and less so the regional climate. Prior to large-scale
876 human destruction of natural vegetation this would not have been a problem,
877 but today climate stations used to amass global climate data are usually
878 located in landscapes highly modified by agriculture and/or urbanisation.
879 These data, on average warmer than those obtained in sub-canopy
880 environments, are those used to evaluate and constrain climate models, and
881 for data/model comparisons for the geological past it is desirable that proxies
882 such as CLAMP use the same meteorological calibration. This is particularly
883 important because there has been a persistent mismatch between
884 palaeoclimate proxies and “warm Earth” model simulations at high latitudes
885 and continental interiors dating back over 25 years (Spicer et al., 2008;
886 Valdes, 2000). Moreover while Wolfe’s calibration data were usually
887 summaries of more than 30 years of observations, the 30 or more years over
888 which the data were collected were not the same.

889 To achieve a CLAMP calibration more akin to that used in climate
890 modelling we utilized the 0.5° x 0.5° global climate data set of (New et al.,
891 1999) supplemented by the ERA40 re-analysis data for atmospheric
892 temperature at upper levels. The New et al. (1999) data are derived from the
893 World Meteorological Organisation (WMO) 1961-1990 global observed
894 standard normals, supplemented by other data from national meteorological
895 agencies collated by those authors. More detailed accounts of data sources,

896 standardizations where more than one definition of a climate variable was
897 used, and uncertainties are given in New et al. (1999). To correct for the effect
898 of altitude, a geographically variable lapse rate was calculated by using the 3-
899 D climatology of temperature from the ECMWF re-analysis project (Uppala et
900 al., 2005). From this the mean lower tropospheric lapse rate for each month of
901 the year was calculated. These data are on a 96 x 73 grid that was then
902 interpolated onto a 720 x 360 grid (0.5° x 0.5°) using a bilinear interpolation
903 scheme that conserves area averages. This interpolation scheme inevitably
904 smoothes altitudes and there is no recognition of aspect.

905 The gridded data were then corrected to the observed altitude of the
906 exact latitude and longitude of the CLAMP calibration site using the monthly
907 lapse rates and a bilinear interpolation scheme. The specific and relative
908 humidities were also recalculated in light of the lapse rates. If the lapse rate
909 corrected temperature was much colder than the original, then potentially the
910 relative humidity could have been greater than 100%. If this was the case, the
911 specific humidity was adjusted until the relative humidity was equal to 100%.
912 From this the commonly returned CLAMP climate variables (mean annual
913 temperature - MAT, warm month mean temperature - WMMT, cold month
914 mean temperature - CMMT, length of the growing season - LGS, mean
915 growing season precipitation - GSP, mean monthly growing season
916 precipitation - MMGSP, precipitation during the three wettest months - 3-WET,
917 precipitation during the three driest months - 3-DRY, specific humidity - SH,
918 relative humidity - RH, and enthalpy) were calculated to provide a
919 meteorological calibration data set (GRIDMET3BR). This was used with the

920 PHYSG3BRC set of foliar physiognomic scores to provide a calibration
921 framework for the fossil sites. Both are provided on the CLAMP website.

922 While transferring Wolfe's original hard copy raw scores of the training
923 sets to the electronic score sheet now routinely used in CLAMP analyses,
924 some minor arithmetic errors were detected. While these are so small that
925 they do not materially affect previously published results, they are now
926 corrected and form the PHYSG3BRC data set used here. A more
927 comprehensive analysis of the new calibration scheme as compared to those
928 used previously is given in Spicer et al. (2009a). Using the new gridded
929 calibration, analyses of Arctic floras are here presented together with selected
930 lower latitude Cretaceous sites.

931

932 **7. Results**

933 The results of the new CLAMP analyses using the gridded
934 meteorological data (GRIDMET3BR) are shown in Table 1.

935

936 **[Table 1 near here]**

937

938 In wet regimes it has long been known that leaf physiognomy is poorly
939 constrained by water availability compared to dry regimes (Wolfe, 1993). At
940 Late Cretaceous high northern latitudes there is ample evidence in the
941 terrestrial realm of an excess of precipitation over evaporation, thus estimates
942 of precipitation must be regarded with great caution, and in any case will be
943 biased towards the growing season. Growing season lengths are shorter for

944 the high palaeolatitude floras and all are between 5 and 8 months, consistent
945 with the temperature and inferred light regimes.

946 Because of the difficulties of comparing sites of different ages and at
947 different palaeolatitudes across all meteorological parameters, data from
948 GRIDMET3BR CLAMP analyses of 9 additional sites in N. America (Lance,
949 Gayhead, Tuscaloosa, Woodbine, Ripley, Edmonton, and Cooper) and one
950 from Kazakhstan (Teretki-Sai) were added to the results reported here to
951 derive latitudinal MAT gradients. These are shown in Figure 9. WMMT and
952 CMMT gradients were similarly derived, and the R^2 values and standard
953 deviations of the residuals as measures of uncertainty are presented in the
954 legend to Figure 9.

955

956 [Figures 9 near here]

957

958 Finally, using these gradients it was possible to calculate temperature
959 parameters at any palaeolatitude using the equations of the regressions in
960 Fig. 9. Figure 10 shows the MAT values so derived for 70 and 80 °N
961 superimposed on plots of observed MATs at different times in the Late
962 Cretaceous as derived by different palaeobotanical methods.

963

964 [Figure 10 near here]

965

966 **8. Discussion**

967

968 CLAMP has already been shown to have parity with open water coastal
969 SSTs derived from oxygen isotope analyses from the Cenomanian of New
970 Zealand (Kennedy et al., 2002), and, more indirectly through palaeoelevation
971 estimates, lacustrine bulk carbonate oxygen isotope values (Currie et al.,
972 2005; Spicer et al., 2003). These, and experimental observations (Gregory,
973 1996), suggest CLAMP not only shows reasonable agreement with oxygen
974 isotope proxies, but also that it appears immune (at least within known
975 uncertainties) to calibration errors associated with changing atmospheric CO₂
976 concentrations.

977 Alaskan North Slope Nanushuk Formation palaeosol $\delta^{18}\text{O}$ siderite
978 analyses used the LMA-based temperature estimates of Parrish and Spicer
979 (1988a) for calibration purposes to obtain quantitative estimates of
980 precipitation (Ufnar et al., 2004). The results (485 mm/yr) are consistent with
981 the widespread early Late Cretaceous peat formation (Lottes and Ziegler,
982 1994) and thereby indirectly confirm the palaeobotanical temperature
983 estimates. These early LMA-derived estimates are also consistent with those
984 obtained from CLAMP (although note that CLAMP analyses have yet to be
985 run for Nanushuk Formation floras). CLAMP does, however, yield higher
986 precipitation values than those of Ufnar et al. (2004), particularly bearing in
987 mind that CLAMP estimates growing season precipitation and not mean
988 annual precipitation. These wetter predictions are unsurprising given the high
989 uncertainties associated with foliar physiognomic techniques in wet regimes,
990 and the fact that leaf preservation is biased towards wet depositional
991 environments where the plants would not experience drought-induced water
992 stress.

993 Overall the new CLAMP values based on the gridded meteorological
994 data calibration are not significantly different from those previously published
995 using Wolfe's (1993) (together with later revisions) calibration data (e.g.
996 Herman and Spicer, 1996b). These, and the earlier CLAMP MAT estimates,
997 are consistent with those obtained by LMA, and show broad agreement with
998 other proxies (Price and Grimes, 2007; Spicer and Corfield, 1992).

999

1000 *8.1 MAT Estimates*

1001

1002 Including uncertainties, CLAMP Arctic MAT estimates reported here,
1003 and those of Tomsich et al. (this volume), range from around 8 °C to around
1004 15 °C with the coolest conditions (Kamchatka Turonian and Denali
1005 Maastrichtian) away from the Arctic Ocean. This is demonstrated most clearly
1006 in Fig. 10 where the N.E Russia floras close to the north Pacific are notably
1007 cooler than those at similar or higher palaeolatitudes bordering the Arctic
1008 Ocean. This is consistent with the post-Cenomanian floral differentiation
1009 between the two regions shown in Fig. 7 and supports the idea of a relatively
1010 cold northern Pacific Ocean gyre (Herman and Spicer, 1996b), compared to a
1011 warmer Arctic Ocean supplied by heat transported northwards along the
1012 Western Interior Seaway. Such a phenomenon has recently been seen
1013 developing in Late Cretaceous coupled atmosphere/ocean GCMs (P.J.
1014 Valdes, personal communication, 2008) and may help explain spatial and
1015 temporal heterogeneities in temperature estimates from the circum north
1016 Pacific region and elsewhere (Zaharov et al., 1999). Small changes in the
1017 position, temperature and intensity of such a gyre would result in significant

1018 differences in temperature proxy estimates depending on their location and
1019 accuracy of dating.

1020 Figure 10 also shows good agreement between the Maastrichtian MAT
1021 at 80 °N derived using the depauparate floral composition (Spicer and Parrish,
1022 1990) and those projected for the latitude using the CLAMP-derived
1023 palaeolatitudinal temperate gradient shown in Fig. 9. The floral estimate is 5
1024 °C as against 6.7 ± 2.2 °C for CLAMP. Note that this uncertainty is obtained
1025 from the CLAMP statistical uncertainty (Table 1) and does not incorporate the
1026 scatter around the MAT/palaeolatitude regression shown in Fig. 9. When the
1027 standard deviation of the residuals is combined with that due to CLAMP
1028 methodology the uncertainty remains more or less the same (± 2.23 °C) due
1029 to the good fit of the data to the regression.

1030 At 82 °N (the predicted latitude of the Ocean Point locality using the
1031 data of Scotese) the MAT is estimated to have been 6.3 °C. This value, and
1032 the CMMT, is likely to be cooler than that actually experienced because the
1033 slope of the line is influenced by the inclusion of the Cantwell (Denali) data,
1034 which is not only derived from the ungridded CLAMP calibration, but is also
1035 from a site well away from the Arctic Ocean coast.

1036 Equally intriguing as the cool northern Pacific gyre is the temperature
1037 of the Late Cretaceous Arctic Ocean. With the possible exception of seasonal
1038 winter ice cover in the Maastrichtian, most authors agree that the Late
1039 Cretaceous Arctic Ocean was ice-free year round. This is despite the lowered
1040 surface salinity arising from high precipitation in the circum-Arctic region, and
1041 restricted connections with the rest of the global oceanic system. Clear
1042 signatures of ice (as opposed to tree root) rafting of drop-stones are absent.

1043 Some authors (Jenkyns et al., 2004) have suggested that the average Arctic
1044 Ocean surface temperature at 70 Ma was 15 ± 1 °C and was above 20 °C in
1045 the Cenomanian. These temperatures were arrived at by using the TEX₈₆
1046 proxy based on the composition of membrane lipids derived from marine
1047 plankton belonging to the Crenarchaeota. The material used was obtained
1048 from drop-stone-free varved Maastrichtian organic-rich sediments recovered
1049 by shallow piston core from the Alpha Ridge. The Cenomanian temperature
1050 was arrived at using the Maastrichtian TEX₈₆ value and a proxy oxygen
1051 isotope curve from the English chalk together with assumptions about a
1052 shallow Cretaceous marine equator-to-pole temperature gradient.

1053 To help support their conclusions Jenkyns et al. (2004) cited an 1890
1054 reference to a putative Cenomanian *Artocarpus* (breadfruit) from Greenland
1055 (Nathorst, 1890). It is worth noting that in the 19th and early 20th centuries it
1056 was common practice to name leaves by reference to morphologically similar
1057 extant forms. As our understanding of early angiosperms improved this
1058 practice was abandoned and names distinct from modern genera were used
1059 instead. To assume the Greenland leaf represented a thermophyllic tropical
1060 plant, and to cite the temperature range of 15 - 38 °C under which modern
1061 breadfruits grow as evidence for a tropical Arctic regime, is to ignore evolution
1062 and the overall temperate character of the hundreds of taxa that comprise the
1063 circum Arctic Cretaceous floras. The highly seasonal growth regimes at high
1064 latitudes is likely to bias the TEX₈₆ proxy towards lipids produced in the
1065 warmer summer season and thus the temperatures obtained will reflect the
1066 WMMT rather than the MAT. The Jenkyns et al. (2004) estimates must

1067 therefore be regarded as overly warm in the absence of additional supporting
1068 information.

1069 If the Arctic Ocean was warm, and it would be imprudent to take the
1070 TEX₈₆ values at face value, how warm was it? Early attempts to use oxygen
1071 isotopes to estimate Cretaceous Arctic Ocean SSTs suffered from both an
1072 imperfect understanding of the effects of diagenesis and, more importantly,
1073 problems associated with reduced salinity.

1074 To avoid salinity uncertainties an alternative approach to estimating
1075 polar temperatures is to examine terrestrial equator-to-pole temperature
1076 gradients. Moreover, to test the palaeobotanical proxies this gradient should
1077 be derived independently of plant fossils. Using oxygen isotope ratios in
1078 vertebrate phosphate, including dinosaur bone remains from Alaska, Amniot
1079 et al. (2004) constructed a Campanian-Maastrichtian latitudinal temperature
1080 gradient that showed good agreement with quantitative palaeobotanical data
1081 across almost all the latitudinal range down to 30° (Wolfe and Upchurch,
1082 1987; Parrish and Spicer, 1988a), although the Alaskan isotopic MAT was
1083 reported as −5 °C. This is colder than palaeobotanical estimates, but within
1084 the thermal tolerance of the inferred Taiga-like palaeovegetation. Given the
1085 lack of sedimentological evidence for freezing this figure of −5 °C may be too
1086 cold, but not by much, and it is within the 2σ uncertainty range about the −2.3°
1087 CMMT CLAMP values obtained by Tomsich et al. (this volume). Taking into
1088 account uncertainties in palaeolatitudes, together with those inherent in the
1089 methodologies, the plant and isotope proxy data are in good agreement.

1090 For the highly seasonal Arctic, with prolonged periods of darkness,
1091 MATs are only part of the story. CLAMP-derived CMMTs attest to a general

1092 lack of freezing conditions and even the coldest CMMT estimates from the
1093 Cenomanian and Turonian of the Russian Far East, on the borders of the
1094 OCVB, suggest only brief periods (several weeks at most) of mild frosts. Even
1095 in the more continental interior position of the Vilui Basin away from oceanic
1096 influences, but above the palaeo-Arctic Circle, there appears to have been a
1097 lack of freezing during the winters (Spicer et al., 2008). Temperatures
1098 bordering the Arctic Ocean (8 °C for the Alaskan Coniacian and 1 °C for
1099 Novaya Sibir' Island, Table 1) are consistent with an Arctic Ocean surface
1100 temperature well above freezing, and possibly as high as 6 - 10 °C in the
1101 Coniacian. Such temperatures are required to maintain warm air temperatures
1102 during the prolonged winter darkness.

1103

1104 *8.2 CMMT Estimates*

1105

1106 A major advantage of CLAMP is that it not only yields MATs but also
1107 provides data on mean annual range of temperature. It is reasonable to ask,
1108 however, how reliable CLAMP is in estimating the CMMT when plants are
1109 dormant during that part of the year. This ability of CLAMP to provide
1110 reasonable CMMT estimates despite deep dormancy and deciduousness has
1111 been investigated using modern samples from the Siberian interior (Spicer et
1112 al., 2004). Although CLAMP tended to provide CMMT estimates that were
1113 somewhat too warm, indications of extreme cold were evident in that CLAMP
1114 analysis of modern leaves suggested a CMMT of around -30 °C in locations
1115 where the observed was closer to -40 °C. The explanation for CLAMP's ability
1116 to estimate CMMT could be that at the critical phase of leaf expansion in the

1117 spring leaf architecture has to be functionally efficient during the steep
1118 warming trajectory from winter to summer temperatures ($>35\text{ }^{\circ}\text{C}$). The CMMT
1119 physiognomic signal is therefore a reflection of rapid spring change. In the
1120 Cretaceous there would have been an equally, if not more, rapid foliar
1121 development phase driven by a fast-changing light regime and some
1122 associated warming despite the relatively warm winters.

1123 CLAMP climate estimates are consistent with the overall composition
1124 of the Arctic vegetation. Vegetation composition ranged from being
1125 characteristic of warm temperate to cool temperate throughout the Late
1126 Cretaceous. Conditions cold enough to lead to the development of vegetation
1127 similar to that of modern Taiga appear to have existed in the Maastrichtian in
1128 northern Alaska, but even then winter CMMT temperatures may not have
1129 been significantly below freezing. Unfortunately the flora representing the
1130 coolest conditions contained insufficient woody dicot taxa for a CLAMP
1131 analysis, but insights can be gleaned from the gradients obtained lower
1132 latitude co-eval North American floras. Using this approach the estimated
1133 CMMT at $82\text{ }^{\circ}\text{N}$ was $-1.98 \pm 3.9\text{ }^{\circ}\text{C}$ (2σ combined uncertainty). It is likely then
1134 that in the northern Alaska Maastrichtian winter episodes of temperatures $< -$
1135 $10\text{ }^{\circ}\text{C}$ probably did occur, but were rare and unlikely to have lasted more than
1136 a week or so. Even so winter cold, combined with prolonged darkness and a
1137 lack of foliage for food, would have posed significant problems for the resident
1138 fauna and those that, due to their size, could not have found shelter for
1139 hibernation or aestivation are likely to have migrated south for over-wintering.

1140 In the earlier (warmer) parts of the Late Cretaceous the Cenomanian
1141 CMMT latitudinal gradient was $1.8\text{ }^{\circ}\text{C}/10^{\circ}$ yielding a CMMT at $82\text{ }^{\circ}\text{N}$ of $+3.8 \pm$

1142 4.1 °C (2σ combined) (R^2 0.84, 2σ of residuals 1.96 °C), while in the
1143 Turonian-Coniacian the gradient is predicted to have been 1.4°C/10° with a
1144 CMMT at 82 °N of $+0.2 \pm 5.3$ °C (2σ combined) (R^2 0.47).

1145 By contrast the Cenomanian of the Czech Republic at a palaeolatitude
1146 just above 30 °N show CMMTs between +7.2 and +10.4 °C (2σ uncertainty \pm
1147 3.8 °C; Table 1). Later in the Cretaceous, consistent with the well known Late
1148 Cretaceous global cooling trend, CMMTs decline (Grünbach +3.2 °C and
1149 Czechsen +0.9°C).

1150

1151 *8.3 WMMT Estimates*

1152

1153 Late Cretaceous Arctic WMMTs suggest cool summers compared to
1154 conditions in Tethyan margin sites (26 - 29 °C). The shallow WMMT latitudinal
1155 gradients of 0.9 °C /10° latitude in the Cenomanian (R^2 0.69, 2σ residuals
1156 1.92 °C), 2.1 °C/10 °C in the Turonian/Coniacian (R^2 0.94, 2σ residuals 1.54
1157 °C) and 2.3 °C /10° latitude in the Maastrichtian (R^2 0.96, 2σ residuals 1.2 °C),
1158 together with those of the CMMT, are consistent with a year-round warmth
1159 and a persistent weak polar high pressure cell. At 82 °N the Maastrichtian
1160 WMMT is projected to have been 14.5 °C with a combined uncertainty (2σ) of
1161 ± 3.1 °C. At 82 °N in the Cenomanian the WMMT is projected to have been
1162 20.5 ± 3.3 °C (2σ combined), while in Turonian-Coniacian it was 17.3 ± 4.8
1163 °C. The difference of 6 °C between Cenomanian WMMT at 82 °N and the
1164 Maastrichtian WMMT at the same palaeolatitude is only marginally greater
1165 than the differences in the CMMTs (5.8 °C) and is certainly not significant
1166 when the uncertainties are taken into consideration. This suggests that

1167 warming and cooling at high latitudes is not just a function of the summer
1168 insolation, but reflects changes to the entire annual thermal cycle.

1169

1170 *8.4 Humidity and a Permanent Polar Cloud Cap*

1171

1172 CLAMP estimates of Arctic annual average relative humidities are all
1173 high, being between 70 and 85%. A persistent polar cloud cap has been
1174 proposed by Herman (1994), who argued that the large leaf sizes found at
1175 high latitudes during the Cretaceous might be a response to diffuse light.
1176 Such a cloud cap would also be a product of an enhanced hydrological cycle
1177 developed under a weak polar high-pressure cell. Recently Mercado et al.
1178 (2009) have shown that the global land carbon sink was enhanced by 23%
1179 between 1960 and 1999 due to increased diffuse solar radiation resulting from
1180 volcanic sulphur aerosols and “global dimming”. This increase in terrestrial
1181 photosynthetic activity resulting from diffuse light would also have operated at
1182 high latitudes under the Late Cretaceous polar cloud cap and particularly
1183 where extensive and frequent volcanism in the OCVB would have contributed
1184 aerosols (see also Herman and Spicer, this volume). Coupled with higher
1185 atmospheric CO₂ concentrations and the possible enhancement of evergreen
1186 over-wintering strategies under warm dark winters (Royer et al., 2003), this
1187 would help explain the high N. polar primary productivity in Late Cretaceous
1188 times.

1189 Figure 11 shows a reconstruction of a summer scene under such a
1190 cloud cap at 70 Ma on the margins of the Arctic Ocean at approximately 82
1191 °N. Such a year round cloud cap has been generated in some Maastrichtian

1192 coupled ocean/atmosphere modelling experiments using a 3 x pre-industrial
1193 atmospheric CO₂ concentration, and a reduced solar constant (run XATTS,
1194 unpublished data of P.J. Valdes). In the experiments the CMMT Arctic Ocean
1195 margin temperatures over land matched those suggested by the vegetation
1196 while the 2 m air temperatures over the ocean were 15 °C. This not only
1197 suggests the possibility of very steep thermal gradients from ocean to land,
1198 but also that the TEX₈₆ estimates should be regarded as reasonable,
1199 despite a likely bias towards the WMMT, pending the emergence of evidence
1200 to the contrary.

1201

1202 [\[Figure 11 near here\]](#)

1203

1204 *8.5 The Effect of Volcanism*

1205

1206 The provenance of the widespread Late Cretaceous bentonites in
1207 northern Alaska has long been a contentious issue (e.g., Smiley, 1969a;
1208 Bergman et al., 1995) because of the dearth of proximal sources. The first
1209 thin Cretaceous bentonites occur in the Albian of the western North Slope.
1210 Thick ashes occur later in the Cretaceous, notably in the Cenomanian, and
1211 are found further east (Brosgé and Whittington, 1966), although a significant
1212 part of Late Cretaceous time is not represented by sediments in the west.
1213 ⁴⁰Ar/³⁹Ar ages on biotite and feldspar from tuffs in northern Alaska range from
1214 100 Ma to 67 Ma, but most are 93 - 85 Ma (Bergman et al., 1995); pers.
1215 comm. to RAS, 1997). The significance of the Kelley et al. (1999) age
1216 reassignment is that it constrains the period of maximum eruptive activity in

1217 the OCVB to coincide with the deposition of the thickest bentonites across
1218 northern Alaska and thus makes it likely that most, if not all, ash deposits in
1219 the Corwin and Umiat Deltas were sourced to the west in N.E. Russia
1220 (Bergman et al., 2006; Kelley et al., 1999) although a Canadian source cannot
1221 be discounted entirely (Bergman et al., 2006). The lack of any ash deposits to
1222 the west of the OCVB constrains predominant Late Cretaceous Arctic wind
1223 directions as being from N.E. Russia eastwards to Alaska. This has
1224 implications for atmospheric heat transport from west to east, and
1225 determinations of air temperatures in N.E. Russia therefore have a bearing on
1226 those in Alaska. Convection over the warm Arctic Ocean coupled with the
1227 rotation of the Earth would have drawn cool air from over the North Pacific
1228 gyre north-eastwards across Alaska leading to offshore winds. This is
1229 reflected in the strong land to sea thermal gradients evident in the XATTS
1230 model run.

1231 The occurrence of ash deposits across the N. Pacific and Arctic Ocean
1232 margin areas raises the possibility that the vegetation, and hence the plant-
1233 derived climate signal, might be compromised by ash-related disturbance.
1234 However studies of modern day eruptions demonstrate that ash falls rarely
1235 cause significant vegetational disturbance in areas distal to volcanic vents
1236 (Spicer, 1989). The lack of vents proximal to northern Alaska suggests
1237 vegetation there was unlikely to have been adversely affected except for
1238 short-term smothering of ground cover. Ash falls undoubtedly contributed
1239 nutrients and stimulated growth in nutrient-poor communities such as on
1240 raised mires. The development of thick palaeosols between ash horizons
1241 suggests long intervals between significant fall events in any given region,

1242 thus the majority of leaf accumulations reflect vegetation minimally impacted
1243 by volcanism. Leaf assemblages used in palaeoclimatic analysis were in the
1244 most part from siliciclastic horizons with no, or very small, volcanogenic
1245 components. The only exception to this was the Grebenka flora at the
1246 Yelisseev locality where the volcanoclastics were all fluvially transported from
1247 the OCVB uplands and only one ash horizon was found within several
1248 hundred metres of section.

1249 In the OCVB frequent heavy ash falls and deep accumulations of
1250 volcanoclastics no doubt did affect vegetation composition such as in the case
1251 of the Chauna flora (Kelley et al., 1999). Such floras were not used for
1252 palaeoclimatic analysis.

1253

1254 **9. Conclusions**

1255

1256 Despite the Late Cretaceous evolutionary novelty of woody dicots and
1257 the lack of any modern analogues for the Late Cretaceous Arctic forests
1258 growing under a polar light regime, new CLAMP palaeoclimate determinations
1259 using a global gridded climate calibration yield values that are similar to
1260 previous estimates using CLAMP, leaf margin analysis, oxygen isotopes
1261 based on terrestrial vertebrate phosphates, marine carbonates from the North
1262 Pacific and palaeosol siderites. CLAMP values are also consistent with
1263 qualitative interpretations arising from plant diversity and vegetation
1264 composition, tree rings and clay minerals. This high level of agreement
1265 between diverse proxies suggests that CLAMP estimates, although bound to
1266 be refined further in the future, are as likely as any method to yield meaningful

1267 climate estimates. To the statistical uncertainties quoted in the CLAMP
1268 methodology have to be added uncertainties arising from palaeolatitudinal
1269 positioning, particularly as latitude is critical to understanding the insolation
1270 regime in near-polar positions.

1271 Palaeobotanical proxies indicate the maintenance of temperatures near
1272 or above freezing over land surrounding the Late Cretaceous Arctic Ocean
1273 despite long periods (possibly as long as 5 months) of continuous darkness
1274 each winter. This argues not only for a persistently warm Arctic Ocean
1275 throughout those months, but an effective thermal “blanket” to keep air
1276 temperatures around the ocean margins high. This is most likely to have been
1277 a more or less permanent polar cloud cap maintained by a strong hydrological
1278 cycle driven by evaporation from the warm ocean. The high precipitation and
1279 humidity values evident from the CLAMP analyses support this, as well as the
1280 precipitation estimates from the palaeosol siderites and the abundance of
1281 coals.

1282 Global cooling towards the end of the Cretaceous, manifest by a
1283 reduction in summer temperatures, was associated with a shallowing and
1284 eventual closure of the Western Interior Seaway and thus a reduction of heat
1285 transport to the Arctic. This may have led to a strengthening of the polar high
1286 and some weakening of the hydrological cycle. This in turn may have reduced
1287 the permanence and thickness of the polar cloud cap compared to earlier
1288 times, and led to further polar cooling and strengthening of the polar high.
1289 Nevertheless, given continued evidence for higher precipitation and
1290 temperatures than today, the cloud cap is likely to have been a feature of the
1291 Arctic throughout the Late Cretaceous and into the Paleocene. Such a cloud

1292 cap can be developed in coupled atmosphere/ocean general circulation
1293 models even for the relatively cool Maastrichtian. Ironically events at the end
1294 of the Cretaceous such as impact(s) and/or Deccan volcanism may have
1295 resulted in a fortuitous interruption of this cooling feedback loop by raising
1296 global CO₂ levels and invigorating the global hydrological cycle. The result
1297 being Arctic warming most notably expressed by increased plant diversity in
1298 the Paleocene palaeolatitudes as high as 85°N (Moiseeva et al., 2009; Spicer
1299 et al., 2009b). There is no evidence for major floristic turnover or extinctions at
1300 or near the end-Cretaceous boundary. Instead there were migrations into the
1301 Arctic as early Paleocene, and possibly late Maastrichtian, warming
1302 progressed.

1303 The data presented here further support our earlier suggestion
1304 (Herman and Spicer, 1996) that a relatively cool gyre developed in the
1305 northern Pacific Ocean during the Late Cretaceous. The effect of the resulting
1306 steep temperature gradient between the northern Pacific coastal plains and
1307 those of the warmer Arctic Ocean led to taxonomically distinct vegetation
1308 developing in these two areas. This differentiation was strongest in mid Late
1309 Cretaceous times and only disappeared after the end of the Cretaceous when
1310 warming led to migration and re-organisation of Arctic vegetation in the
1311 Paleocene.

1312

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1314

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1329

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1906

1907 Figure Legends

1908

1909 Figure 1. Palaeogeographic reconstructions of the North Polar Region at
1910 100Ma, 88 Ma and 70 Ma based on Hay et al. (1999) and obtained
1911 from the ODSN website (<http://www.odsn.de/>). Plant fossil location
1912 numbers used here and in Table 1 as follows: 1 Coniacian Tuluvak
1913 Formation, 2 Turonian Novaya Sibir', 3 Cenomanian-Turonian Yukon-
1914 Koyukuk Basin, 4 Cenomanian to Maastrichtian Vilui Basin, 5
1915 Cenomanian Grebenka Flora, 6 Turonian Kamchatka, 7 Coniacian
1916 Kamchatka, 8 Turonian-Coniacian Arman River, 9. Coniacian
1917 Tylpergergenai, 10 Cenomanian Chuchle, 11 Cenomanian
1918 Vyshehorovice, 12 Cenomanian Peruc, 13 Campanian Grünbach, 14.
1919 Senonian of the Czech Republic, 15 Albian-Cenomanian Nanushuk
1920 Formation, 16 Albian-Cenomanian Kukpowruk Flora, 17 Maastrichtian
1921 Prince Creek (Kogosukruk) Flora, 18 Maastrichtian Kakanaut Flora, 19
1922 Albian Buor-Kemyus (Zyrianka River Basin), 20 Albian Buor-Kemyus
1923 (Ainakhkurgen Depression), 21 Maastrichtian Lance, 22 Maastrichtian
1924 Medine Bow, 23 Maastrichtian Ripley, 24 Cenomanian Tuscaloosa, 25
1925 Cenomanian Woodbine, 26 Santonian Gayhead, 27 Maastrichtian
1926 Edmonton, 28 Maastrichtian Cooper Pit, 29 Cenomanian Tetetky-Sai.
1927

1928 Figure 2. Map of present day N.E. Russia and Alaska showing positions of
1929 areas yielding taphofloras. 1) Novaya Sibir Island; 2) Arman River; 3)
1930 Yelistratov Peninsula; 4) Cape Valizhgen; 5) Cape Konglomeratoyi; 6)
1931 Grebenka-Orlovka; 7) Chukotskaya-Bystraya River Basin; 8) Ubienka-

1932 Krestovaya River Basin; 9) Chineiveem River; 10) Rarytkin Ridge
1933 (southwestern part); 11) Pekul'nei Ridge western slope; 12) Pekul'nei
1934 Ridge eastern slope; 13) Anadyr River middle reaches; 14) Velikaya
1935 River lower reaches; 15) Khatyrka River; 16) Pekul'neiskoe Lake; 17)
1936 south Bering Peninsula; 18) Ugol'naya Bay; 19) Amaam Lagoon; 20)
1937 Corwin Bluffs area; 21) Utukok, Kokolik, Kukpowruk Rivers; 22) Kuk -
1938 Kaolak area; 23) Yukon-Koyukuk Basin; 24) Denali; 25) Colville (Umiat)
1939 - Chandler River region; 26) Sagavanirktok River area.

1940

1941 Figure 3. Chronostratigraphic column for the late Albian and Late Cretaceous
1942 of the Colville Basin, North Slope, Alaska, based on Mull et al. (2003).

1943

1944 Figure 4. Correlation chart for Albian to Paleocene deposits of the Arctic
1945 Slope of N. Alaska: (l) lower, (m) middle, (u) upper. LU – Lithological
1946 Unit of Smiley (1966). Modified from Herman (2007b).

1947

1948 Figure 5. Correlation chart for Albian and some Cenomanian floras of
1949 Northeastern Russia and Northern Alaska. Modified from Herman
1950 (2007c).

1951

1952 Figure 6. Correlation chart for upper Albian to early Paleocene deposits in
1953 N.E. Russia (the Anadyr-Koryak subregion): (l) lower, (m) middle, (u)
1954 upper. Modified from Herman (2007a).

1955

1956 Figure 7. Correlation chart of N.E. Russia and N. Alaska regional floras.
1957 Similar shades of grey indicate similar floral composition. Arrows
1958 indicate floral connections/migrations. Modified from Herman (2007c).
1959
1960 Figure 8. The Arctic light regime with present day obliquity showing the
1961 number of hours of daylight per 24 hour period throughout the year
1962 (Anonymous, 1978).
1963
1964 Figure 9. Graph showing the MAT palaeolatitudinal temperature gradients for
1965 floras grouped into age bins 98.6 – 95 Ma (Cenomanian), 90-85 Ma
1966 (Turonian – Coniacian) and 70 Ma (Maastrichtian). Internally consistent
1967 palaeolatitudes for these age intervals were obtained using
1968 PointTracker v.5.5 supplied by C. Scotese. The ages and
1969 palaeolatitudes used here are shown after the name of the flora.
1970 Teretky-Sai (Kazakhstan) (95 Ma: 42 °N); Vilui Basin (Siberia) (95 Ma:
1971 68 °N); Tuscaloosa (Alabama) (95 Ma: 32 °N); Woodbine (Texas) (95
1972 Ma: 32 °N); Chucle and Vyshehoroviche (Czech Republic) (95 Ma:
1973 40°N); Grebenka (98.6 Ma: 81°N); Kaysayam (Kamchataka) (88 Ma: 72
1974 °N); Penslina (Kamchataka) (90 Ma; 72 °N); Tuluvak (N. Slope) (88 Ma:
1975 80 °N); Arman River (N.E. Russia) (88 Ma: 68 °N); Tylpergergenai
1976 (N.E. Russia) (88 Ma: 79 °N); CzechSen (85 Ma: 39 °N); Lance
1977 (Wyoming) (70 Ma: 49 °N); Ripley (Mississippi) (70 Ma: 38 °N); Denali
1978 (Alaska) (70 Ma: 74 °N) (data from Tomsich et al. this volume); Cooper
1979 Pit (Alabama) (70 Ma: 36 °N); Edmonton (Canada) (70 Ma: 55 °N)
1980 (data from Tomsich et al. this volume). Regression equations -

1981 Cenomanian $y = -0.1421x + 23.348$, R^2 0.909, 2σ of residuals 1.08 °C;
1982 Turonian-Coniacian $y = -0.1674x + 22.434$, R^2 0.809, 2σ of residuals
1983 2.6°C. Maastrichtian $y = -0.1947x + 22.242$, R^2 0.9716, 2σ of residuals
1984 0.36 °C.

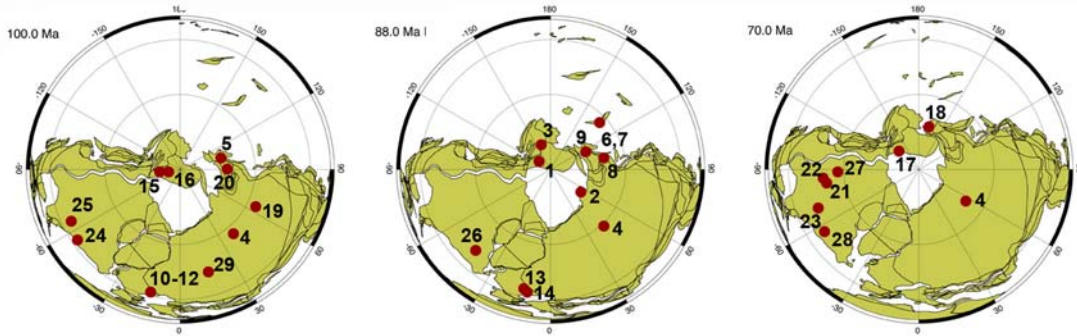
1985 Figure 10. Graph showing the MAT /Age relationship derived from floras in
1986 N.E. Russia and N. Alaska using floral composition, Leaf Margin
1987 Analysis and CLAMP. The Arctic Ocean surface temperature estimate
1988 (assumed to be MAT but possibly biased towards the WMMT) derived
1989 from the TEX_{86} proxy is also shown (Jenkyns et al., 2004).

1990

1991 Figure 11. Reconstruction of a summer scene at 70 Ma close to the Arctic
1992 Ocean shoreline (80 – 85 °N).

1993

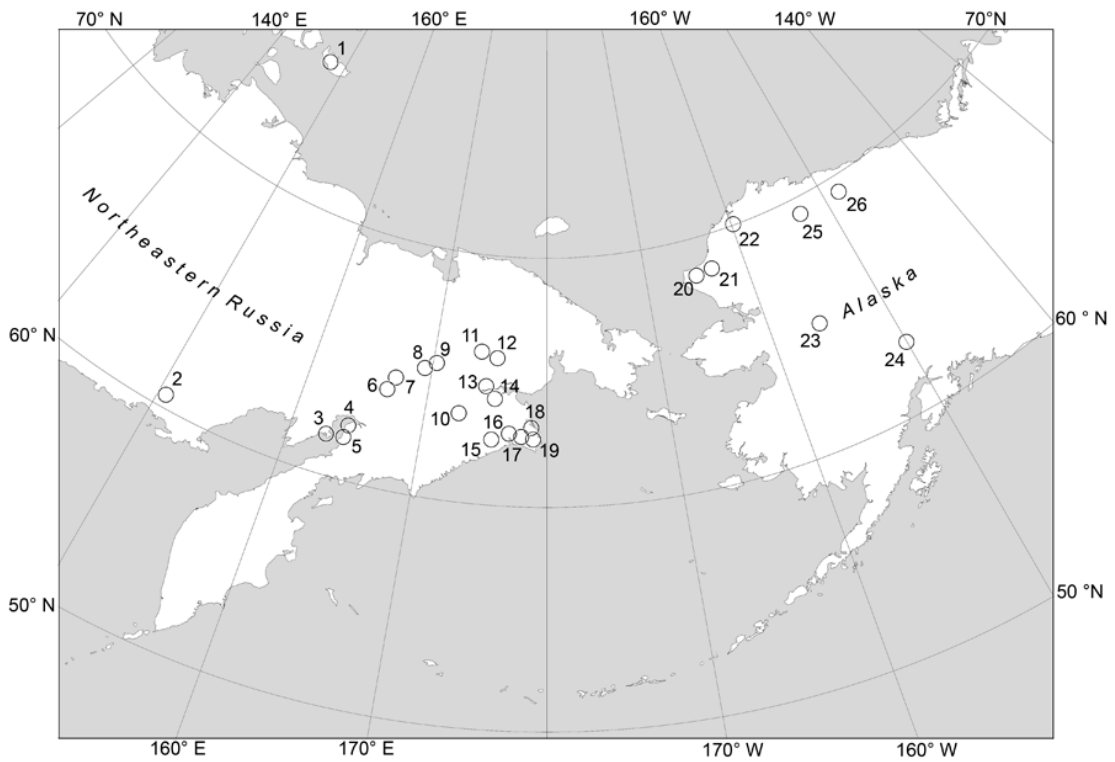
1994 Table 1. Palaeoclimate determinations for Late Cretaceous fossil floras using
1995 CLAMP calibrated with global gridded climate data derived from New et
1996 al. (1999) using the methodology of Spicer et al. (2009). The floras are
1997 numbered as in Fig. 1. Abbreviations: MAT, warm month mean
1998 temperature - WMMT, cold month mean temperature - CMMT, length
1999 of the growing season - LGS, mean growing season precipitation -
2000 GSP, mean monthly growing season precipitation - MMGSP,
2001 precipitation during the three wettest months - 3-WET, precipitation
2002 during the three driest months - 3-DRY, specific humidity - SH, relative
2003 humidity - RH, enthalpy – ENTHAL and Palaeolatitude - P Lat.
2004 Palaeolatitudes were obtained using Point Tracker supplied by C.
2005 Scotese.



2006

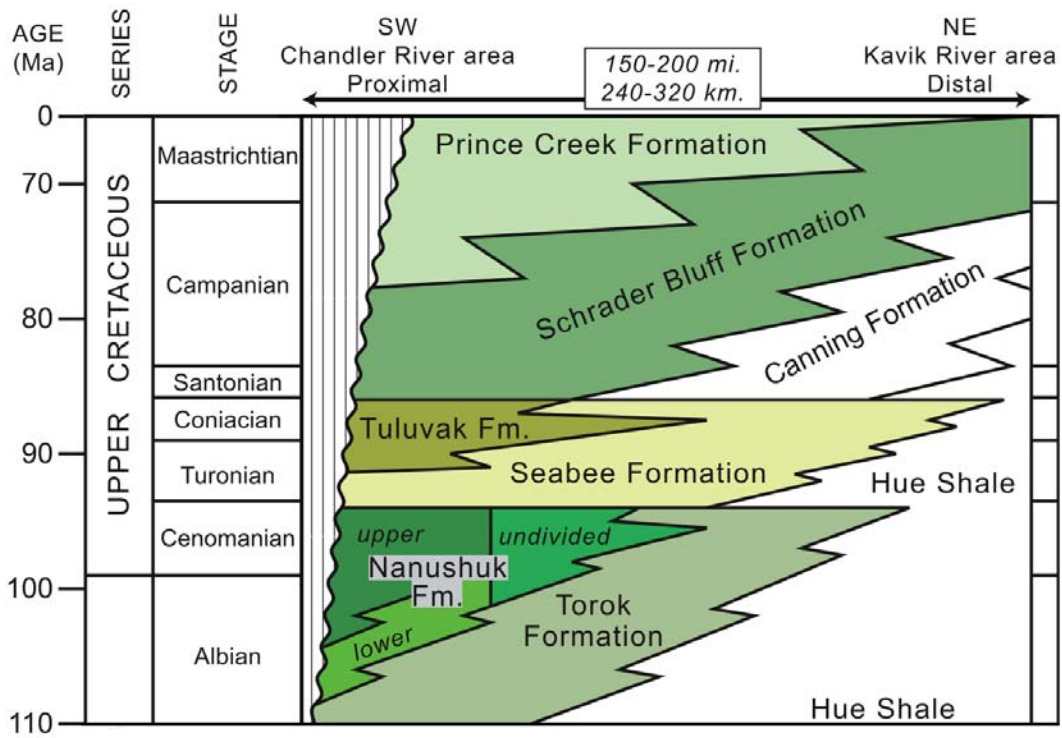
2007 Figure 1

2008



2009

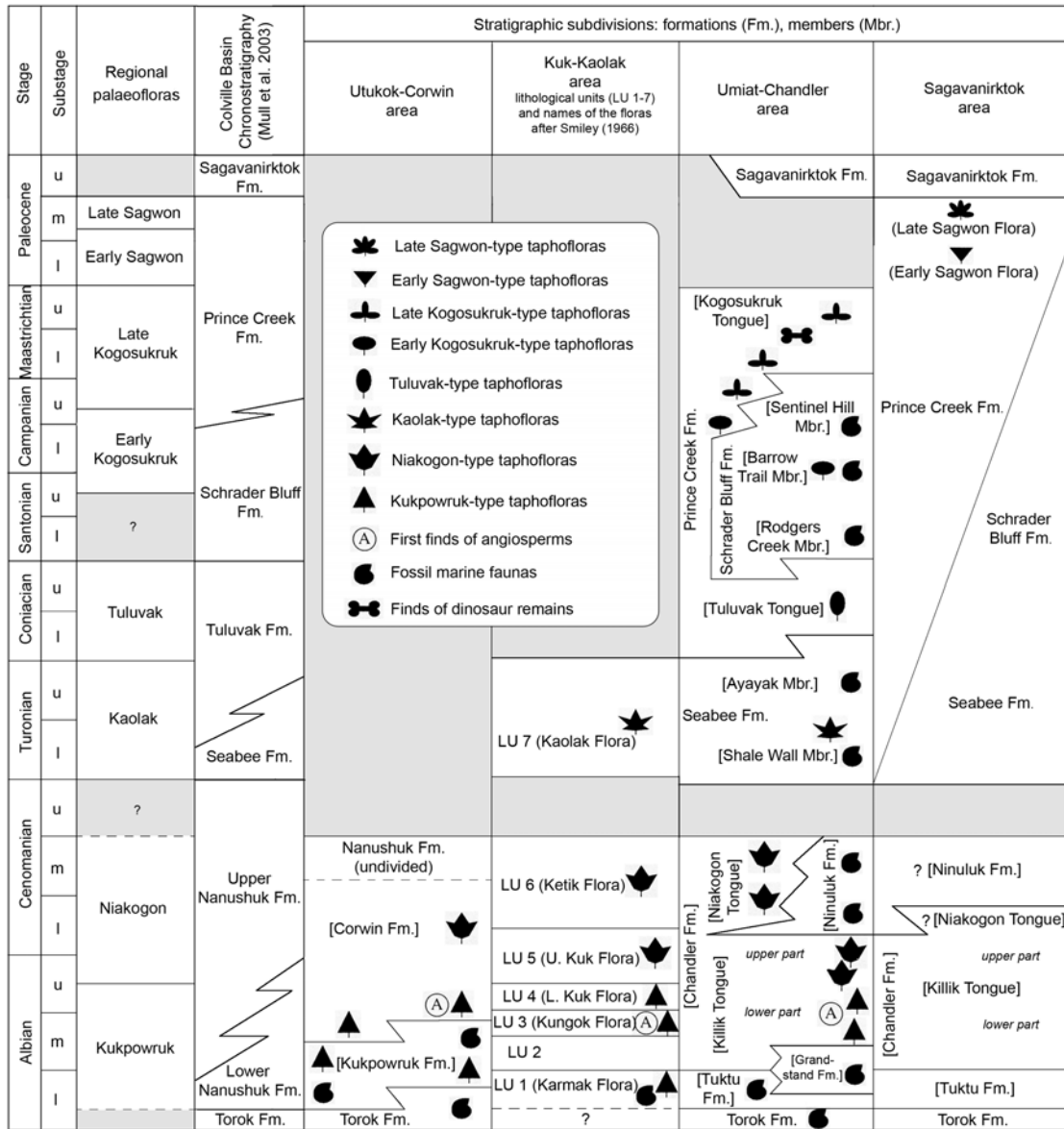
2010 Figure 2



2011

2012 Figure 3.

2013

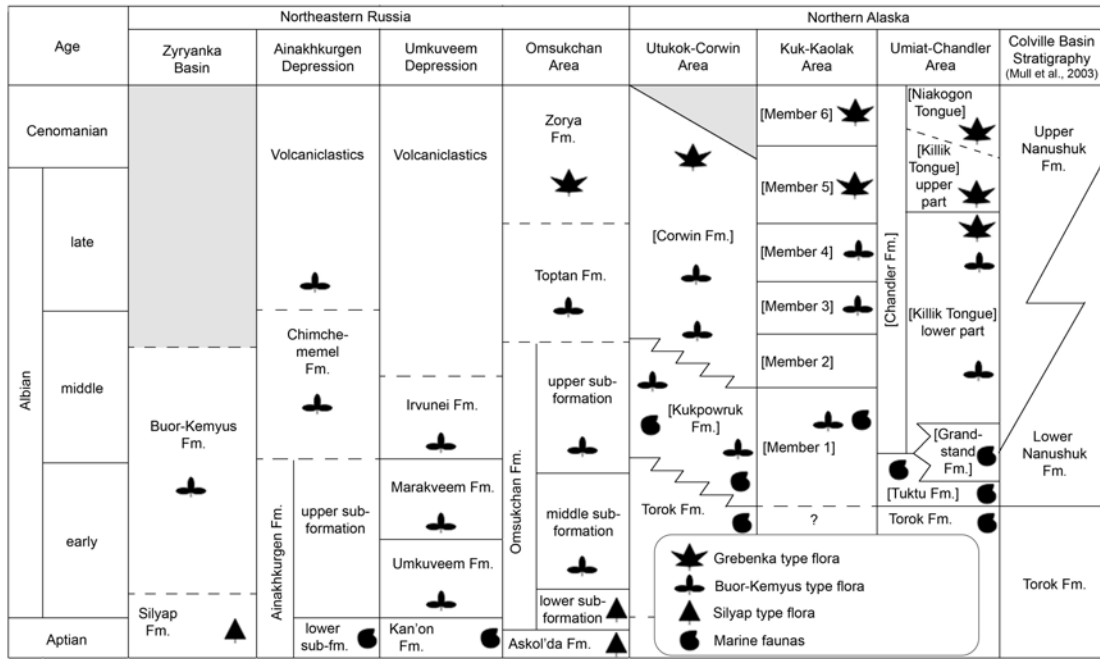


2014

2015 Figure 4.

2016

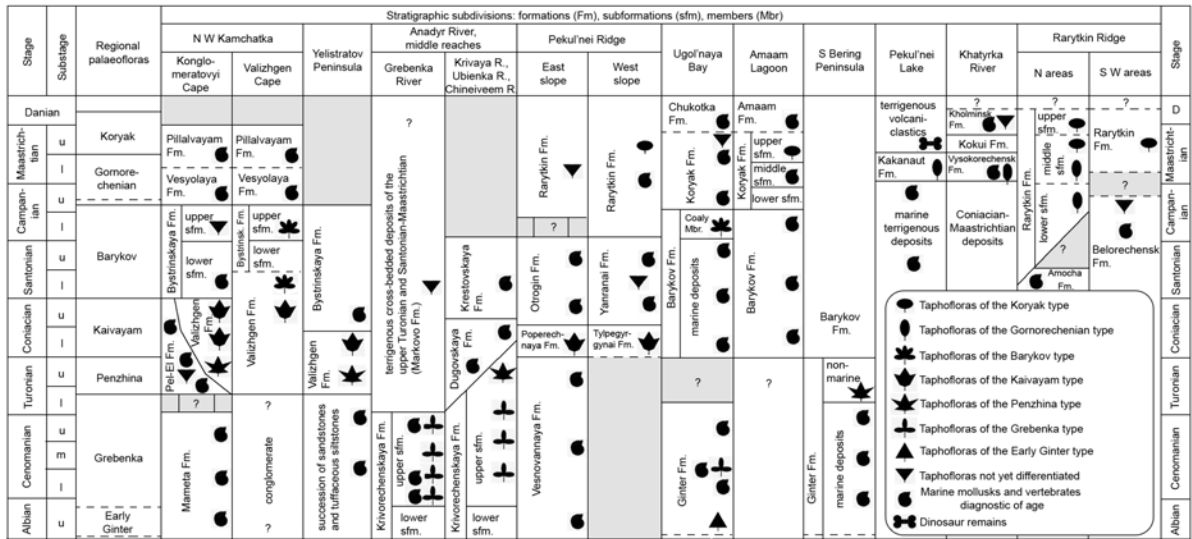
2017



2018

2019 Figure 5.

2020



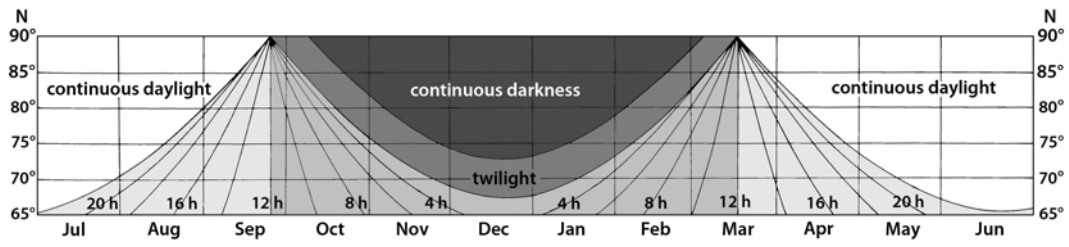
2021

2022 Figure 6.

Stage	Substage	Regional palaeofloras		Stage
		Anadyr-Koryak region	Northern Alaska region	
Thanetian			<i>Late Sagwon</i>	Thanetian
Selandian			<i>Early Sagwon</i>	Selandian
Danian			?	Danian
Maastrichtian	upper	<i>Koryak</i>	<i>Late Kogosukruk</i>	Maastrichtian
	lower	<i>Gornorechenian</i>		
Campanian	upper	<i>Barykov</i>	<i>Early Kogosukruk</i>	Campanian
	lower			
Santonian	upper		?	Santonian
	lower		?	
Coniacian	upper	<i>Kaivayam</i>	<i>Tuluvak</i>	Coniacian
	middle			
	lower			
Turonian	upper	<i>Penzhina</i>	<i>Kaolak</i>	Turonian
	middle			
	lower			
Cenomanian	upper	<i>Grebenka</i>	?	Cenomanian
	middle		?	
	lower		<i>Niakogon</i>	
Albian	upper	<i>Early Ginter</i>	<i>Kukpowruk</i>	Albian
	middle			
	lower			

2023

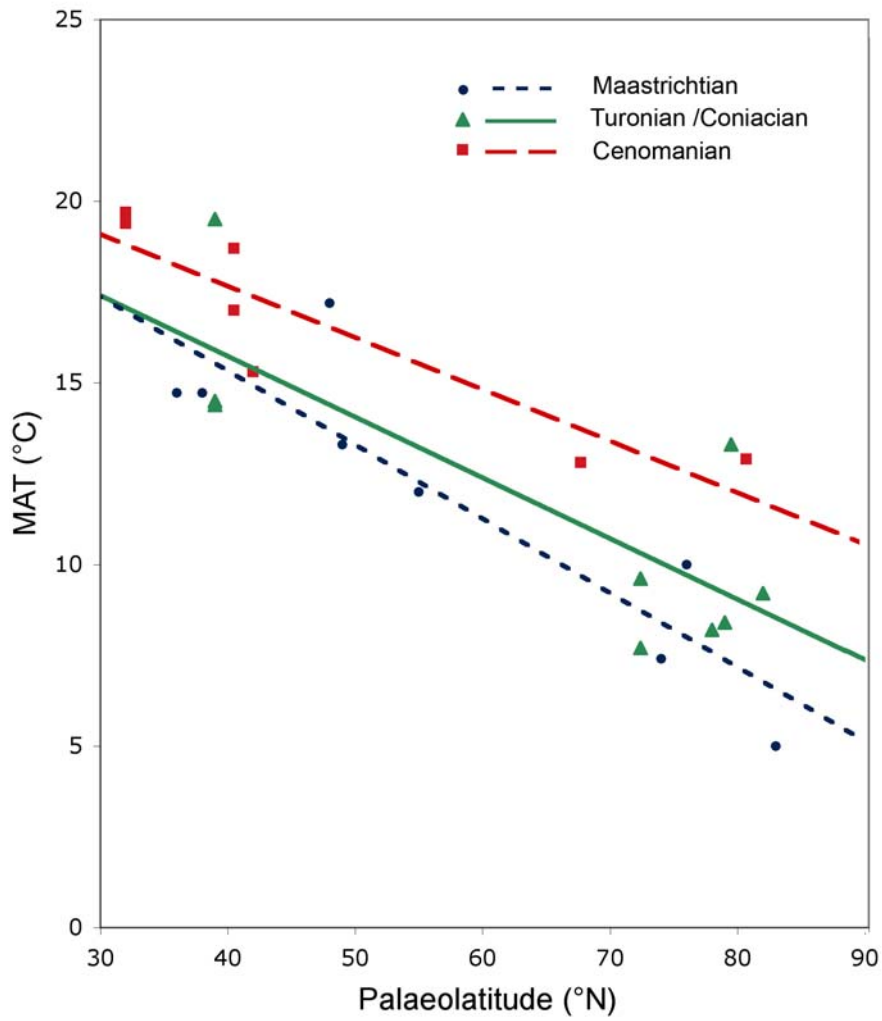
2024 Figure 7.



2025

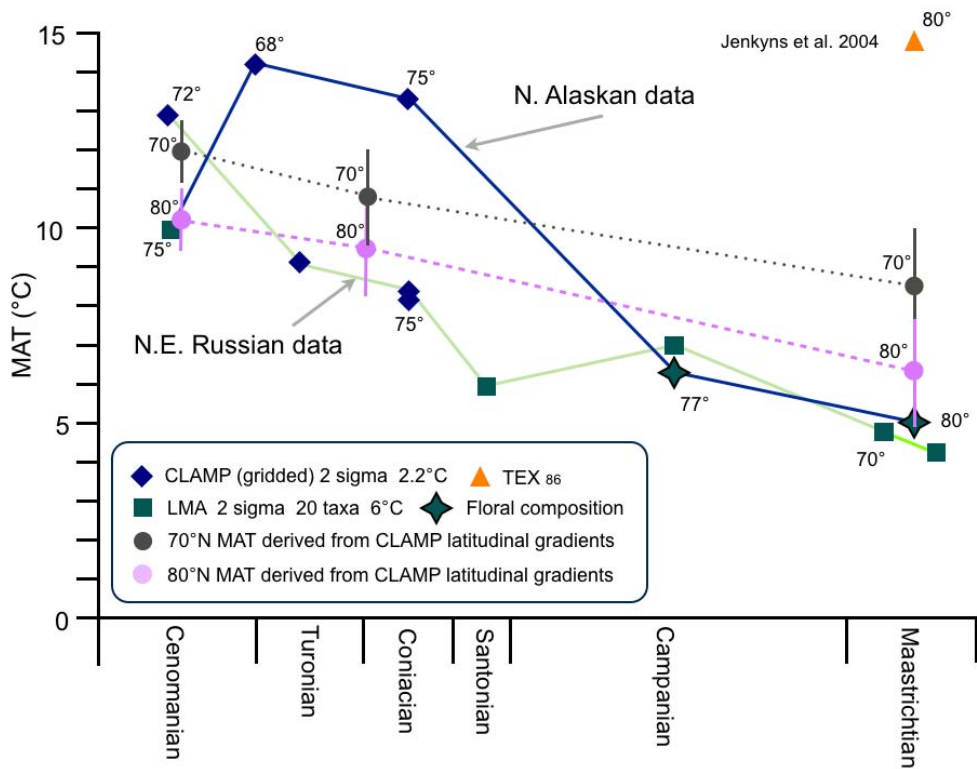
2026 Figure 8.

2027



2028

2029 Figure 9.



2030

2031 Figure 10.

2032



2033

2034 Figure 10.

2035

FLORA (Age Ma)	MAT (°C)	WMMT (°C)	CMMT (°C)	LGS (months)	GSP (mm)	MMGSP (mm)	3WET (mm)	3DRY (mm)	RH (%)	SH (g/kg)	ENTHAL (kJ/kg)	P.Lat (°N)
1. N. Alaska (88)	13.3	19.1	7.9	7.6	793	88	526	184	80.4	10.1	326	80
2. Nov. Sibir (90)	9.2	17.2	1.1	5.8	537	79	504	151	76.7	7.0	310	82
3. Yukon (90)	14.3	21.1	8.0	8.1	1248	158	619	267	84.9	12.4	336	73
4. Vilui B. (95)	12.8	21.0	5.3	7.4	1054	135	623	209	79.9	9.6	324	68
5. Grebenka (98)	12.9	20.8	5.9	7.4	823	93	577	145	72.5	8.0	317	81
6. Kamchatka (90)	7.7	17.7	-2.4	5.1	427	82	466	144	75.4	5.8	304	72
7. Kamchatka (88)	9.6	18.3	1.1	6.0	598	88	519	154	76.0	7.0	310	72
8. Arman R. (88)	8.2	18.7	-2.0	5.3	481	88	481	143	74.1	5.8	304	68
9. Tylpegyrgynai (88)	8.4	18.8	-1.6	5.4	476	82	493	126	71.2	5.3	303	79
10. Chuchle (95)	18.7	27.3	10.4	10.1	1432	158	676	170	67.6	9.4	329	41
11. Vyshehor. (95)	17.0	26.2	8.3	9.4	1786	210	825	231	75.6	10.4	331	41
12. Peruc (95)	16.2	26.2	7.2	8.9	1208	146	631	162	67.9	8.4	322	41
13. Grunbach (80)	14.4	26.7	3.2	8.1	1196	165	654	154	63.6	6.8	314	39
14. Czechsen (85)	14.5	28.7	0.9	8.2	1357	199	707	146	54.9	5.4	309	39
21. Lance (70)	13.3	22.2	5.3	7.6	793	100	540	144	71.6	7.7	317	49
22. Medicine B. (70)	17.2	23.6	11.2	9.4	1440	152	713	218	79.0	11.0	333	48
23. Ripley (70)	14.7	25.3	5.3	8.3	983	119	623	109	57.6	6.3	312	38
24. Tuscaloosa (95)	19.7	24.5	15.1	10.5	1456	131	696	195	76	11.6	338	32
25. Woodbine (95)	19.4	23.5	15.4	10.3	1305	104	686	160	72.3	10.8	334	32
26. Gayhead (84)	19.5	24.3	14.8	10.4	1551	133	794	171	70.4	10.6	334	39
28. Cooper (70)	14.7	25.3	5.3	8.3	983	119	623	108	57.6	6.3	312	36
29. Terekty-Sai (95)	15.3	22.2	9.2	8.5	1070	120	600	191	78	10.3	310	42

STDEV Residuals 1.1 1.4 1.9 0.7 196 26 138 32 5.2 1.0 5 ≈ 5