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1   **Late Quaternary glacier sensitivity to temperature and precipitation**
2   **distribution in the Southern Alps of New Zealand**

3

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17

18   **Abstract**

19   Glaciers respond to climate variations and leave geomorphic evidence that represents an
20   important terrestrial paleoclimate record. However, the accuracy of paleoclimate
21   reconstructions from glacial geology is limited by the challenge of representing mountain
22   meteorology in numerical models. Precipitation is usually treated in a simple manner and yet
23   represents difficult-to-characterize variables such as amount, distribution and phase.
24   Furthermore, precipitation distributions during a glacial probably differed from present-day
25   interglacial patterns. We applied two models to investigate glacier sensitivity to temperature
26   and precipitation in the eastern Southern Alps of New Zealand. A 2-D model was used to
27   quantify variations in the length of the reconstructed glaciers resulting from plausible
28   precipitation distributions compared to variations in length resulting from change in mean
29   annual air temperature and precipitation amount. A 1-D model was used to quantify
30   variations in length resulting from interannual climate variability. Assuming that present-day
31   interglacial values represent precipitation distributions during the last glacial, a range of
32   plausible present-day precipitation distributions resulted in uncertainty in the Last Glacial
33   Maximum length of the Pukaki Glacier of 17.1 km (24%) and the Rakaia Glacier of 9.3 km

34 (25%), corresponding to a 0.5°C difference in temperature. Smaller changes in glacier length

35 resulted from a 50% decrease in precipitation amount from present-day values (−14% and −

36 18%), and from a 50% increase in precipitation amount (5% and 9%). Our results

37 demonstrate that precipitation distribution can produce considerable variation in simulated

38 glacier extents, and that reconstructions of paleoglaciers should include this uncertainty.

39

40 **1. Introduction**

41 Glacial geology is an important terrestrial record of past climate change [e.g. Kaplan et al.,

42 2010; Putnam et al., 2010]. Paleoclimate conditions can be inferred from this record using

43 equilibrium line altitude (ELA) reconstructions based on mapping of paleoglacier shape [e.g.

44 Porter, 1975] or using ice flow models that determine glacier volume [e.g. Anderson and

45 Mackintosh, 2006; Doughty et al., 2013; Kaplan et al., 2013]. While both methods have

46 advantages and disadvantages, the accuracy of the inferred paleoclimate is limited by the

47 challenge of representing mountain meteorology in glacier models. Describing the spatial and

48 seasonal variations of an essentially unchanging climate, and the temporal changes in

49 climatic conditions that are likely to affect the glacier balance also presents potential

50 difficulties to model applications. Near-surface air temperature and precipitation rates are

51 typically assumed to have a linear relationship with altitude, but the interaction of air masses

52 with high topography modifies the distribution of precipitation. Reconstructions of glaciers

53 located in the temperate, westerly-dominated mid-latitudes—for example, the Patagonian

54 Andes [Glasser et al., 2005; Kaplan et al., 2008] and the Southern Alps of New Zealand

55 [Anderson and Mackintosh, 2006; Rother and Shulmeister, 2006; Golledge et al., 2012;

56 Rowan et al., 2013]—reveal compelling evidence for glacier sensitivity to both temperature

57 and precipitation distribution.

58

59 The interaction between rugged, evolving topography and variable air circulation patterns is

60 complex, and the distribution of precipitation in mountainous regions is often difficult to

61 predict. Precipitation peaks do not correlate with the highest topography [Henderson and

62 Thompson, 1999; Schultz et al., 2002; Steenburgh, 2003; Roe, 2005; Anders et al., 2006],

63 and observations are scarce, as high-elevation rain gauges are frequently lacking and these

64 data typically only represent short timespans [Groisman and Legates, 1994]. Moreover,

65 precipitation amount, spatial distribution, temporal distribution and phase will vary with

66 climate change, so present-day precipitation data may not represent conditions during a

67 glacial. As a result, the representation of precipitation in glacier models may be

unsatisfactory and could result in unaccounted-for uncertainties in paleoclimate reconstructions [Rother and Shulmeister, 2006]. Seasonality in lapse rate [Doughty et al., 2013], air temperature and precipitation amount [Golledge et al., 2010] may also modify mass balance. Many glacier modeling studies use summer–winter climatologies that do not consider the detail of these seasonal variations in meteorological variables. Furthermore, many glacier-modeling studies assume that glaciers were in equilibrium with the long-term mean values of temperature and precipitation amount. However, because glaciers also respond to interannual climate variations (climate noise) this assumption is likely to be invalid [Anderson et al., 2014].

The purpose of this paper is to quantify variations in the extents of reconstructed glaciers resulting from a realistic range of precipitation distributions for the Southern Alps of New Zealand. We apply two glacier models to the eastern Southern Alps to demonstrate the need for more realistic representations of orographic patterns of rain and snowfall. We consider whether these glacier reconstructions are precise indicators of past climate change, or if a more realistic approach is to quantify glacier sensitivities to climate and use these to infer an envelope of likely paleoclimate change—an approach previously used by Anderson and Mackintosh [2006] and Golledge et al. [2012] for temperature and precipitation amount. Our paper builds on these previous studies to quantify uncertainties in simulated glacier extents due to precipitation distribution, precipitation phase, interannual climate variability and seasonality.

1.1 The Southern Alps

The Southern Alps of New Zealand (Fig. 1) are an excellent location to explore the climate sensitivity of glaciers [Oerlemans, 1997; Anderson and Mackintosh, 2006; Anderson et al., 2010; Putnam et al., 2010; Doughty et al., 2013]. This 400 km long, ~120 km wide, northeast–southwest-trending mountain range has summit elevations exceeding 3000 m [Tippett and Kamp, 1995; Willett, 1999]. The axial trend of the range is perpendicular to the prevailing westerly winds, resulting in a steep west–east precipitation gradient. The central Southern Alps experience extremely high precipitation of up to 14 m per year on the western (upwind) side of the range, which decreases rapidly to the east [Henderson and Thompson, 1999; Wratt et al., 2000]. The trend in ELA is strongly influenced by this precipitation gradient [Chinn, 1995], suggesting that orographic precipitation exerts a primary control on glaciation [Porter, 1975].

103   New Zealand is one of few landmasses in the southern mid-latitudes, and paleoclimate

104   reconstructions from New Zealand are important for comparison with global records [e.g.

105   Kaplan et al., 2010; Putnam et al., 2012]. The Last Glacial Maximum (LGM) occurred in

106   New Zealand between 24–18 ka [Putnam et al., 2013b]. The late Quaternary geology is well

107   preserved and records frequent and rapid climate change [Alloway et al., 2007; Barrell et al.,

108   2011]. There is little regional variation in bedrock lithology [Cox and Barrell, 2007], so

109   glaciers are unlikely to be modified by their geological setting. Numerical simulations of the

110   Southern Alps icefield [Golledge et al. 2012], the Ohau [Putnam et al., 2013b], the Pukaki

111   [McKinnon et al., 2012] and the Rakaia [Rowan et al., 2013] Glaciers (Fig. 1) demonstrated

112   that LGM mean annual air temperature was between 6°C and 8°C cooler than present-day

113   values and may have been accompanied by a reduction of up to 25% in precipitation.

1.2 Applications of glacier models in the Southern Alps

116   Previous glacier modeling studies in the Southern Alps have focused on the Franz Josef

117   Glacier (Fig. 1) to examine the climate sensitivity of this glacier [Oerlemans, 1997] and the

118   drivers of the advance to the well-preserved Waiho Loop moraine [Anderson and

119   Mackintosh, 2006; Anderson et al., 2008; Alexander et al., 2011]. Oerlemans [1997] and

120   Anderson and Mackintosh [2006] demonstrated that differences in temperature rather than

121   precipitation amount were the major control on the length of this glacier, but that high

122   precipitation values enhanced temperature sensitivity; Oerlemans [1997] showed that the

123   Franz Josef Glacier receded 1.5 km per °C of warming, whereas Anderson and Mackintosh

124   [2006] showed that the glacier advanced at a rate of 3.3 km per °C of cooling. This difference

125   was attributed to *Oerlemans'* unrealistically low precipitation values [Tovar et al., 2008;

126   Shulmeister et al., 2009; Alexander et al., 2011]. Energy-balance calculations for the

127   Brewster Glacier indicated high temperature sensitivity; a 50% change in precipitation

128   amount was required to offset a temperature difference of 1°C [Anderson et al., 2010]. The

129   Pukaki Glacier has a greater temperature sensitivity than the Brewster Glacier; an 82%

130   increase in precipitation amount is required to offset a temperature difference of −1°C

131   [Anderson and Mackintosh, 2012], probably due to the difference in the hypsometry of these

132   glaciers. There may be uncertainty in simulated glacier extents due to bed geometry and

133   subglacial erosion independent of climate change. A flowline model of the Pukaki Glacier

134   indicated that variations in the bed topography could have forced kilometer-scale variation in

135 glacier length to form the two distinct LGM moraine sequences in this valley [McKinnon et
136 al., 2012].

137

138 The poor fit of some simulated glaciers to mapped moraines, particularly when the model
139 features more than one glacier, indicates the need to quantify the uncertainties associated
140 with the application of glacier models to avoid misleadingly precise paleoclimate estimates.
141 In a model reconstruction of glaciers in the eastern Southern Alps, the Rakaia Glacier was
142 under-represented by the LGM simulation that provided the best fit to the geological data
143 compared to the neighbouring Rangitata and Ashburton Glaciers [Rowan et al., 2013]. A
144 model reconstruction of the Southern Alps LGM icefield generally provided a good fit to the
145 glacial geology, although for some glaciers including the Rakaia, this simulation
146 underestimated the LGM terminus positions [Golledge et al., 2012]. Increasing the simulated
147 length of the Rakaia Glacier to reach the LGM extents required either further cooling of −
148 0.25°C from an LGM simulation with a temperature difference of −6.5°C and no change in
149 precipitation amount from present-day values [Rowan et al., 2013], or further cooling of −
150 1.75°C from an LGM simulation with a temperature difference of −6.25°C and a 25%
151 reduction in precipitation amount from present-day values [Golledge et al., 2012].

152

153 **2. Methods**
154 **2.1 The 2-D and 1-D glacier models**
155 We applied a 2-D energy–mass balance and ice flow model implementing the shallow-ice
156 approximation [Plummer and Phillips, 2003], and a 1-D shallow-ice approximation flowline
157 model [*Roe and O'Neal*, 2009] to catchments in the eastern Southern Alps (Fig. 1). We used
158 the 2-D model to investigate how temperature and precipitation modify the energy balance
159 and extents of these glaciers, while the 1-D model was used for experiments investigating
160 fluctuations in glacier length forced by interannual climate variability [cf. Anderson et al.,
161 2014]. The 2-D glacier model has previously been applied to glaciers in the USA [Plummer
162 and Phillips, 2003; Laabs et al., 2006; Refsnider et al., 2008] and New Zealand [Rowan et
163 al., 2013; Putnam et al., 2013a]. These glacier models are based on the shallow-ice
164 approximation developed for large ice sheets with shallow bed topography [Hutter, 1983],
165 which represents the longitudinal but not the transverse stresses of flowing ice and is
166 unsuitable for glaciers with dominantly steep bed topography [Le Meur et al., 2004].
167 Previous studies have successfully applied the shallow-ice approximation to glaciers in New
168 Zealand [Anderson and Mackintosh, 2006; Rowan et al., 2013] and elsewhere [Oerlemans et

169 al., 1998; Kessler and Anderson, 2006; Refsnider et al., 2008] and we consider this

170 approximation valid for the large, low-angle valley glaciers that occupied the eastern

171 Southern Alps.

172

173 The model domain includes the Rakaia to the Pukaki valleys (Fig. 1). The Land Information

174 New Zealand (LINZ) 50-m digital elevation model (DEM) was resampled to a 200-m grid

175 spacing to describe topography (Table 1). Present-day ice volumes were removed from the

176 DEM before applying the glacier models following the method of Golledge et al. [2012]

177 using glacier outlines defined by LINZ and assuming a uniform basal shear stress (τ_b) of 150

178 kPa

179

180 $H = \tau_b / (\rho * g * sin \alpha)$, (1)

181

182 where H is ice thickness, ρ is the density of pure glacier ice (917 kg m^{-3}) [Cuffey and

183 Paterson, 2010], g is acceleration due to gravity (9.81 m s^{-1}) and α is the glacier surface slope

184 taken from the resampled DEM. Model parameter values followed Rowan et al. [2013] for

185 the Rakaia–Rangitata Glaciers (Table 1). After an initial simulation for a particular

186 temperature difference, the simulated glaciers were added to the DEM to iteratively

187 recalculate mass balance across the glacier allowing for the increase in surface elevation with

188 greater ice volume. Calculated mass balance and DEM topography were used as inputs to the

189 ice flow model to calculate ice thickness. Results from the ice flow model were considered

190 acceptable when the integrated mass balance (the difference between accumulation and

191 ablation across the entire glacier) was within 5% of steady state.

192

193 We tested the variability in glacier volume from a baseline model of the present-day climate

194 resulting from; uniform differences in mean annual air temperature (hereafter referred to as

195 temperature), for example, present-day mean annual air temperature minus 1°C (hereafter

196 ΔT); and multiplicative differences in precipitation amount, for example, 75% of present-day

197 precipitation amount (hereafter P). Temperature difference is defined here as an increase or

198 decrease in temperature calculated as 30-year means from daily measurements. Elsewhere in

199 the glaciological literature, temperature difference may be referred to as "temperature

200 change", implying variation in temperature throughout each simulation.

201

202 **2.2 Climatological data**

203 The climate inputs to our baseline model (Table 1) were based on 123 automatic weather

204 stations (AWS) in the national climate database CliFlo (http://cliflo.niwa.co.nz/). We used

205 30-year (1971–2000) monthly mean and daily standard deviation values for temperature,

206 monthly means for relative humidity and wind speed, and 30-year mean annual values for

207 cloudiness. Although interannual variability was observed in the meteorological data, we

208 used 30-year mean values as input to the 2-D model, as variations in climate with a shorter

209 period than the glacier's response time are unlikely to produce the magnitude of length

210 fluctuations we are examining (10–80 km length fluctuations).

211

212 In the baseline model, precipitation distribution was defined using the National Institute of

213 Water and Atmospheric Research (NIWA) 500-m gridded data [Tait et al., 2006].

214 Comparison with river flow measurements indicated that the NIWA data are within 25% of

215 the total water input to the catchments in question [Tait et al., 2006], and probably record

216 most rainfall but only some snowfall due to the limitations of standard precipitation gauging

217 techniques [Goodison, 1978; Yang et al., 1998]. We applied the method of Yang et al. [1998]

218 for a standard rain gauge to estimate the proportion of both precipitation phases that are not

219 recorded by these gauges. The difference in amount between the gauge-estimated values and

220 the modeled precipitation for the model domain was 11% for rainfall and 49% for snowfall,

221 implying that the total annual precipitation amount was 144% of that recorded. To reflect this

222 estimate of gauge undercatch, we increased the precipitation input to the 2-D baseline model

223 by these ratios and again simulated glacier lengths. After increasing precipitation amount in

224 line with this estimate, the LGM Rakaia Glacier simulated under the same ΔT (–6.5°C) was

225 4.1 km (10%) longer, which equated to a ΔT of less than –0.5°C.

226

227 **2.3 Experimental design**

228 Glacier sensitivity to temperature, precipitation amount and distribution, interannual climate

229 variability, temperature seasonality, precipitation seasonality and precipitation phase was

230 investigated for the eastern Southern Alps. We considered glacier sensitivity to climate

231 change in terms of both change in mass balance and change in glacier length (volume) [cf.

232 Oerlemans, 1997]. We performed five sets of experiments, each comprising multiple model

233 simulations, to quantify uncertainty in simulated glacier extents resulting from:

234 • Experiment 1: Differences in temperature (ΔT) from the baseline model describing
235 present-day climate

- Experiment 2: Differences in precipitation amount (P) from the baseline model within a plausible worldwide present-day range
- Experiment 3: Precipitation distribution using five estimated precipitation distributions and three statistical approximations of precipitation for the central Southern Alps (Table 2)
- Experiment 4: Interannual climate variability defined by the present-day standard deviation of mean melt season (December–February) temperature and annual precipitation amount.
- Experiment 5: Change in seasonality (S), defined here as an increase in monthly summer (October–March) temperatures of up to 3°C while winter temperatures remain unchanged, combined with change in winter and summer monthly precipitation amounts relative to present-day values

Despite the possible reduction in LGM precipitation amount of up to 25% indicated by previous glacier modeling [Golledge et al., 2012], all experiments used the same values for P in the present-day and LGM simulations, apart from those simulations where P was explicitly varied. This approach allowed us to isolate the sensitivity to P and to compare this directly to differences in temperature and precipitation distribution over a range of climate scenarios. Experiments 1 and 2 tested a range of plausible values of ΔT and P during the glacial. Experiments 3 and 5 were designed to simulate specific climate scenarios: (1) the present-day climate applied to the study area by Rowan et al. [2013] (Table 1); (2) a late glacial paleoclimate indicated by the advance of the Rakaia Glacier to produce the Prospect Hill moraine at 16.25 ± 0.34 ka, equivalent to $\Delta T = -3.0$°C [Putnam et al., 2013a]; (3) a late glacial paleoclimate indicated by the Rakaia Glacier advance to the Double Hill moraine at 16.96 ± 0.37 ka, equivalent to $\Delta T = -4.5$°C [Putnam et al., 2013a]; and (4) a paleoclimate representing the LGM at ~21 ka, equivalent to $\Delta T = -6.5$°C [Golledge et al., 2012; Rowan et al., 2013]. Experiment 4 simulated two scenarios: (1) a Late Glacial advance resulting in a reduction in ELA of ~100 m around ~11 ka [Kaplan et al., 2013] equivalent to $\Delta T = -1.25$°C; and (2) the LGM scenario.

Our 2-D glacier model calculated snowfall using the number of days per month for which the daily air temperature in each cell was below a critical value for rain–snow partitioning, using the mean monthly air temperature and its daily standard deviation [Plummer and Phillips,

269 2003]. The value used for the critical temperature at which precipitation falls as snow varies

270 between modeling studies. We tested a range of critical temperatures from 0–3°C (results not

271 presented here) which resulted in a 1.9 km (7%) uncertainty in the length of the Pukaki

272 Glacier under present-day climate, and a 1.1 km (2%) uncertainty in the glacier length under

273 the Double Hill scenario. The critical temperature was set to 2°C for all simulations reported

274 in this paper. The proportion of annual precipitation falling as snow across the model domain

275 was 7% under the present-day scenario, 46% under the Prospect Hill scenario, 86% under the

276 Double Hill scenario, and 92% under the LGM scenario.

277

278 2.3.1 Experiment 1: Temperature

279 Variations in glacier extent due to ΔT were tested for the Rakaia–Rangitata catchments from

280 –9.0°C to 0°C in 0.5°C increments to find an ELA equivalent to the LGM (799 ± 50 m) and

281 the Prospect Hill advance (1540 ± 50 m). Results are presented in Section 3.1.

282

283 2.3.2 Experiment 2: Precipitation amount

284 P was varied from 25% to 400% of present-day values in 10% or 25% increments to

285 investigate glacier sensitivities in the Rakaia–Rangitata catchments. ELAs were calculated

286 and glacier length simulated under each of the four climate scenarios. Results are presented in

287 Sections 3.1 and 3.2. Results from Experiments 1 and 2 (Fig. 2) are compared to those

288 produced for the Franz Josef Glacier [Anderson and Mackintosh, 2006] and for the Irishman

289 Glacier [Doughty et al., 2013].

290

291 2.3.3 Experiment 3: Precipitation distribution

292 Orography regulates precipitation distribution over the Southern Alps as the range axis trends

293 perpendicular to the prevailing westerlies. Therefore, the distribution of precipitation is

294 primarily a function of the distance from the west coast of the South Island rather than a

295 function of elevation [Griffiths and McSaveney, 1983; Sinclair et al., 1997; Henderson and

296 Thompson, 1999; Ibbitt et al., 2001; Tait et al., 2006] (Fig. 3B and C). When plotted across

297 the range, rain-gauge data show a rather wet region upwind on the western side of the range

298 (3–4 m per year) compared to a much drier region east of the range (less than 2 m per year)

299 (Fig. 3A). The scarcity of rain gauges in the high-elevation region 20–60 km downwind of

300 the west coast with which to document this dramatic precipitation gradient leaves open the

301 possibility that glacier simulations could be highly sensitive to the peaks and distribution of

302 precipitation in this region. We assume that precipitation during the LGM was unlikely to

have increased beyond the present-day worldwide maximum, equivalent to an 85% increase in Southern Alps precipitation [Henderson and Thompson, 1999]. As the last glacial precipitation distribution is unknown, we instead experiment with present-day precipitation distributions for the region and acknowledge that there is unresolved uncertainty when using these to represent last glacial precipitation.

To explore the sensitivity of glacier model results to uncertainty in precipitation distribution, we tested five different estimated present-day precipitation distributions and three statistical approximations of these data for the central Southern Alps: (1–3) three annual rainfall profiles [Griffiths and McSaveney, 1983; Wratt et al., 1996; Henderson and Thompson, 1999]; (4) gridded 500-m monthly NIWA rainfall data [Tait et al., 2006]; and (5) rain-gauge data from the CliFlo database fitted with a least-squares cubic spline approximation that preserves shape. Based on these rainfall data, we tested three statistical approximations of precipitation distribution; (6) a linear relationship linking rainfall to elevation derived from regression of the CliFlo data, and grids of (7) median, and (8) mean monthly values derived from the NIWA data. Where rainfall data provided only annual values, these were divided into monthly totals using the present-day distribution in the NIWA data. These 1-D profiles represent rainfall along a section oriented at 130° through the center of the model domain (Fig. 1A). A mean topographic profile defined from a 40 km-wide swath centered on this transect (Fig. 3) was used to interpolate the linear precipitation values in test (6). The rainfall data were converted to 2-D grids by duplicating these profiles along-strike of the range axis perpendicular to the transect. The mean annual precipitation amount varied with the choice of precipitation distribution. For the distributions in tests (1) to (4), mean annual precipitation across the model domain varied between the different datasets from 1134–1602 mm (Table 2). Results are presented in Section 3.3.

2.3.4 Experiment 4: Interannual climate variability

Interannual variability in mean melt season temperature and annual precipitation amount (often described as white noise) can cause kilometer-scale fluctuations in glacier length independent of climate change [e.g. Oerlemans, 2001; *Roe and O'Neal*, 2009; Roe, 2011]. These fluctuations add a one-sided bias to paleoclimate estimates derived from the moraine record, as the terminal moraine position for a particular advance represents the maximum down-valley excursion of the glacier rather than the mean glacier length [Anderson et al., 2014]. We used a 1-D flowline model with variable width to determine the mean length for

337 the Late Glacial (ΔT = −1.25°C) and LGM (ΔT = −6.5°C) Rakaia Glacier. To maintain

338 coherence between our glacier models, mass balance calculations produced using the 2-D

339 model were used to describe the 1-D mass balance profiles for this glacier. The advantage of

340 using a 1-D model to test sensitivity to interannual climate variability is that we can

341 efficiently run hundreds of simulations with independent white-noise realizations, therefore

342 allowing us to establish the most probable mean glacier length for a particular advance.

343

344 Two independent white-noise realizations for mean melt-season temperature and annual

345 precipitation amount were used for each simulation. The temperature realization was

346 modified by a random normal distribution of annual values using the standard deviation of

347 mean December–February temperature (0.8°C) from the Lake Coleridge AWS (Fig. 1). The

348 annual precipitation realizations were modified by a random normal distribution of annual

349 values using the standard deviation of precipitation data from AWS on the west side of the

350 range (870 mm per year). Data derived from AWS in the eastern Southern Alps do not

351 capture the precipitation variability in present-day glacier accumulation areas (Fig. 3) [Woo

352 and Fitzharris, 1992]. However, the Woo and Fitzharris [1992] data provide a minimum

353 estimate of the annual precipitation variability in the glacier accumulation areas, because

354 these data are derived from a low-elevation AWS and neglect the effects of wind-blown snow

355 and avalanching. Mass balance was perturbed from a mean state using a melt factor of 0.9 m

356 water equivalent per °C per year; a representative value based on a global compilation of

357 present-day melt factors for ice [Anderson et al., 2014]. Results are presented in Section 3.4.

358

359 2.3.5 Experiment 5: Seasonality

360 The range of monthly mean summer and winter temperatures and precipitation amounts are

361 likely to change during a glacial [Nelson et al., 2000; Golledge and Hubbard, 2009]. Sea

362 surface temperature records for the LGM indicate that temperature seasonality (S) was 3°C in

363 Canterbury Bight [Nelson et al., 2000; Drost et al., 2007] and 2°C across the region [Barrows

364 and Juggins, 2005]. Previous LGM regional climate modeling indicated S equivalent to an

365 increase in summer temperatures of 0.7°C in eastern South Island, lower than the mean value

366 of 1.1°C for New Zealand, and summer precipitation slightly higher and winter precipitation

367 slightly lower than present-day values [Drost et al., 2007]. To quantify how a realistic

368 variation in S from present-day values affects ELA, we compared glaciers simulated with the

369 baseline model to simulations using an estimated maximum LGM seasonality of

370   • Summer temperature (October–March) = Present-day temperature + 3°C

11

371 • Summer precipitation (October–March) = Present-day precipitation * 1.11

372 • Winter temperature (April–September) = Present-day temperature

373 • Winter precipitation (April–September) = Present-day precipitation * 0.97

374 These values for S follow the results of regional meteorological modeling of the LGM in

375 New Zealand by Drost et al. [2007]. Results are presented in Section 3.5.

376

377 **3. Results**

378 The variability in simulated glacier lengths and ELAs resulting from sensitivities to

379 difference in temperature, precipitation amount, precipitation distribution, interannual climate

380 variability and seasonality are presented here. Transient ice flow calculations show that

381 steady state was reached within 400 years for $\Delta T = 0.5°C$, consistent with the response times

382 of up to hundreds of years estimated using analytical solutions [Jóhannesson et al., 1989].

383 Response time for the present-day Tasman Glacier in the Pukaki valley was estimated as 20–

384 200 years by assuming terminus ice thickness of ~500 m and an ablation rate beneath the

385 thick debris layer of 2.5 m per year. The range of values indicates the potentially large

386 uncertainty in predictions of melt rates beneath supraglacial debris [Herman et al., 2011].

387

388 **3.1 Experiment 1: Glacier sensitivity to change in temperature and precipitation**

389 Present-day and LGM ELAs simulated using the 2-D glacier model were compared to ELAs

390 reconstructed using the accumulation-area ratio method (with a value for the ratio of the

391 accumulation area to the total glacier area of 0.6 ± 0.05) for the Pukaki and Tekapo valleys

392 [Porter, 1975]. The parameter space required to produce LGM and Prospect Hill ELAs was

393 tested for the Rakaia–Rangitata catchment and compared to previous estimates for the Franz

394 Josef Glacier [Anderson and Mackintosh, 2006] and the Irishman Glacier [Doughty et al.,

395 2013]. We tested ΔT and P values to produce ELAs required to advance glaciers to the LGM

396 and Prospect Hill terminal moraines. P was limited within a realistic range for the present-

397 day interannual variability in the Southern Alps (80–140% of present-day values) and up to

398 the present-day worldwide maximum (185%) [Henderson and Thompson, 1999].

399

400 LGM ELAs were simulated under conditions where $\Delta T = -8.0°C$ to $-5.5°C$ and P = 80% to

401 175% of present-day values. If change in precipitation amounts were restricted to the range of

402 regional interannual variability, the LGM would have occurred with ΔT of $-8.25°C$ to $-$

403 6.0°C. We assume little or no change in precipitation amount during the LGM, to give

404 solutions of ΔT = −6.5°C and P = 100% for the LGM, and ΔT = −3.0°C and P = 100% for the

405 Prospect Hill advance [Putnam et al., 2013a]. The sensitivity of glacier extent to ΔT was

406 tested for the Rakaia and Rangitata Glaciers and varied with the absolute value of ΔT (Fig. 4).

407 The length of the Rakaia Glacier increased by at least 37% with ΔT = 0.5°C when the

408 absolute value of ΔT was minimal (less than −2.0°C) compared to present-day conditions.

409 The relative change in glacier length decreased with moderate differences in temperature (ΔT

410 = −3.5°C to 5.0°C) to 8%, then increased with greater absolute differences in temperature (ΔT

411 = −5.5°C to −6.5°C) to ~23%. A similar trend was found for the Rangitata Glacier (Fig. 4),

412 and the decrease in glacier length change occurs when the glaciers extend into the trunk

413 valleys where bed slopes are lower. Under the LGM scenario, ΔT = −1°C increased the length

414 of the Rakaia Glacier by 28.7 km (51%).

415

416 **3.2 Experiment 2: Glacier sensitivity to precipitation amount**

417 Glacier sensitivity to precipitation amount was tested for the Rakaia Glacier by varying P

418 from half to twice the present-day values under present-day, Double Hill, and LGM scenarios

419 (Table 3). This range of P values tested exceeds the present-day interannual variability of

420 precipitation amount in the Southern Alps. Under present-day conditions, halving P produced

421 no glacier ice at the headwall of the Rakaia Glacier and reduced the length of the Pukaki

422 Glacier by 13.4 km (−51%). A 50% increase in P increased the length of the Rakaia Glacier

423 by 2.2 km (78%) and the Pukaki Glacier by 3.7 km (14%). Doubling P increased the length

424 of the Rakaia Glacier by 5.7 km (203%) and the Pukaki Glacier by 13.9 km (53%). Under

425 Double Hill conditions, halving P reduced the length of the Rakaia Glacier by 6.6 km (−18%)

426 and reduced the length of the Pukaki Glacier by 10.2 km (−14%), whereas a 50% increase in

427 P increased the length of the Rakaia Glacier by 3.1 km (9%) and the Pukaki Glacier by 3.5

428 km (5%). The experiment under Double Hill conditions where P was doubled did not reach a

429 stable solution due to an unrealistically positive mass balance. Under LGM conditions, a 25%

430 increase in P produced the equivalent increase in length of the Rakaia Glacier to ΔT = −

431 0.5°C. Sensitivity to P was lower for the Double Hill simulations, requiring a greater increase

432 (P = 150% rather than 125%) to produce the change in glacier extent resulting from ΔT = −

433 0.5°C (Fig. 2). For the LGM scenario, glacier sensitivity to ΔT decreases as P exceeds the

434 present-day worldwide maximum; if P was double the present-day value, ΔT of −0.5°C

435 would be equivalent to an increase in P of 50%, indicating that P modifies the temperature

436 sensitivity of these glaciers.

437

3.3 Experiment 3: Glacier sensitivity to precipitation distribution

The choice of precipitation distribution had a considerable influence on the extent of the simulated glaciers (Fig. 5 and Table 3) and the regional ELA (Table 2). The precipitation peak in each profile is located upwind of the main drainage divide and precipitation amounts are similar downwind of 70 km from the west coast (Fig. 3), suggesting that glaciers are sensitive to the volume of precipitation delivered within the zone up to 30 km downwind of the main drainage divide (32 km from the west coast; Fig. 3). Under both present-day and Double Hill conditions, glacier extents simulated using published precipitation distribution profiles were greatest when using the NIWA data, followed by the CliFlo, then Wratt et al. [1996] profiles, and least with the Griffiths and McSaveney [1983] and Henderson and Thompson [1999] profiles (Fig. 5 and Table 3). The change in ELA using published precipitation distribution profiles relative to the NIWA results was greatest with the Henderson and Thompson [1999] data; ELA change of 333 m under present-day conditions and 154 m under the Double Hill scenario (Table 2). Across the range of precipitation distributions tested, the variation in glacier length under present-day conditions was 24.5 km (93.5%) for the Pukaki Glacier and 2.8 km (100%) for the Rakaia Glacier. Under Double Hill conditions, the variation in length was less for the Pukaki Glacier (17.1 km; 24%) and greater for the Rakaia Glacier (9.3 km; 25%) compared to present-day conditions although the absence of ice in four of the present-day Rakaia simulations affected this result (Table 3).

Results produced using mean and median precipitation distributions show that, although the maximum precipitation amount for these experiments was less than that in the NIWA gridded data, ELAs were lower (–438 m and –451 m) as the value for the minimum precipitation amount was greater than that from the NIWA data. Change in ELA due to precipitation distribution was greatest under present-day climate conditions (191 m) compared to the Double Hill advance (91 m) (Table 2). The linear regression of precipitation measurements gave the lowest value for total precipitation amount and produced an ELA ~100 m lower than those simulated using the NIWA data for the present-day and Double Hill scenarios (Table 2). Under LGM conditions, glacier extents calculated using the linear regression were similar to those produced under Double Hill conditions using the NIWA data, equivalent to $\Delta T = 2°C$. If we exclude those glaciers simulated using statistical approximations of precipitation distribution and consider only the glaciers simulated using the five estimated rainfall distributions, then the change in ELA due to precipitation distribution was 171 m under present-day conditions and 91 m for the Double Hill scenario. Of the variables tested in our

14

472 experiments, after difference in temperature, glaciers were most sensitive to precipitation

473 distribution.

474

475 **3.4 Experiment 4: Glacier sensitivity to interannual climate variability**

476 We tested the effect of interannual climate variability on glacier length using a 1-D model of

477 the Late Glacial ($\Delta T = -1.25°C$; ~12 km long glacier) and LGM ($\Delta T = -6.5°C$; ~80 km long

478 glacier) advances of the Rakaia Glacier. The Late Glacial glacier was formed by three

479 tributaries converging within 3 km of the maximum glacier extent. To capture this complex

480 glacier geometry, we modeled all three tributary glaciers and fed the two smaller tributaries

481 into the larger trunk glacier. As a result, these simulations consider the terminus fluctuations

482 resulting from the independent response of each of the three tributary glaciers [cf.

483 MacGregor et al., 2000; Zuo and Oerlemans, 1997]. We compared glacier extents simulated

484 using the 1-D model to those from 2-D simulations; a 1.25°C increase in mean summer

485 temperature produced a 5 km recession from the Late Glacial maximum that is similar to the

486 difference in extent between the Late Glacial and present-day glaciers, demonstrating that our

487 1-D model is reasonably sensitive to summer temperature perturbations in comparison to the

488 2-D model.

489

490 Each Late Glacial simulation ran for 1000 years [e.g. Kaplan et al., 2010]. 1000 simulations

491 were used to estimate the most probable mean glacier length for the Late Glacial advance,

492 which was ~1300m shorter (10% of the maximum glacier length; equivalent to $\Delta T = +0.2°C$)

493 than the terminal moraine. The standard deviation of the mean length from the most likely

494 mean length was 560 m, and the standard deviation of glacier length was 825 m. These

495 results imply that paleoclimate estimates from the Late Glacial terminal moraine position will

496 overestimate the mean glacier length by 10%. This variability accounts for 26% of the change

497 in length required to advance from the present-day position to the Late Glacial terminus. The

498 standard deviation of the modeled snowline elevation was 110 m, which is within the window

499 of present-day snowline variability in New Zealand [WGMS, 2013]. The standard deviation

500 of net mass balance, summer balance and winter balance were 1.15 m, 0.9 m, and 0.7 m per

501 year. These simulated values are similar to mass balance measurements made for the Ivory

502 Glacier (~15 km to the north of the Rakaia) where the standard deviation of annual-mean

503 mass balance, summer balance and winter balance were 1.1 m, 0.63 m and 0.87 m per year

504 between 1970 and 1975 [WGMS, 2013].

505

506 Each LGM simulation ran for 4000-years. 100 simulations were used to estimate that the

507 most probable LGM mean glacier length was 2.3 km shorter (2.8% of the maximum glacier

508 length; equivalent to $\Delta T = +0.1°C$) than the terminal moraine defined by Shulmeister et al.

509 [2010]. The standard deviation of the mean length from the most likely mean length was 840

510 m, and the standard deviation of the glacier length was 1200 m. These results imply that

511 paleoclimate estimates using the LGM terminal moraine position will overestimate the mean

512 glacier length by 2.8%. The standard deviation of the modeled snowline elevation was 100 m.

513 The standard deviation of annual-mean mass balance, summer balance and winter balance

514 were 1.09 m, 0.72 m and 0.82 m per year.

515

516 **3.5 Experiment 5: Glacier sensitivity to seasonality**

517 Glacier sensitivity to S was tested for ΔT from 0°C to −7°C using the maximum estimated

518 LGM seasonality (Fig. 6). Under present-day conditions, the ELA was reduced by just 3 m

519 due to S. ELA change due to S was greatest when glaciers were less extensive, but did not

520 exceed 41 m across the range of ΔT values tested. LGM ELAs were 13 m lower due to

521 changes in S than those for the same climate scenario using present-day values for

522 seasonality. In comparison, ΔT of −1°C under the LGM scenario resulted in a decrease in

523 ELA of 146 m—much greater than that due to changes in S (Fig. 6). Glacier sensitivity to

524 seasonality was not sufficient to be resolved beyond the model uncertainty, and our results

525 indicate the limitations of glacier modeling as a means of reconstructing the finer details of

526 LGM paleoclimates from the geological record.

527

528 **3.6 Summary of results**

529 Glaciers were sensitive to differences in mean annual air temperature (ΔT), the distribution of

530 precipitation and precipitation amount (P) (Table 3). Glacier sensitivity to seasonality in

531 temperature and precipitation amount and to interannual climate variability was within the

532 uncertainty ascribed to the climatological parameter values used in our simulations. Based on

533 our results and previous testing of the uncertainty associated with the model parameter values

534 by Plummer and Phillips [2003], we consider the minimum ΔT value that can be resolved to

535 be 0.25°C. Therefore, we consider only those variations in simulated glacier length that

536 exceed those produced by $\Delta T = 0.25°C$ to indicate significant climate sensitivity. For the

537 Rakaia Glacier under present-day conditions, the change in glacier length indicating

538 significant climate sensitivity is 1.9 km (13%). Under the Double Hill scenario this value is

539 3.0 km (15%). Under the LGM scenario this is 4.7 km (12%).

16

540

541 The percentage change in glacier length varied with the magnitude of ΔT. Glacier length

542 percentage change from the present-day extent was least under intermediate differences in

543 temperature (ΔT = −3.5°C to 5.0°C). Glacier length change per degree ΔT was greatest when

544 glaciers were very small or very large (approaching their LGM limits) (Fig. 4). Small glaciers

545 advanced more rapidly with relatively small changes in mass balance, and they advanced at a

546 faster rate as tributary glaciers merged into the main valley glaciers. ΔT of 0.5°C offset

547 values of P within the present-day worldwide range [Henderson and Thompson, 1999],

548 indicating that change in LGM precipitation amount had a minor effect on glacier extents in

549 the Southern Alps in comparison to difference in temperature.

550

551 Glaciers were sensitive to change in precipitation distribution, with change in glacier length

552 of 25% occurring across the range of precipitation distributions tested under the Double Hill

553 scenario, equivalent to ΔT of at least 0.5°C. If a linear regression linking precipitation

554 distribution to the topographic surface based on measurements from AWS was used instead,

555 then the offset in simulated mass balance for LGM conditions was equivalent to ΔT of 2°C.

556 Using uniform precipitation distributions with values taken from the mean and median of

557 regional values gave unrealistic mass balance for each climate scenario, with ELA

558 depressions of ~400 m relative to using the NIWA data. Glacier sensitivity to precipitation

559 distribution was greatest within 30 km downwind of the main drainage divide where the

560 largest accumulation areas occur, and sensitivity to both precipitation amount and distribution

561 decreased with increasing cooling (increased ΔT). Using our definition of model resolution

562 (ΔT = 0.25°C), the importance of variations in glacier length produced by P decreased with

563 increased ΔT, but the variation in glacier length produced using a range of plausible present-

564 day precipitation distributions remained significant across all of our climate scenarios.

565

566 **4. Discussion**

567 The use of glacier models to estimate paleoclimate requires assumptions about a number of

568 climatological and glaciological parameters. In this paper, we explored glacier sensitivity to

569 paleoclimate variables by comparing the glacier volumes simulated with a realistic range of

570 values for difference in temperature, precipitation amount, precipitation distribution,

571 interannual climate variability and seasonality. Here we discuss the implications of these

572 sensitivities for the reconstruction of LGM glaciers in the Southern Alps. The Rakaia Glacier,

573 which has proved particularly challenging in previous modeling studies, is discussed in

574 detail. We also discuss further sources of uncertainty that should be considered in the
575 application of glacier models—factors that influence mass balance such as radiative fluxes
576 and avalanching, the representation of topography, and the choice of model domain and grid
577 spacing.

578

579 **4.1 LGM climate variability in the Southern Alps**

580 ELA sensitivity to ΔT and P for the Rakaia–Rangitata Glaciers show a similar relationship to
581 those calculated for the Franz Josef Glacier [Anderson and Mackintosh, 2006]. Our present-
582 day climate simulations gave an ELA for the Pukaki and Tekapo valleys similar to those
583 reconstructed by Porter [1975] (Fig. 6). Our LGM simulations produced an ELA slightly
584 lower than those estimated for the Pukaki and Tekapo Glaciers although the value is within
585 the uncertainty window stated by Porter [1975]. The gradient of simulated ELA along the
586 eastern side of the range indicated that slightly greater cooling was needed to simulate
587 glaciers that extended to the LGM moraines further north. ΔT of –4.5°C forced the advance
588 of the Rakaia Glacier to Double Hill at ~17 ka [Putnam et al., 2013] for which there is no
589 equivalent advance identified elsewhere in the Southern Alps. ΔT of –4.5°C also forced the
590 advance of the Pukaki Glacier to the Birch Hill moraines at ~13 ka [Putnam et al., 2010], and
591 a similar ΔT of –4.0°C forced the advance of the Franz Josef Glacier to the Waiho Loop
592 moraine at ~13 ka [Anderson and Mackintosh, 2006], although the climatic significance of
593 this advance is unclear [e.g. Tovar et al., 2008]. The cooling required for simulated glaciers
594 to reach LGM extents (ΔT = –6.5°C) is in agreement with the 4–7°C of cooling estimated
595 from LGM sea surface temperatures [Barrows et al., 2007; Bostock et al., 2013] and previous
596 regional glacier modeling [Golledge et al., 2012].

597

598 **4.2 The Rakaia Glacier**

599 Golledge et al. [2012] identified those glaciers with large overdeepenings farthest from their
600 accumulation areas, including the Rakaia, as those most challenging to model
601 reconstructions; to obtain a successful simulation of the Rakaia Glacier, temperature for the
602 entire Southern Alps icefield was –1.5°C to –2.0°C below that used for the LGM in other
603 eastern valleys [Golledge et al., 2012]. Such large uncertainties may be attributed to: (1)
604 catchment-scale meteorological variability forced by orography and therefore not represented
605 in regional climate data; (2) the limitations of the approximations used in glacier modeling
606 that do not completely conserve mass; or (3) the composition of the bed and its influence on
607 subglacial motion. Finer-resolution (200-m compared to 500-m grid spacing) simulations for

608 the Rakaia–Rangitata Glaciers [Rowan et al., 2013] reached an improved solution which

609 showed a greater degree of synchroneity between these glaciers, but still had a ΔT of –0.25°C

610 between the LGM extent of the Rakaia and that of both the neighboring Rangitata and

611 Ashburton Glaciers.

612

613 **4.3 Uncertainty due to the description of mass balance**

614 Many glacier models use empirically-derived accumulation and ablation rates measured in

615 the field or based on relevant degree-day factors to describe mass balance [Braithwaite, 1995;

616 Hock, 2003]. However, the assumption of a linear relationship between mass balance and

617 elevation may be inaccurate because steep topography modifies radiative energy fluxes and

618 the redistribution of snow by avalanching. We instead used a 2-D energy balance calculation

619 based on a monthly climatology derived from local AWS, and incorporated the effects of

620 topographic shading and avalanching into the calculation of mass balance [Plummer and

621 Phillips, 2003]. Using 2-D meteorological data allows the investigation of glacier sensitivity

622 to spatial variability in mass balance which is not considered in 1-D models. This paper uses

623 present-day 30-year mean meteorological data to capture the regional orographic trend in

624 precipitation distribution (Fig. 3). However, precipitation may exhibit a more complex

625 distribution than can be captured at the resolution of these gridded data (the NIWA data have

626 a grid spacing of ~4 km), and catchment-scale variations may be unaccounted for in regional

627 gridded precipitation data.

628

629 **4.4 Uncertainty due to interannual variability**

630 The position of a glacier terminus can fluctuate even in the absence of a change in climate.

631 Variations in mass balance forced by interannual climate variability can produce nested sets

632 of moraines. For a given advance, the outermost terminal moraine will therefore represent the

633 maximum rather than the mean glacier length [Anderson et al., 2014]. While the magnitude

634 of the most likely maximum fluctuation of the Rakaia Glacier was larger for the LGM than

635 for the Late Glacial (2.3 km compared to 1.3 km), the LGM fluctuations represent a smaller

636 percentage of the maximum glacier length (2.8% compared to 10%). Ice extents preserved

637 within 2.8% and 10% of the terminal moraine position for the LGM and Late Glacial

638 advances could therefore be explained by climate noise rather than climate change.

639

640 The standard deviation of annual precipitation amount ($\sigma_P = 0.9$ m per year) and summer

641 temperature ($\sigma_T = 0.8°C$) for the Southern Alps are comparable to those for other maritime

642   regions. For example σ_P = 1.0 m per year and σ_T = 0.8°C in the North Cascade Mountains,

643   USA [*Roe and O'Neal*, 2009]; σ_P = 0.9 m per year and σ_T = 0.7°C at Nigardsbreen in Norway

644   [Roe and Baker, submitted to the Journal of Glaciology], and an exceptional case with a

645   continental climate of σ_P = 0.22 m per year and σ_T = 1.3°C for the Colorado Front Range,

646   USA [Anderson et al., 2014]. The standard deviations of the length of the simulated Rakaia

647   Glacier (850 m and 1200 m) are large compared to those for other simulated glaciers forced

648   by interannual variability. For example, 180 m for the Rhonegletscher and 360 m for

649   Nigardsbreen [Reichert et al., 2002]; 415 m for the North Cascades, USA [*Roe and O'Neal*,

650   2009]; and 280–960 m for the Colorado Front Range [Anderson et al., 2014]. The large

651   standard deviations of the length of the Rakaia Glacier are the result of the large variability in

652   annual precipitation and mean melt-season temperature typical of maritime climates.

653

654   Our results imply that interannual climate variability may significantly affect advances that

655   are less than or as extensive as the Late Glacial Rakaia advance. Future modeling studies

656   should consider that smaller advances could be explained by climate noise without

657   implicating changes in climate. The magnitude of the standard deviations of the length of the

658   Late Glacial and LGM advances is amplified by the low bed slopes of these glaciers (~5%

659   slope for the Late Glacial and ~0.5% slope for the LGM Rakaia), as large glacier area relative

660   to the ablation zone slopes enhances length fluctuations due to climate noise [Roe and

661   *O'Neal*, 2009; Eqn. 11]. The uncertainties in the 2-D mass balance model inputs overwhelm

662   the effect of interannual climate variability on paleoclimate estimates for the Late Glacial and

663   LGM Rakaia advances. However, the nearest present-day glacier to the Late Glacial extent is

664   ~5 km upvalley, and interannual variability could have forced an advance accounting for up

665   to 26% of the total Late Glacial advance from the present-day glacier position. Furthermore,

666   current New Zealand meteorological records do not cover a long enough time span to

667   confidently test for conditional probability in the climate system [e.g. Burke and Roe, 2013],

668   which could greatly enhance the magnitude of interannual climate varibility-forced advances.

669

670   **4.5 Uncertainty due to the choice of model domain and grid spacing**

671   Adjacent glaciers may not reach steady state synchronously [Oerlemans et al., 1998]. The

672   simulations presented in our study were performed over the model domain composed of three

673   major catchments—the Pukaki, Tekapo and Rakaia–Rangitata (Fig. 1). We compared steady-

674   state ice volumes between two domains with the same grid spacing (200 m) but with different

675   grid extents. We applied identical simulations for present-day and Double Hill scenarios to

676 the entire model domain and to sub-domains representing just the Pukaki or Rakaia–

677 Rangitata catchments. Under present-day conditions ($\Delta T = 0°C$), there was no difference in

678 simulated glacier extents between domains. However, under the Double Hill scenario ($\Delta T = -$

679 4.5°C), the Rakaia Glacier was 0.81 km (2%) shorter and the Pukaki Glacier was 4.3 km

680 (6%) longer for the small domain simulations compared to those for the large domain (Fig.

681 7).

682

683 The conservation of ice mass was monitored in each simulation, as spurious output may

684 result when the calculated flux from a cell is greater than the ice mass in that cell [Jarosch et

685 al., 2013]. Mass conservation was improved for simulations with a smaller grid spacing (100

686 m compared to 200 m). However, there is a tension in the choice of grid spacing, as halving

687 this value quadruples the number of cells in the model domain and computation time

688 increases exponentially. We compared identical simulations with different grid spacings to

689 calculate the change in simulated length of the Rakaia Glacier. As the 100-m grid spacing

690 simulations were more computationally expensive to run, we only tested small values for ΔT.

691 Compared to the 200-m grid spacing; with $\Delta T = -1.0°C$ a 100-m grid spacing produced a

692 glacier 0.7 km (6%) shorter; and with $\Delta T = -2.25°C$ equivalent to the Reischek Knob 1

693 advance [Putnam et al., 2013a] a 100-m grid spacing produced a glacier 0.8 km (3%) shorter.

694 Conservation of mass was improved by an order of magnitude using a 100-m grid spacing,

695 from an integrated mass balance of between –5% and –1% to less than –0.5%. Simulations

696 with a smaller grid spacing resulted in a systematically slightly less-extensive glacier than

697 those with a coarser grid.

698

699 The variations in glacier length resulting from the use of different model domains and either a

700 200-m or 100-m grid spacing is equivalent to $\Delta T < 0.1°C$ which is smaller than the

701 uncertainty ascribed to the application of the glacier model (equivalent to change in glacier

702 length of 13% and $\Delta T = 0.25°C$). Within the range of values tested, the size of the model

703 domain and the grid spacing are not considered to represent a significant source of

704 uncertainty and the larger grid spacing is preferable to allow less expensive computation.

705 Conservation of mass improved as ΔT increased; integrated glacier balance was no greater

706 than –2% when ΔT exceeded –5°C, indicating that this uncertainty is relatively small for

707 larger glaciers.

708

709 **5. Conclusions**

710 The sensitivity of glaciers in the Southern Alps of New Zealand to mean annual air

711 temperature, precipitation amount, precipitation distribution, interannual climate variability

712 and seasonality was tested using two glacier models. Variations in mass balance and glacier

713 length were governed primarily by differences in mean annual air temperature and the

714 distribution of precipitation. The variations in glacier lengths resulting from the choice of

715 precipitation data were equivalent to those resulting from a difference in temperature of

716 0.5°C. However, if precipitation were calculated as a function of elevation a larger

717 uncertainty in the simulated glacier length would be produced, equivalent to a difference in

718 temperature of 2°C. Interannual climate variability and seasonality added relatively minor

719 uncertainties to our paleoclimate estimates, although the effect of interannual variability was

720 important for advances comparable to or smaller than that during the Late Glacial extent at

721 ~11 ka.

722

723 Within a plausible range of precipitation variability for the Southern Alps (80–140% of

724 present-day regional values), the Last Glacial Maximum (LGM) occurred with a difference in

725 temperature from present-day values of –8.25°C to –6.0°C, or up to –5.5°C if the present-day

726 worldwide maximum precipitation amount (185%) was used. This LGM paleoclimate

727 envelope captures the possible climatic variability indicated by the glacial geological record,

728 based on the assumption that precipitation during the glacial was in the present-day

729 worldwide range, and includes the ±0.25°C uncertainty assigned to the choice of precipitation

730 data. To account for the total uncertainty in the LGM simulations resulting from other

731 climatological variables (the critical temperature for rain–snow partitioning, interannual

732 variability and seasonality) we assign an additional uncertainty of ±0.25°C.

733

734 Glacier models require a spatial representation of precipitation, the distribution of which is

735 difficult to quantify even under present-day conditions. Rainfall data are collected at a

736 relatively large number of weather stations in the Southern Alps and some snowfall

737 measurements are also made, but estimates of orographic precipitation distributions still

738 contain substantial uncertainties. Our results demonstrate the importance of quantifying

739 sensitivity to a range of precipitation distributions when making glacier model

740 reconstructions, and of considering the uncertainty resulting from how precipitation

741 distribution may have varied from present-day values during the last glacial. Future glacier

742 modeling studies should test a range of plausible precipitation distributions to quantify these

743 uncertainties rather than relying on empirical relationships between precipitation amount and

744 elevation. Without these data describing precipitation distribution, we expect the resolution of

745 our Southern Alps glacier model to be ±0.5°C. However, by testing a range of precipitation

746 distributions we can estimate the paleoclimate envelope represented by a particular set of

747 moraines, and can resolve past differences in temperature greater than ±0.25°C from the late

748 Quaternary moraine record.

749

763

764 **Table captions**

765 Table 1. 2-D glacier model parameter values and variables used in the simulations described

766 in this paper.

767

768 Table 2. Precipitation data and the change in simulated ELA resulting from the use of

769 different precipitation distributions under present-day ($\Delta T = 0$°C) and Double Hill ($\Delta T = -$

770 4.5°C) scenarios.

771

772 Table 3. Comparison of the change in length of the simulated Pukaki and Rakaia Glaciers

773 resulting from experiments testing temperature, precipitation amount and distribution under

774 present-day ($\Delta T = 0$°C) and Double Hill ($\Delta T = -4.5$°C) scenarios. The Rakaia Glacier length

775 is zero in some present-day simulations as no ice is present at the headwall of this catchment.

776 In Experiments 1 and 2, the precipitation data used are the NIWA grids. NB: The solution for

777 the simulation where precipitation was doubled under the Double Hill scenario did not reach

778 steady state due to an unrealistically high mass balance—these results are not presented (*).

779

780 **Figure captions**

781 Figure 1. Location of the study area in the Southern Alps of New Zealand (A), showing the

782 model domain used in the 2-D experiments (red shaded area), the maximum glaciated extent

783 during the LGM (white shading) and location of the precipitation transect shown in Fig. 3

784 (blue line). Glacier volumes simulated under (B) the present-day ($\Delta T = 0$°C) scenario (the

785 baseline model), and (C) the Double Hill scenario ($\Delta T = -4.5$°C), overlain on a shaded relief

786 map of the model domain. Catchment boundaries (solid red lines; dashed sections indicate

787 interpretation over areas of low relief) and the flowlines used to measure length of the Pukaki

788 and Rakaia Glaciers (solid green lines) are shown.

789

790 Figure 2. Parameter sets (ΔT and P) for the advance of the Rakaia Glacier (RG) to Prospect

791 Hill (green dotted shading) and the LGM limit (purple diagonal-hatched shading), compared

792 to results for the advance of the Franz Josef Glacier (FJG) to the Waiho Loop moraine (grey

793 horizontal lined shading) from Anderson and Mackintosh [2006], and for the Lateglacial

794 advance of the Irishman Glacier (red cross-hatched shading) from Doughty et al. [2013]. The

795 present-day interannual precipitation variability at the FJG (blue shading) and the present-day

796 worldwide precipitation maximum (blue dashed line) are shown [Henderson and Thompson,

797 1999]. As any change in LGM precipitation amount is unlikely to have exceeded the present-

798 day worldwide maximum, the change in climate for these advances probably lies within the

799 blue shaded area.

800

801 Figure 3. Precipitation amount along a 1-D transect through the center of the model domain

802 orientated at 130°. (A) Annual 30-year mean precipitation data collected from rain gauges

803 within 50 km of the transect (filled points), 30-year monthly mean precipitation amount for

804 (B) April and (C) October, showing 1-D precipitation profiles for central Southern Alps

805 [Griffiths and McSaveney, 1983; Wratt et al., 1996; Henderson and Thompson, 1999; Wratt

806 et al., 2000; Tait et al., 2006], CliFlo rain gauge data fitted with a least-squares cubic spline

807 that preserves shape, and a linear function linking rainfall to elevation derived from the

808 CliFlo data. The mean topographic profile of the model domain (green shading), the position

809 of the main drainage divide (red dashed line) and the 30-km region downwind of the main

810 drainage divide where glacier sensitivity to precipitation amount is greatest (pink shading)

are shown. Note that some points in (A) occur above the mean topographic profile as the stations are located above the mean elevation along this transect. Also note the different units for elevation (m) and precipitation (mm) on the y-axis of B and C.

Figure 4. Change in length of Rakaia and Rangitata Glaciers with ΔT. Inset shows the relative change in glacier length as a percentage of the total LGM glacier length. The location of Prospect Hill and Double Hill are noted.

Figure 5. Glaciers simulated using five different precipitation distributions under (A) the present-day (ΔT = 0°C) and (B) the Double Hill (ΔT = –4.5°C) scenarios. The blue shading shows glaciers simulated using the Henderson and Thompson [1999] profile. The glaciers simulated using the NIWA precipitation data [Tait et al., 2006] (orange lines), CliFlo (black lines), Wratt et al. [1996] (purple lines), and Griffiths and McSaveney [1983] (green lines) profiles for the same climate scenarios are also shown. Inset to (A) shows the glaciated area at the main drainage divide in detail.

Figure 6. Change in ELA with ΔT for the Rakaia–Rangitata domain (black points) with one standard deviation uncertainty compared to ELAs reconstructed for the eastern Southern Alps [Porter, 1975], and variations in ELA resulting from an estimated maximum LGM seasonality (red bar).

Figure 7. Double Hill (ΔT = –4.5°C) simulations for the (A) Pukaki and (B) Rakaia Glaciers. In each case, ice extent and terminus position are shown for the simulation using the model domain including the Pukaki, Tekapo and Rakaia–Rangitata catchments (solid black line). Terminus position for the simulations applied to a smaller domain covering just the catchment in question (dashed black lines) are also shown.

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