**Modelling sea level data from China and Malay-Thailand to estimate Holocene
ice-volume equivalent sea level change.**

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26 **Abstract.**

27

28 This study presents a new model of Holocene ice-volume equivalent sea level (ESL),
29 extending a previously published global ice sheet model (Bassett et al., 2005), which
30 was unconstrained from 10 kyr BP to present. This new model was developed by
31 comparing relative sea level (RSL) predictions from a glacial isostatic adjustment
32 (GIA) model to a suite of Holocene sea level index points from China and Malay-
33 Thailand. Three consistent data-model misfits were found using the Bassett et al.,
34 (2005) model: an over-prediction in the height of maximum sea level, the timing of
35 this maximum, and the temporal variation of sea level from the time of the highstand
36 to present.

37

38 The data-model misfits were examined for a large suite of ESL scenarios and a range
39 of earth model parameters to determine an optimum model of Holocene ESL. This
40 model is characterised by a slowdown in melting at ~7 kyr BP, associated with the
41 final deglaciation of the Laurentide Ice Sheet, followed by a continued rise in ESL
42 until ~1kyr BP of ~5.8 m associated with melting from the Antarctic Ice Sheet. It was
43 not possible to identify an earth viscosity model that provided good fits for both
44 regions; with the China data preferring viscosity values in the upper mantle of less
45 than 1.5×10^{20} Pa s and the Malay-Thailand data preferring greater values. We suggest
46 that this inference of a very weak upper mantle for the China data originates from the
47 nearby subduction zone and Hainan Plume. The low viscosity values may also
48 account for the lack of a well-defined highstand at the China sites.

49

50

51 **1: Introduction**

52

53 Relative sea level (RSL) across the globe is spatially and temporally variable, with a
54 range of processes driving this variation (Milne et al., 2009). On millennial time
55 scales during the Quaternary, the global pattern of sea-level change was strongly
56 controlled by the growth and retreat of ice sheets (Clark et al., 1978; Lambeck et al.,
57 2014). In areas of glaciation – the “near-field” - the RSL signal was dominated by the
58 solid earth response to the deglaciation/glaciation of the overlying ice sheets. In these
59 regions, RSL data is commonly used to constrain the regional ice sheet history and
60 earth rheology (Bradley et al., 2011; Lambeck et al., 2010; Lecavalier et al., 2014;
61 Tarasov et al., 2012). In contrast, at locations distant from the major glaciation centres
62 – the “far-field” - the RSL signal was dominated by the rate and magnitude of
63 meltwater input into the oceans and, as such, data from these regions are commonly
64 used to constrain the chronology of changes in the volume of grounded ice, often
65 referred to as the ice-volume equivalent sea level (henceforth abbreviated to ESL)
66 (Fleming et al., 1998; Lambeck and Nakada, 1990; Milne and Peros, 2013; Peltier,
67 2002)

68

69 ESL is defined as the spatially uniform change in sea level due to the mass exchange
70 between the ice sheets and the oceans (Farrell and Clark, 1976) and is important to
71 constrain for a number of reasons. It can be used as a constraint for global ice sheet
72 modelling reconstructions, independent from the ice volume signal derived from $\delta^{18}O$
73 marine sediments (Lisiecki and Raymo, 2005). Extracting the global ice volume
74 signal from such records requires knowledge of both changes in temperature and
75 salinity (Duplessy et al., 2002; Waelbroeck et al., 2002) and so ESL reconstructions

76 can help to calibrate these data as a proxy for total ice volume. Finally, the rate and
77 magnitude of meltwater influx during the mid-to-late Holocene is a useful baseline for
78 interpreting the rate of 20th century global mean sea level change (Gehrels and
79 Woodworth, 2013).

80

81 There have been a number of previous studies that estimated the rate and magnitude
82 of ESL change. For example, the ICE-5G (Peltier, 2004; Peltier and Fairbanks, 2006;
83 Toscano et al., 2011) and the ICE-6G (Argus et al., 2014) models both infer a marked
84 slowdown in the rate of global melting at 8 kyr BP, followed by a rise in ESL
85 between 7 and 4 kyr BP of ~ 4 m in ICE-5G and ~ 2 m in ICE-6G, with no melt
86 subsequent to 4 kyr BP. From the most recent study by Lambeck and colleagues
87 (Lambeck et al., 2014), a markedly different conclusion was drawn about the
88 temporal pattern of Holocene melting. In this study, a 5 m rise between 7- 2 kyr BP
89 was inferred, with no melt following 2 kyr BP. This result is compatible with an
90 earlier study by Fleming et al., (1998), which concluded that ESL rose by 3-5 m
91 between 7-2 kyr BP. As it is widely agreed that the main source of meltwater for the
92 ESL rise during the mid-to-late Holocene was the Antarctic Ice Sheet (AIS) (Bentley
93 et al., 2014), the differences between these two ESL models has implications for the
94 timing and deglaciation history of the AIS.

95

96 The aim of this study is to develop a new model of ESL change over the Holocene
97 using far-field sea level data from China and Malay-Thailand. A revised global ice
98 model will be produced, which extends a previously published model (Bassett et al.,
99 2005) that was tuned to fit a suite of far-field sea level data from Last Glacial
100 maximum (LGM) to 10 kyr BP (and so unconstrained over the Holocene) to examine

101 the volume partitioning of global ice loss during Meltwater Pulse 1A (MWP-1A). In
102 the following, we first review the RSL database from China and Malay-Thailand to
103 identify the key trends and patterns across the South East Asia region (Section 2). We
104 then go on to estimate ESL models for each region (China and Malay-Thailand)
105 (Section 3.3) as well as a single model that provides a good fit to the combined data
106 set (Section 3.4).

107

108 **2: Relative Sea Level Data**

109

110 The RSL database used in this study contains 314 sea level index points (SLIPs) and
111 was compiled by Horton et al., (2005) and Zong (2004). These databases of SLIPs
112 were compiled following the methodology of International Geoscience Programme
113 (IGCP) projects 61, 200, 495 and 588 (e.g., Horton et al., 2009; Preuss, 1979;
114 Shennan, 1982; Shennan et al., 2015; Switzer et al., 2012; Tooley, 1982; Van de
115 Plassche, 1982; Van de Plassche, 1986; Van de Plassche et al., 2014). A SLIP is a
116 discrete reconstruction of the unique position of RSL in time and space (van de
117 Plassche et al., 2014). Where a suite of SLIPs exists for a locality or region, they
118 describe changes in RSL through time. Horton et al., (2005) and Zong (2004)
119 validated all index points in their databases by insuring that each sea-level indicator
120 had: (1) a location; (2) a calibrated age; (3) an elevation at which the indicator was
121 sampled; and (4) a known relationship between the indicator and a contemporaneous
122 tidal level (termed the indicative meaning; van de Plassche, 1986; Shennan et al.,
123 2015). The SLIPs are sub-divided into 16 areas for the purpose of producing RSL
124 curves. The locations of all SLIPs and the 16 areas (the red triangle indicates the

125 mean position of all SLIPs in each area) are shown on Figure 1. Figure 2 shows the
 126 SLIPs at the 9 localities in China and Figure 3 for the 7 localities in Malay-Thailand.
 127
 128 The study region is located within the Eurasian Plate and is surrounded by active
 129 subduction zones. Along the eastern coast of China, the Pacific and Philippine sub-
 130 plate subducts below the Eurasian plate. However the coast of China, residing a
 131 significant distance (up to 900 km) away from the active plate boundary and with its
 132 extensive continental shelf and infrequent seismic activity, is considered to be
 133 geological stable (see Zong (2004) and references there in). The Malay-Thailand
 134 Peninsula sits within the geologically stable region called Sundaland and is bounded
 135 to the south by the active Sunda arc, where the Indo-Australian plate subducts below
 136 the Eurasian Plate. Tjia, (1996) using a range of shoreline indicators estimated that the
 137 Malay-Peninsula was stable over the Holocene, with rates of vertical crustal motion of
 138 less than 0.1 mm yr^{-1} . Therefore no correction for vertical tectonic motion was applied
 139 to the SLIPs. However, there might be some site-specific localised tectonic influences
 140 that were not corrected for in the development of the SLIPs and could be the source of
 141 data-model misfits.
 142
 143 The database consists of a range of sea-level indicators, including oyster reefs,
 144 beachrocks and organic silts (from China) and corals and fossilised mangrove swamps
 145 (from Malay-Thailand) with the associated errors described in greater detail by Zong,
 146 (2004) and Horton et al., (2005). Deltaic areas (i.e Yangtze River Delta (YRD), Han
 147 River Delta (HRD) and Chao Phraya delta (CP)) can be problematic for sea-level
 148 studies due to long term subsidence associated with sediment loading (Ferrier et al.,
 149 2015; Ivins et al., 2007; Wolstencroft et al., 2014) and sediment consolidation (Brain,

2015; Horton and Shennan, 2009; Teatini et al., 2011). The former can produce a RSL
fall of several metres over millennial time scales while the latter can result in a larger
contribution within shorter timescales. For example, the SLIP's at the Chao Phraya
delta (CP) may have been affected by sediment compaction, which would result in the
SLIP's plotting at lower elevations than originally deposited. At the site Yangtze
River Delta (YRD), the SLIPs were taken from the edge of the delta plain, into the
lowlands region, which is considered to be only moderately affected by the
subsidence processes mentioned above (Zong, 2004). This is not the case for the other
deltaic localities. To determine if the inclusion of SLIPs from these areas strongly
affected the GIA model parameters inferred below, we considered a subset of the
China data set that does not include Yangtze River Delta (YRD) or Han River Delta
(HRD). This subset also excluded the data from Fujian Province (FP) that is located
near a plate margin and thus possibly affected by local tectonic motion. This China
dataset includes the sites Bohai Bay (BB), Jingasu Province (JP), East Gaungdong
(EG), Hainan Province (HP) and West Guangdong (WG) and will be referred to as the
'China-reduced data set'.

Spatial variability in the observed sea-level change is evident along the coast of China
(Fig. 2). At the northern (Bohai Bay (BB), Jingasu Province (JP), Yangtze River Delta
(YRD)) and southern (West Guangdong (WG), Hainan Province (HP)) sites, sea level
rises from around -20 m at 9 kyr BP, with a marked slowdown between 7-8 kyr BP, as
it approaches present day levels. There is little evidence for sea levels above present
day values at these sites. For example, Hainan Province (HP) displays sea levels close
to 0 m over the last 6.5 kyr BP. In contrast, the central sites (Fujian Province (FP),
Han River Delta (HRD), East Gaungdong (EG), Pearl River Delta (PRD)) exhibit a

175 gradual slowdown in the rate of sea-level rise, approaching present day level later,
176 between 5-6 kyr BP. A small highstand of 1-1.5 m can be identified at some of these
177 central sites. The possible origin of this spatial variability will be discussed further
178 below.

179

180 At most of the Malay-Thailand data sites, there are gaps in the sea-level record
181 between 6-3 kyr BP. For example see data at sites Chao Phraya (CP) and East Coast
182 of Thailand (ET) (Fig. 3). This limits the constraint offered by these data for
183 determining an optimal Holocene melt history. In general, the data indicate a marked
184 slowdown between 7-8 kyr BP, with sea level rising steadily to form a highstand of ~
185 2- 4 m between 6-4 kyr BP. This is followed by a steady fall, reaching present day
186 levels by ~1 kyr BP. There is some spatial variability in the observed highstand
187 height, with a maximum sea level of ~ 4 m at sites along the western coast within the
188 Strait of Malacca (Keelang (KG), Strait of Malacca (MAL)) and site Chao Phraya
189 (CP), the most northerly site in Thailand. This contrasts with the lower sea level
190 highstand of ~1.5 m at the more central sites. This can be explained by the influence
191 of hydro-isostasy across the region, which will be discussed in greater detail below.

192

193 Between 5-2 kyr BP, the observed data at Tiomin Island (TI) and Phuket (PH) (Fig. 3)
194 imply a very low rate of sea-level change, which is not seen in the other data records
195 or captured by the starting model. We therefore hypothesise that this observed trend is
196 related to either local processes or errors in the interpretation of the sea-level
197 indicators used.

198

199

200 **3: Sea Level Modelling**

201

202 *3.1: Model description*

203

204 Sea level predictions were generated at each of the 16 sites (Figs. 4a and 4b) using a
205 glacial isostatic adjustment (GIA) model of RSL change. Such a model typically
206 comprises three sub-components: a global model of the Late Pleistocene ice sheet
207 history (Bassett et al., 2005; Peltier, 2004); an earth model that reproduces the solid
208 earth deformation due to changes in the surface mass redistribution between the ice
209 sheets and the oceans (Peltier, 1974); a model of sea-level change to calculate the
210 redistribution of ocean mass for a specified ice history and earth model (Farrell and
211 Clark, 1976). Note that the sea-level model adopted here includes perturbations to the
212 rotation vector (Milne and Mitrovica, 1998; Mitrovica et al., 2001; Mitrovica et al.,
213 2005) as well as time-dependant shoreline migration and an accurate treatment of sea-
214 level change in areas characterised by ablating marine based ice (Kendall et al., 2005;
215 Mitrovica and Milne, 2003).

216 Given that the primary aim of this study is to estimate the ESL changes from
217 the field data described in Section 2, it is worth defining some underlying concepts
218 related to ESL. At any given location in the oceans, the change in RSL
219 ($\Delta S_{rsl}(\theta, \psi, t)$) at a given time t with respect to values at present can be written as
220 follows:

221

$$222 \Delta S_{rsl}(\theta, \psi, t) = \Delta S_{esl}(t) + \Delta S_I(\theta, \psi, t) + \Delta S_T(\theta, \psi, t) + \Delta S_o(\theta, \psi, t) \quad (1)$$

223

Where $\Delta S_{est}(t)$ is the ESL component, $\Delta S_I(\theta, \psi, t)$ is the total isostatic effect associated with the GIA process and includes the effects of ice (glacio-isostatic) and water (hydro-isostatic) loads and earth rotation. The term $\Delta S_T(\theta, \psi, t)$ represents the relative sea level contribution from tectonic processes and the term $\Delta S_o(\theta, \psi, t)$ represents the contributions from ‘other’ processes such as sediment compaction, sediment loading, and changes in tidal amplitudes through time. The possible contributions from tectonics and ‘other’ processes are not modelled here but they are considered as possible reasons for poor data-model fits in certain areas. The ESL component is defined as:

$$\Delta S_{est}(t) = -\frac{\rho_{ice}}{\rho_w} \int_0^t \left(\frac{\dot{V}_{ice}(t)}{A_o(t)} \right) dt \quad (2)$$

where ρ_{ice} and ρ_w are the densities of ice and water, respectively; $\dot{V}_{ice}$ is the time derivative of global grounded ice volume relative to the present value; A_o is the surface area of the ocean. In what follows, our estimates of ESL are determined via equation (2); however, we note that these estimates include marine-based ice below sea level and so are an overestimate of the true ESL value. This overestimate is Earth model dependent but is around 10 m for the period LGM to present, but less than 0.6 m over the last 7 kyr BP (Fig. 5).

The starting ice model, and hence starting ESL model (referred to as $\Delta S_{est}^{sm}(t)$, where the notation *sm* refers to the starting model), was taken from Bassett et al., (2005, 2007) (model is shown on Fig. 5). This model was tuned to fit a small subset of far-field records spanning (completely or partially) the period LGM to ~10 kyr BP (Bassett et al., 2005) and near-field Antarctic RSL records (Bassett et al., 2007). It is

249 characterised by a relatively rapid rise in the early Holocene up to 6 kyr BP,
250 associated with the final stages of global melting. Major melting ceased at 6 kyr BP in
251 this model, with a small amount melting (0.5 m) between 6-5 kyr. In all the revised
252 ESL scenarios discussed in this study, the total global ESL change between LGM and
253 present day was not altered from that defined in the starting ESL model (compare the
254 solid grey and black lines on Fig. 5a), only the timing and relative contribution of the
255 two main ice sheets (Laurentide Ice Sheet (LIS) and AIS) over the past 10 kyr (Figure
256 5). This ensures that the high quality model fits to the far-field data between LGM
257 and the late glacial in Bassett et al. (2005) are preserved. In the starting model, the
258 rapid rise in ESL associated with MWP-1A (~14.5 kyr BP) is primarily driven by the
259 melting of the AIS, with the deglaciation of this ice sheet ending at 7 kyr BP. From ~
260 13 kyr BP to present, the total ESL is strongly controlled by the ESL rise from the
261 NHIS (compare the solid and dotted grey lines Fig. 5b), specifically the rapid rise
262 during the early Holocene up to 5 kyr BP is driven by the final retreat of the LIS.

263

264 The Earth model is a spherically symmetric, self-gravitating Maxwell body (Peltier,
265 1974). The radial elastic and density structure were taken from a seismic model
266 (Dziewonski and Anderson, 1981). The viscous structure is crudely parameterized
267 into three layers: a high viscosity (10^{43} Pas) outer shell to simulate an elastic
268 lithosphere, an upper mantle region of uniform viscosity extending from beneath the
269 model lithosphere to the 660 km seismic discontinuity, and a lower mantle region
270 extending from this depth to the core-mantle boundary. The thickness of the
271 lithosphere and the viscosity within the upper and lower mantle regions are free
272 parameters in the modelling. The starting ESL model was combined with an earth
273 model characterised by a 96 km lithosphere, and upper and lower mantle viscosity

274 values of 5×10^{20} Pas and 1×10^{22} Pa s, respectively, which will be referred to as the
275 reference earth model. The viscosity parameters in this model fall approximately
276 within the middle of the range of values commonly inferred (and considered below).

277

278 *3.2 Preliminary data-model comparison*

279

280 Sea-level predictions based on the starting ESL model (solid black line on Figs. 4a
281 and 4b) display a similar trend across the study region, with a slow steady rise up to a
282 highstand of several metres at 6 kyr BP (which marks the end of main stage of
283 melting within the starting ESL model), followed by a subsequent steady fall towards
284 zero. Two isostatic processes dominate this fall: hydro-isostasy (or continental
285 levering) and ocean syphoning (Clark et al., 1978; Mitrovica and Milne, 2002). Ocean
286 syphoning generates a fall in global mean sea level due to the influence of GIA on the
287 relative motion between the sea surface and sea floor, which increases during
288 deglaciations due, largely, to peripheral bulge subsidence associated with glacial
289 isostasy in mid-to-high latitudes and so-called “continental levering” associated with
290 hydro-isostasy around continental margins (Mitrovica and Milne, 2002). The latter
291 generates subtle sea level gradients perpendicular to the shoreline (e.g. see RSL
292 contours in Fig. 1).

293

294 Three consistent data-model misfits are clearly evident: (1) an overprediction in the
295 height of the maximum sea level (or highstand) attained, by up to 7 m across China
296 (for example, at site Yangetese River Delta (YRD) Fig. 4a) and up to 4 m across
297 Malay-Thailand (Fig. 4b); (2) the timing of the highstand and (3) temporal variation
298 of sea-level from the time of the highstand to present.

299

300 The approach to present day sea-level prior to the development of the highstand is
301 captured well by the starting model at the Northern and Southern China sites and at all
302 the Malay-Thailand data sites. For example, at both Strait of Malacca (MAL) and
303 Chao Phraya (CP) (see Fig. 4b) and Jingasu Province (JP) (see Fig. 4a), the
304 predictions lie within error limits of the SLIPs pre-dating ~ 7 kyr. The starting model
305 does not, however, capture the timing of the slowdown between 5-6 kyr BP evident at
306 the sites in central China.

307

308 The height of the predicted highstand is more spatially variable across Malay-
309 Thailand, by up to 4 m, than across China, where it remains between 6 - 7 m at most
310 sites. The SLIPs at the China sites generally rule out sea level across the region
311 reaching values greater than 1.5 m. Due to the noted gaps in the sea level record at
312 many sites across Malay-Thailand, constraints on the timing and maximum height of
313 the sea level high stand are limited.

314

315 Previous investigations into the spatial and temporal variability of sea level at far-
316 field sites (Lambeck and Nakada, 1990) have identified that the main processes
317 affecting the variability are hydro-isostatic loading (continental levering), ocean
318 syphoning and/or local tectonism. Examining the spatial pattern of predicted sea level
319 using the starting ESL model along the coastline of China at 6 kyr BP (contoured on
320 Fig. 1), there is no notable difference between the Northern and Central sites. This
321 implies that the observed spatial variability is not due to GIA but other processes that
322 are active in areas with a large sediment supply such as sediment compaction/loading

323 and growth faulting (e.g. Han River Delta, (HRD)) or tectonic activity in areas
324 proximal to a plate margin (Fujian Province (FP)).

325

326 In contrast, across Malay-Thailand, there is significant spatial variability in the pattern
327 of the predicted sea level (see Fig. 1) due to GIA processes. For example, hydro-
328 isostasy results in sea level gradients that are generally perpendicular to the coastline
329 (see Fig. 1) with an amplified sea-level fall (due to land uplift) further inland. As
330 such, the highest sea levels are found at sites closer to larger landmasses, such as at
331 Keelang (KG) where it reaches ~ 8 m, compared to only ~ 4 m at Phuket (PH). This
332 implies that the spatial variability in the observed highstand is due to GIA driven
333 processes.

334

335 It is well established that predicted sea levels at far-field sites are sensitive to both the
336 choice of input ice model (particularly the ESL component) and earth model.

337 Therefore it is important to assess if the described above misfits can be resolved
338 through reasonable adjustments to the viscous structure in the earth model.

339 Predictions were therefore generated using seven earth viscosity models (see Fig. 4a
340 and 4b). The minimum and maximum values are: 71, 120 km for lithosphere
341 thickness, $0.05\text{--}1 \times 10^{21}$ Pa s for upper mantle viscosity, and $1\text{--}50 \times 10^{21}$ Pa s for lower
342 mantle viscosity.

343

344 From Figure 4a and 4b it is apparent that the majority of the data-model misfit cannot
345 be resolved with these large changes to the earth model parameters. Adjusting the
346 input earth model parameters primarily alters the amplitude of the predicted
347 highstand, with the shape and timing of the predicted sea level curve remaining

348 relatively unaltered. Over the range of lithosphere thicknesses investigated (71- 120
349 km), there is a difference in the predicted RSL of up to 1 m. Reducing the viscosity of
350 the upper or lower mantle (dotted and dot-dashed lines, respectively, on Fig. 4a and
351 Fig. 4b) reduced the height of the highstand by up to 3 m around Malay-Thailand and
352 2 m around China, but still did not fully resolve the significant misfit between 7-5 kyr
353 BP. The choice of mantle viscosity primarily controls the rate at which the earth will
354 respond to changes in the overlying loading; hence with a weaker mantle viscosity,
355 the response rate is increased and hence the predicted sea level lower. Using a weak
356 lower mantle viscosity (10^{21} Pa s) at some sites around Malay-Thailand (Paracub
357 (PA), Keelang (KG) and Tiomin Island (TI)) did remove the misfits, with predictions
358 lying within the error limits of the observed data. However, this relatively weak value
359 for the lower mantle viscosity did not improve the fit at all sites across the region
360 (such as Chao Phraya (CP)) or the China sites. We conclude that varying the
361 viscosity structure alone cannot effectively reduce the primary data-model misfits
362 noted above. Therefore, in the following section, the starting ESL model will be
363 revised as a more effective means to improving the data-model fits.

364

365 *3.3 Determining first-iteration ESL models from the China and Malay-*

366 *Thailand data*

367

368 This section focuses on determining two ESL models which when combined with the
369 reference earth model (a 96 km lithosphere, and upper and lower mantle viscosity
370 values of 5×10^{20} Pa s and 1×10^{22} Pa s respectively) produce an improved fit to the
371 two regional data sets.

372 The first step, in reference to equation (1), was to subtract from the predicted sea level
373 $\Delta S_{rsl}(\theta, \psi, t)$ (generated using the starting ESL model and the reference earth model)
374 at each SLIP location the model component $\Delta S_{esl}^{sm}(t)$ (the ESL from the starting ESL
375 model) thus producing the quantity $\Delta S_I(\theta, \psi, t)$ (we are assuming that $\Delta S_T(\theta, \psi, t)$
376 and $\Delta S_o(\theta, \psi, t)$ are negligible).

377 Second, a large set of ~ 300 ESL scenarios ($\Delta S_{esl}^i(t)$) were then generated and added
378 to $\Delta S_I(\theta, \psi, t)$ to produce a suite of new predicted RSL, $\Delta S_{rsl}(\theta, \psi, t)$. Thirdly, the
379 χ^2 misfit was calculated in order to assess how certain changes to the ESL function
380 affected the data-model fits and identify the optimum ESL model. Finally, the starting
381 ESL model was adjusted to generate this optimum ESL model.

382

383 Note that this series of steps (and that also discussed in Section 3.4) does not take into
384 account changes in the ocean-loading signal associated with the difference between
385 the starting model eustatic function $\Delta S_{esl}^{sm}(t)$ and that of each of the revised ESL
386 scenarios $\Delta S_{esl}^i(t)$. This contribution can be significant (> 1 m) in cases where the
387 revised ESL scenario is very different from the starting model used in generating
388 $\Delta S_I(\theta, \psi, t)$ (e.g. compare solid grey and dashed black lines in Fig. 6). We do not
389 take this into account when generating this suite of 300 ESL scenarios. However,
390 once a good fitting ESL model is identified for each region, the starting ice model was
391 manually altered (by revising the deglaciation history of the input LIS and AIS) to
392 recreate the optimal ESL functions.

393

394 To generate the ~ 300 ESL scenarios, the partitioning of melt in the period 10 kyr to
395 present in the starting ESL model ($\Delta S_{esl}^{sm}(t)$) was systematically altered to address the
396 key misfits identified between the observed data and the starting ESL model. Four

example ESL scenarios referred to in the following text are shown on Figure 6. Key aspects investigated were: (1) timing of the melt slowdown in the mid-Holocene, ranging from 6 kyr (*melt1*, which is similar to the starting model) to 7 kyr (*melt2* and *melt3a/3b*); (2) the duration and variation of melt following this initial slowdown (abrupt in *melt1* or gradual in *melt2* and *melt3*); and (3) timing for the end of global melting, 3 kyr BP in *melt2*; 2 kyr BP in *melt1*; 1 kyr BP in *melt3a/3b*. These revisions were implemented by considering different partitions of rate of melt rise between the early, mid and late Holocene (i.e from 10 kyr to present). For example, *melt3a* and *melt3b* illustrate two end-member scenarios where the partitioning of melt following the initial slowdown is varied; the greater the amount of melt added to the ocean prior to the slowdown the higher the resultant highstand. Therefore the *melt3a* scenario, with 8 m of melt after the slowdown at 7 kyr BP will produce a smaller (or no) highstand compared to *melt3b*.

Analysing the χ^2 results for the entire suite of 300 ESL scenarios, two optimum ESL models were chosen which produced a good fit to each data set, and will be referred to as CHINA1 (for the China-complete data set) and MT1 (for Malay-Thailand data set) (Fig. 6). The LIS and AIS components of the starting ESL model were manually altered to produce the CHINA1 and MT1 models such that the change in ocean-loading signal will be fully accounted for in the revised RSL predictions.

Both of these models are characterised by an initial slowdown in the ESL rise at 7 kyr BP, 1 kyr earlier than in the starting model, followed by a gradual reduction in the rate of ESL rise until the end of melting at 1 kyr BP. This chronology of melting is comparable to the ‘melt3’ models (dashed black and grey lines, Fig. 6). However, these two models differ in the amount of melt partitioned before and after the 7 kyr

BP slowdown, with an extra 2.4 m of melt prior to 7 kyr BP preferred by the Malay-
Thailand data compared to the China data set (compare MT1 and CHINA1 on Fig. 6).

We note that the choice of earth model will influence the specific ESL model,
particularly the volume of melt partitioned before and after the initial timing of the
melt slowdown. This parameter trade-off is considered in the next section in which
we determine optimal ESL histories and earth model parameters for each region
independently as well as a final optimal ESL model for the combined data set.

3.4: Determining an optimal ESL model

In order to determine an optimal earth and ESL model first for each region and then
for the combined data set, we follow a procedure similar to that discussed in Section
3.3. First, a suite of 13 melt models was produced based around the chronology of
melting defined by the *melt3* model (see Fig. 6) which provided the optimum fits for
the reference earth model. The amount of melt partitioned before and following the
initial slowdown at 7 kyr BP was varied to produce these 13 melt models, a subset of
which is shown in Figure 7. These models are identified based on their value at the
time of initial slowdown (7 kyr BP), which ranges from -4.7 m to -8 m. The CHINA1
and MT1 ice models produced in the previous section were used to generate the
 $\Delta S_I(\theta, \psi, t)$ values used in these calculations, with the ice model most closely
matching each curve shown in Figure 6 being used. Second, for each melt model,
RSL predictions were generated for 72 earth models: a lithosphere thickness of 96 km
with viscosity values for the upper and lower mantle regions sampled from the ranges
0.05- 1×10^{21} Pa s and 1- 50×10^{21} Pa s, respectively.

447

448 From this procedure, optimum ESL models and earth model parameters were
449 determined for the complete and reduced China and Malay-Thailand data sets by
450 examining the data-model fits (χ^2 values) as function of earth model parameters and
451 adopted ESL curves (Fig. 7). These optimum ESL models will be referred to as
452 CHINA2, CHINA3 for the complete and reduced China data sets, respectively, and
453 MT2 for the Malay-Thailand data. They are shown in Figure 7.

454

455 There are a number of points to conclude from Figure 7, which highlights the
456 different sensitivity of the two datasets to choice of earth model parameters. The
457 China data rule out lower mantle viscosities greater than $\sim 5 \times 10^{21}$ Pa s regardless of
458 the ESL model considered (Fig. 8a, d & g). Despite the differences in the sensitivity
459 to the choice of upper mantle viscosity between the two China datasets, there is
460 evident overlap in the 95% confidence region and thus a range of possible earth
461 model parameters. Therefore, when determining the region specific earth model,
462 parameters were chosen which produced an optimum fit to both datasets, although
463 this may not directly correspond to the minimum χ^2 value. These parameters were a
464 lithosphere thickness of 96 km combined with an upper and lower mantle viscosity of
465 8×10^{19} Pa s and 1×10^{21} Pa s respectively. We note that, when comparing the 95%
466 confidence limits, it may appear that a better fit is produced with the CHINA2 model
467 for the reduced data set compared to the CHINA3 model, (c.f. Fig. 8b and 7e), the χ^2
468 value using this region-specific earth model is higher for the CHINA2 model (3.51)
469 compared to the CHINA2 (3.36).

470

471 The Malay-Thailand data rule out upper mantle viscosities less than 1.5×10^{20} Pa s for
 472 the CHINA2 and CHINA3 ESL models (Fig. 7c & f), in which there is relatively little
 473 melt prior to 7 kyr BP. However for the MT2 model, with greater melt after 7 kyr BP,
 474 the data rule out earth models with both weak upper and lower mantle viscosities (i.e.
 475 the optimum parameters for the China data). The optimum earth model parameters
 476 chosen for the MT2 model are a lithosphere thickness of 96 km and upper and lower
 477 mantle viscosities of 2×10^{20} Pa s and 5×10^{21} Pa s respectively (Fig. 7i).
 478
 479 Comparing the ESL histories for these optimum models (Fig. 9) it is again apparent
 480 that the China data prefer a greater amount of melt following the slowdown than the
 481 Malay-Thailand data. This is consistent with the results from Section 3.3. However, it
 482 is noted that the new optimum models differ to those inferred earlier based on the
 483 reference earth model (i.e. CHINA1 and MT1). Specifically, ESL at 7 kyr BP is -9 m
 484 in CHINA1 and -6.6 m in MT1, compared to -7 m in CHINA2 and -5.5 m in MT2.
 485 These differences reflect the earth model parameters used and so highlight the trade-
 486 off between ice and earth model parameters.
 487
 488 Using the region specific optimum earth and ESL (CHINA2, CHINA3 or MT2)
 489 models, a minimum χ^2 value was defined for each data set from which a 95%
 490 confidence upper limit was estimated (using *F*-test tables): 6.56 for China-complete,
 491 4.88 for China-reduced and 25.2 for Malay-Thailand data (see Fig. 8). These values
 492 were then used to identify a range of possible melt models that satisfy the combined
 493 data set within the defined confidence limit (see Fig. 10). It is evident from Figure 10
 494 that the upper 95% confidence limit of ESL at 7 kyr BP is constrained by the Malay-
 495 Thailand data and the lower limit by the China data set. The range of possible models

496 (at the specified confidence limit) is indicated by the shaded region on Figure 10a for
497 the China-complete data set (ESL between -5.7 m and -6.6 m at 7 kyr BP) and the
498 shaded region on Figure. 10b for the China-reduced data set (ESL between -5.3 and -
499 6.6 m at 7 kyr BP). An ESL of -6.0 m at 7 kyr BP represents the middle model within
500 this range (represented by the solid grey line on Fig. 7).

501

502 This model defines the ESL history that is optimised to fit the combined China and
503 Malay-Thailand data sets. The LIS and AIS components within the CHINA1 ice
504 model were revised further to recreate this final model (shown in Fig. 5 and Fig. 9).
505 (We note that this model has been referred to in previous publications as EUST3
506 (Milne and Peros, 2013; Woodroffe et al., 2015). As Fig. 5a illustrates the total LGM
507 – present day ESL contribution and the timing and nature of the pre-10 kyr BP ESL
508 rise from the NHIS and AIS remains unaltered from the starting model. However, in
509 the final model, the timing for the main LIS retreat, and rise in ESL from the NHIS is
510 shifted 1 kyr earlier than in the starting model, to 7 kyr BP (compare the dotted grey
511 and black lines on Fig. 5b). It is again noted that the total contribution over the last 10
512 kyr from the NHIS is the same in both models (27.7 m). Most of the change in the
513 post 10 kyr BP ESL rise in the final model was generated by a revision to the AIS,
514 where instead of a sharp rise of 7.5 m between 10 – 7 kyr BP (as in the starting
515 model), there is now a gradual rise in ESL between 10 and 1 kyr BP (compare the
516 dashed black and grey lines on Fig. 5b). As mentioned earlier, when fully accounting
517 for marine-based ice in the ESL estimate, this will reduce this ESL rise during the last
518 7 kyr BP by 0.6 m.

519

520 In summary the final model is characterised by: (1) an initial slowdown in the rate of
521 ESL rise at 7 kyr BP, associated with the final deglaciation of the Laurentide ice
522 sheet, (2) continued decelerations in the rate of ESL from 5 kyr BP until cessation of
523 melting at 1 kyr BP, (3) a ESL rise of 1.71 m from 5 kyr BP to present associated with
524 the deglaciation of the AIS.

525
526 There is a slight increase in the χ^2 minimum computed using our final model for the
527 China data set compared to the CHINA2 and CHINA3 models (see Fig. 8a, 7e and
528 Fig. 11a and 10b). We note that the final model provides a tighter bound on the range
529 of possible earth model values, ruling out upper and lower mantle viscosities greater
530 than $\sim 1.5 \times 10^{20}$ Pa s and $\sim 2 \times 10^{21}$ Pa s, respectively. Unlike the results based on the
531 CHINA2 and CHINA3 models, the minimum χ^2 value for both the complete and
532 reduced china data sets is the same, an upper mantle and lower mantle viscosity of $5 \times$
533 10^{19} Pa s and 1×10^{21} Pa s respectively. For the Malay-Thailand data (Fig. 11c), there
534 is also little change in the optimum region of earth model parameters between the
535 final model and the MT2 model (Fig. 8i). The optimum parameters are an upper and
536 lower mantle viscosity of 2×10^{20} Pa s and 8×10^{21} Pa s respectively. Again, it is not
537 possible to define an earth model that will fit within the 95% confidence region for
538 both data sets. Possible reasons for this difference in earth model parameters will be
539 discussed in the following section.

540
541 Combining our final optimised ice model with these two region-specific earth models,
542 RSL predictions were generated at each of the 16 locations indicated in Figure 1 and
543 compared to the observations from China (Fig. 12) and Malay-Thailand (Fig. 13).
544 Predictions using the starting model and the optimised region specific ESL models

545 (CHINA2 and MT2) are also shown. The CHINA2 and final model both display a
546 broadly similar trend in predicted RSL with a gradual rise towards present day levels
547 following a slowdown in the rate of rise at 7 kyr and 5 kyr BP. The previously
548 identified key misfits to the data using the starting ice model have been significantly
549 reduced with both models. The final model generates a very minor highstand (less
550 than 0.5 m) at the more northern sites at ~ 5 kyr BP which is within the error
551 uncertainties of the SLIPs. Prior to 5 kyr BP, the final model produces slightly higher
552 sea levels (less than 0.5 m) compared to CHINA2 but there is relatively little
553 difference between the two models from 5 kyr BP onwards. The remaining misfits at
554 the site Han River Delta (HRD) could be due to the impact of localised subsidence,
555 which would result in the data sitting below the predictions.

556

557 At Fujian Province (FP) and East Guangdong (EG), localised uplift may account for
558 the SLIPs sitting higher than the predictions using either CHINA2 or the final model
559 between 4 kyr BP and present . In a recent study Pedoja et al. (2008) estimated this
560 uplift rate to be 0.18 mm yr⁻¹ and 0.13 mm yr⁻¹ or a displacement of 0.7 m and 0.52 m,
561 respectively, over the past 4 kyr. At the two southern sites, Hainan Province (HP) and
562 West Guangdong (WG), which are considered to be most stable (Zong, 2004), the
563 predictions based on the final model capture the general trend of the SLIPs; predicting
564 the ~0 m sea level from 6.5 kyr BP at Hainan Province (HP) and the cluster of SLIPs
565 at 7 kyr BP at West Guangdong (WG).

566

567 For Malay-Thailand the final and MT2 models predict similar patterns (Fig. 12), with
568 a gradual rise towards present day sea level at ~ 7.5 kyr BP, rising to a maximum sea
569 level at ~ 5 kyr BP of between 1.5-4 m, followed by a gradual fall towards present.

570 However, the MT2 model predictions are higher at 5 kyr BP, by up to 1 m compared
571 to the final model which is to be expected given that the ESL is ~ 0.85 m higher in
572 MT2 compared to the final model at this time (see Fig. 9). Both models show a
573 significantly improved fit when compared to that of the starting ice model.
574
575 The most marked improvement is at sites Chao Phraya (CP) and Strait of Malacca
576 (MAL), where the final model captures both the timing of slowdown and the gradual
577 fall towards present from 5 kyr BP onwards. The predicted fall over the last 5 kyr
578 also captures the trend at sites Paracub (PA), Keelang (KG) and Tiomin Island (TI),
579 which supports the need for continued ice melting through the majority of the late
580 Holocene. There are a few SLIPs which are not captured: (1) the previously
581 mentioned uniform trend at Phuket (PH); (2) the cluster of points at sites Strait of
582 Malacca (MAL) and East Coast Thailand (ET) at 5 kyr and 8 kyr, respectively, which
583 display large vertical scatter. Clearly, it is not feasible to create a sea-level prediction,
584 which could capture these trends, and, as such, this misfit is not considered important
585 to the results of this modelling study.

586

587 **4: Discussion**

588

589 The final model is compared to the two most widely cited ESL models (Fig. 9)
590 (1) ICE5G (Peltier, 2004), and (2) the most recent in a series of models developed by
591 Kurt Lambeck and colleagues (Lambeck et al., 2014), which will be referred to as
592 LAM in the following discussion.

593

594 The Holocene ESL history in the ICE5G model is characterised by a rapid rise to 8
595 kyr BP, when there is a marked slowdown in the rate of melting. Following a
596 secondary minor slowdown at 7 kyr, there is a continued steady rise until melting
597 ceases at 4 kyr BP. The LAM ESL model is characterised by a steady rise in ESL
598 until a staged slowdown in the rate of melting at 7 kyr BP. From 5 kyr onwards ESL
599 rises at an average rate of 0.56 mm yr^{-1} until 2 kyr BP, which is similar to the rates
600 and timings of our final model (see Fig. 9), where the ESL rises at an average rate of
601 0.53 mm yr^{-1} between 5 and 2 kyr BP, and 0.12 mm yr^{-1} from 2 kyr to 1 kyr BP.
602 These rates are much lower than the estimated mean rate of global mean sea level rise
603 from 1901 to 2010, $1.5\text{-}1.9 \text{ mm yr}^{-1}$ (Rhein et al., 2013)
604
605 Our final model compared to the LAM and ICE5G models display broadly similar
606 trends over the last 10 kyr and constrain the Holocene ESL to within a relatively
607 narrow range. The two most apparent differences between the three models are the
608 timing of the initial slowdown in global melting (7 kyr BP versus 8 kyr BP) and the
609 timing of the cessation of melting (4 versus 2-1 kyr BP).
610
611 Taking ICE5G and our final model as two end member scenarios which span the
612 possible range for Holocene ESL, RSL predictions were generated with the ICE5G
613 model using VM2 earth model to explore the impact of the differences in the ESL
614 history on the RSL predictions (Fig. 12 and Fig. 13). We note that the ICE5G model
615 was calibrated using the VM2 earth model, which is a global earth model, rather than
616 region specific as used with the final ESL model.
617

618 It is apparent that the earlier timing for the initial slowdown in global melting in
619 ICE5G (8 kyr BP versus 7 kyr BP) produces significant misfit at the data sites with
620 pre-6.5 kyr BP SLIPs. For example, at the northern China sites (Bohai Bay (BB),
621 Jingasu Province (JP) and Yangetese River Delta (YRD)) and Strait of Malacca
622 (MAL) and Chao Phraya (CP) in the Malay-Thailand sites, the ICE5G predictions are
623 several metres above the observed SLIPs. As to be expected, given the stronger
624 mantle viscosity in the VM2 earth model, there is a significant highstand at 8 kyr BP
625 generated across all of the China sites with the ICE5G model which is several metres
626 higher and several kyr's earlier than the observed SLIP's. Therefore, the ICE5G-VM2
627 model does not capture the observed SLIP trend at the China sites. However, at the
628 Malay-Thailand sites, there is only a minor difference between the RSL predictions of
629 the final model and ICE5G from 4 kyr BP, which is due in part to the similarity in the
630 optimum earth structure and VM2. Therefore, it is not possible within the resolution
631 of the Malay-Thailand data to resolve the difference in the two ESL models over the
632 last 4 kyr BP.

633

634 Additional datasets from this region that were not included in this study also support a
635 later timing for the slowdown in global melting. Yu et al. (2002) investigated the
636 growth of corals at Leizhou Peninsula, South China Sea (close to the West
637 Guangdong (WG) data locality) and estimated the timing for the slowdown in sea-
638 level rise between 7.2-6.7 kyr BP. Radiocarbon dating of plant matter and shells from
639 sites around Hong Kong were combined with additional sedimentological and
640 paleontological evidence to infer that sea level initially reached present day levels
641 between 8-6 kyr BP (Yim, 1999).

642

643 It is useful to also compare the timing and chronology of Holocene ESL in the final
644 model to estimates made from other global far-field sites. In a recent study
645 (Woodroffe et al., 2015) where RSL predictions from the final model and ICE5G
646 were compared to sea level data from the Seychelles, it was concluded that the ESL
647 rise over the last 2 kyr was less than 2 m and that the observed RSL signal was more
648 compatible with the results of our final model. Milne and Peros (2013) investigated
649 sea level data from the circum-Caribbean and estimate a contribution of 3-4 m
650 between 7 and 3-2 kyr BP. Numerous studies have examined the extensive range of
651 Holocene sea level data from Australia. Some studies concluded that RSL was at least
652 1.5 m above present between 6.2 and 2.2 kyr BP (Sloss et al., 2007; Woodroffe,
653 2009), inferring ice continued to melt into the late Holocene, whereas other studies
654 suggest ESL rise ended closer to 4 kyr BP (Lewis et al., 2013). This earlier (6-4 kyr
655 BP) end to the rise in ESL is supported by sea level data from Japan (Yokoyama et
656 al., 2012), South America (Milne et al., 2005) and the Mediterranean (Mauz et al.,
657 2015). It is apparent, therefore, that there is no consensus on the timing for the end of
658 global melting, with significant variations depending on the location of the far-field
659 sea level site. This implies there is a need for more high resolution, spatially
660 distributed data sets and GIA models that can accommodate lateral variations in Earth
661 structure.

662

663 As the main source for the late Holocene ESL rise in the LAM and our final model is
664 the AIS, the inferred ESL history has implications for the Holocene deglaciation of
665 the AIS. There is near-field evidence to support this continued melting, with a number
666 of recent papers summarising the most up to date onshore and offshore
667 geomorphological evidence (Bentley et al., 2014; Hillenbrand et al., 2014; Larter et

al., 2014). As an in-depth review of the Holocene retreat of West AIS is beyond the scope of the study, the reader is referred to these papers and references therein. A number of recent studies have adopted near-field Antarctic constraints to produce improved ice model reconstructions for this region (Briggs et al., 2014; Golledge et al., 2014; Ivins et al., 2013; Mackintosh et al., 2011; Whitehouse et al., 2012). These model reconstructions give ESL values for the mid-to-late Holocene period that ranges from less than a metre to almost 5 m around 7 kyr BP (see Fig. 2 of Ivins et al. (2013) and related discussion). The results found here give an ESL value of 5.8 m since 7 kyr BP, which is reduced to 5.2 m when taking account of marine-based ice that would not contribute to a sea-level rise (Fig. 9b). This is compatible with models that include a relatively large amount of AIS melt during this period, such as the model of Ivins et al. (2013).

An interesting result from our modelling analysis is the distinctly different optimum earth viscosity structure identified for each region; specifically a very weak upper mantle, less than 1.5×10^{20} Pa s preferred by the China data. We suggest that this very weak upper mantle may be one of the reason for the distinct lack of a Holocene highstand at the China data sites, compared to the other sites across the Indo-Pacific region (Woodroffe and Horton, 2005).

As described above both regions are located close to active subduction zones, which are defined as dynamic plate margins where the relatively cold oceanic lithosphere subducts into the hotter, weaker oceanic mantle. This complex geodynamic process will affect the viscosity of the surrounding mantle. Numerical modelling and empirical flow law experiments both infer that the viscosity of the background mantle

693 around subduction zones is between 10^{20} and 10^{21} Pa s, with lower viscosities, 10^{18} -
 694 10^{19} Pa s in the mantle wedge, the back-arc region and surrounding the subducted slab
 695 (Billen, 2008; Currie and Hyndman, 2006; Hirth and Kohlstedt, 2013). We suggest,
 696 therefore, that the hydro-isostatic deformation affecting the China sites is significantly
 697 influenced by these weaker viscosity regions, which produce a faster response rate in
 698 the deformation of the solid earth and lowering predicted RSL in the mid-Holocene.
 699 This suggestion is supported by GIA modelling of RSL data in south west British
 700 Columbia in the vicinity of the Cascadia subduction zone, which concluded that the
 701 data required a weak upper mantle, between 3×10^{18} and 4×10^{19} Pa s (James et al.,
 702 2009). A second GIA modelling study of GPS data across the Southern Patagonian
 703 Icefield (Lange et al., 2014) also concluded the data required a weak upper mantle,
 704 between 1.6×10^{18} and 8×10^{18} Pas (depending on the Little Ice age loading history)
 705 due to the influence of local mantle upwelling from the eastern edge of the Pacific rim
 706 subduction zone. We note that a more recent study (Roy and Peltier, 2015) which
 707 compared RSL predictions using ICE-6G_C, combined with VM6 to a new RSL
 708 database for sites along the entire west coast of the USA (beyond the region in James
 709 et al., 2009) captured the regional observed Holocene RSL trend without the need for
 710 a low upper mantle viscosity. However, in this study, the predicted RSL fails to
 711 capture the observed RSL along the west coast of Vancouver, which it attributes to
 712 the tectonic impact of the closely located subduction zone.

713

714 Despite a similar general tectonic setting, the shallow earth structure below China
 715 appears to be more geodynamically complex than that below Malay-Thailand. One of
 716 the main methods to investigate the upper mantle structure around China is seismic
 717 tomography. Recent tomography studies have imaged the Pacific slab below China,

718 suggesting it flattens or deflects near the mantle transition zone which may generate
719 zones of weaker mantle (Li and van der Hilst, 2010). More extensive 3D tomography
720 studies have imaged low velocity anomalies (suggestive of high temperatures and
721 therefore low viscosity) close to the China study sites, compared to the high velocity
722 anomalies around Malay-Thailand (Lebedev and Nolet, 2003; Replumaz et al., 2004).
723 The most pronounced feature supporting a weaker region within the upper mantle is
724 the Hainan Plume, which was imaged below the southern coast and Hainan Island
725 (near to sites Hainan Province (HP) and West Guangdong (WG)), extending to depths
726 of up to 900 km (Huang, 2014; Lei et al., 2009). Clift et al. (2002) combined seismic
727 data, basin geometry and subsidence modelling to also estimate that a low viscosity
728 layer of between 10^{18} and 10^{19} Pa s underlies the southern coast of China. In a recent
729 review of the structure and properties of the mantle below China (Zang et al., 2008), a
730 degree of partial melt was identified at a depth of 410km with an estimated viscosity
731 of between 1.2×10^{18} Pa s and 3.6×10^{19} Pa s.

732

733 The range of constraints considered above support the low viscosities inferred from
734 the China data and significant lateral structure in the upper mantle, which is not
735 captured by the 1-D earth model adopted here. The influence of lateral earth structure
736 on Holocene RSL has been examined in previous studies (Spada et al., 2006; Wu and
737 van der Wal, 2003), with the results indicating differences of up to several metres at
738 far-field sites.

739

740 **5: Conclusions**

741

742 Using sea level data from China and Malay-Thailand a new model of late Holocene
743 ESL was developed. This model is characterised by: (1) an initial slowdown in the
744 rate of ESL rise at 7 kyr associated with the final deglaciation of the Laurentide ice
745 sheet, (2) continued deceleration in the rate of ESL rise from 5 kyr BP until cessation
746 of melting at 1 kyr BP (3) a total ESL rise since slowdown at 7 kyr BP of 5.8 m. It is
747 suggested, based on a variety of evidence, that the most likely source for the
748 continuing rise in ESL in the late Holocene is the West Antarctic Ice sheet (WAIS).
749

750 Our final model and two previously published ESL models, ICE5G (Peltier, 2004)
751 and LAM (Lambeck et al., 2014) display broadly similar trends over the last 10 kyr
752 and constrain the Holocene ESL history to a relatively narrow range. The two most
753 apparent differences between the three models are the timing of the initial slowdown
754 in global melting (~7 kyr BP, LAM and our final model versus 8 kyr BP in ICE5G)
755 and the timing of the cessation of melting (4 in ICE5G versus 2-1 kyr BP in the other
756 models). Our final model is more consistent with the overall characteristics of LAM,
757 where ESL rises steadily at an average rate of 0.56 mm yr^{-1} compared to 0.53 mm yr^{-1}
758 between 5 and 2 kyr BP. Comparison of RSL predictions using our optimum region
759 specific earth model and the ICE5G-VM2 model found significant misfits to the pre
760 6.5 kyr BP SLIPs; confirming that the data set prefer an ESL with a slowdown in
761 melting 7 kyr BP or later. We acknowledge that some of these discrepancies may be
762 attributed to the limitation associated with the application of a 1D Earth model in a
763 region where lateral variations are likely significant.

764

765 The data-model misfits indicate distinctly different optimum earth viscosity values for
766 the China and Malay-Thailand regions. Both data sets were relatively insensitive to

767 variations in lower mantle viscosity but prefer markedly different upper mantle
768 values. For China the best fit is achieved with a weak upper mantle, less than 1.5
769 $\times 10^{20}$ Pa s, compared to Malay-Thailand data, where a stronger mantle, greater than
770 2×10^{20} Pa s, is preferred. It is thought that this difference is related to distinct lateral
771 variations with the upper mantle structure below China, associated with the nearby
772 active subduction zones and the Hainan Plume.

773

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775

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782

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784 **References:**

785

786 Ai, Y.S., Zheng, T.Y., Xu, W.W., Li, Q., 2008. Small scale hot upwelling near the
787 North Yellow Sea of eastern China. *Geophysical Research Letters* 35, 5.

788 Argus, D.F., Peltier, W.R., Drummond, R., Moore, A.W., 2014. The Antarctica
789 component of postglacial rebound model ICE-6G_C (VM5a) based on GPS
790 positioning, exposure age dating of ice thicknesses, and relative sea level histories.
791 *Geophysical Journal International* 198, 537-563.

792 Bassett, S.E., Milne, G.A., Bentley, M.J., Huybrechts, P., 2007. Modelling Antarctic
793 sea-level data to explore the possibility of a dominant Antarctic contribution to
794 meltwater pulse 1A. *Quaternary Sci Rev* 26, 2113-2127.

- 795 Bassett, S.E., Milne, G.A., Mitrovica, J.X., Clark, P.U., 2005. Ice sheet and solid
796 earth influences on far-field sea-level histories. *Science* 309, 925-928.
- 797 Bentley, M.J., Cofaigh, C.O., Anderson, J.B., Conway, H., Davies, B., Graham,
798 A.G.C., Hillenbrand, C.D., Hodgson, D.A., Jamieson, S.S.R., Larter, R.D.,
799 Mackintosh, A., Smith, J.A., Verleyen, E., Ackert, R.P., Bart, P.J., Berg, S.,
800 Brunstein, D., Canals, M., Colhoun, E.A., Crosta, X., Dickens, W.A., Domack, E.,
801 Dowdeswell, J.A., Dunbar, R., Ehrmann, W., Evans, J., Favier, V., Fink, D., Fogwill,
802 C.J., Glasser, N.F., Gohl, K., Golledge, N.R., Goodwin, I., Gore, D.B., Greenwood,
803 S.L., Hall, B.L., Hall, K., Hedding, D.W., Hein, A.S., Hocking, E.P., Jakobsson, M.,
804 Johnson, J.S., Jomelli, V., Jones, R.S., Klages, J.P., Kristoffersen, Y., Kuhn, G.,
805 Leventer, A., Licht, K., Lilly, K., Lindow, J., Livingstone, S.J., Masse, G., McGlone,
806 M.S., McKay, R.M., Melles, M., Miura, H., Mulvaney, R., Nel, W., Nitsche, F.O.,
807 O'Brien, P.E., Post, A.L., Roberts, S.J., Saunders, K.M., Selkirk, P.M., Simms, A.R.,
808 Spiegel, C., Stollendorf, T.D., Sugden, D.E., van der Putten, N., van Ommen, T.,
809 Verfaillie, D., Vyverman, W., Wagner, B., White, D.A., Witus, A.E., Zwartz, D.,
810 Consortium, R., 2014. A community-based geological reconstruction of Antarctic Ice
811 Sheet deglaciation since the Last Glacial Maximum. *Quaternary Science Reviews*
812 100, 1-9.
- 813 Billen, M.I., 2008. Modeling the dynamics of subducting slabs, *Annual reviews of*
814 *Earth and Planetary Sciences*, pp. 325-356.
- 815 Bradley, S.L., Milne, G.A., Shennan, I., Edwards, R., 2011. An improved Glacial
816 Isostatic Adjustment model for the British Isles. *Journal of Quaternary Science* 26,
817 541-552.
- 818 Brain, M.J., 2015. Compaction, in: Shennan, I., Long, A.J., Horton, B.P. (Eds.),
819 *Handbook of Sea-Level Research*. Wiley, pp. 452-469.
- 820 Briggs, R.D., Pollard, D., Tarasov, L., 2014. A data-constrained large ensemble
821 analysis of Antarctic evolution since the Eemian. *Quaternary Science Reviews* 103,
822 91-115.
- 823 Clark, J.A., Farrell, W.E., Peltier, W.R., 1978. GLOBAL CHANGES IN POST-
824 GLACIAL SEA-LEVEL - NUMERICAL-CALCULATION. *Quaternary Research* 9,
825 265-287.
- 826 Clift, P., Lin, J., Barckhausen, U., 2002. Evidence of low flexural rigidity and low
827 viscosity lower continental crust during continental break-up in the South China Sea.
828 *Marine and Petroleum Geology* 19, 951-970.
- 829 Currie, C.A., Hyndman, R.D., 2006. The thermal structure of subduction zone back
830 arcs. *Journal of Geophysical Research-Solid Earth* 111, 22.
- 831 Duplessy, J.C., Labeyrie, L., Waelbroeck, C., 2002. Constraints on the ocean oxygen
832 isotopic enrichment between the Last Glacial Maximum and the Holocene:
833 Paleoceanographic implications. *Quaternary Science Reviews* 21, 315-330.
- 834 Dziewonski, A.M., Anderson, D.L., 1981. Preliminary Reference Earth Model. *Phys*
835 *Earth Planet In* 25, 297-356.

- 836 Farrell, W.E., Clark, J.A., 1976. POSTGLACIAL SEA-LEVEL. *Geophysical Journal*
837 *of the Royal Astronomical Society* 46, 647-667.
- 838 Ferrier, K.L., Mitrovica, J.X., Giosan, L., Clift, P.D., 2015. Sea-level responses to
839 erosion and deposition of sediment in the Indus River basin and the Arabian Sea.
840 *Earth and Planetary Science Letters* 416, 12-20.
- 841 Fleming, K., Johnston, P., Zwart, D., Yokoyama, Y., Lambeck, K., Chappell, J.,
842 1998. Refining the eustatic sea-level curve since the Last Glacial Maximum using far-
843 and intermediate-field sites. *Earth and Planetary Science Letters* 163, 327-342.
- 844 Gehrels, W.R., Woodworth, P.L., 2013. When did modern rates of sea-level rise start?
845 *Global and Planetary Change* 100, 263-277.
- 846 Golledge, N.R., Menviel, L., Carter, L., Fogwill, C.J., England, M.H., Cortese, G.,
847 Levy, R.H., 2014. Antarctic contribution to meltwater pulse 1A from reduced
848 Southern Ocean overturning. *Nature Communications* 5, 10.
- 849 Hillenbrand, C.D., Bentley, M.J., Stoll, T.D., Hein, A.S., Kuhn, G., Graham,
850 A.G.C., Fogwill, C.J., Kristoffersen, Y., Smith, J.A., Anderson, J.B., Larter, R.D.,
851 Melles, M., Hodgson, D.A., Mulvaney, R., Sugden, D.E., 2014. Reconstruction of
852 changes in the Weddell Sea sector of the Antarctic Ice Sheet since the Last Glacial
853 Maximum. *Quaternary Science Reviews* 100, 111-136.
- 854 Hirth, G., Kohlstedt, D., 2013. Rheology of the Upper Mantle and the Mantle Wedge:
855 A View from the Experimentalists, Inside the Subduction Factory. *American*
856 *Geophysical Union*, pp. 83-105.
- 857 Horton, B.P., Gibbard, P.L., Milne, G.M., Morley, R.J., Purintavaragul, C., Stargardt,
858 J.M., 2005. Holocene sea levels and palaeoenvironments, Malay-Thai Peninsula,
859 southeast Asia. *Holocene* 15, 1199-1213.
- 860 Horton, B.P., Shennan, I., 2009. Compaction of Holocene strata and the implications
861 for relative sea-level change on the east coast of England. *Geology* 37, 1083-1086.
- 862 Huang, J.L., 2014. P- and S-wave tomography of the Hainan and surrounding regions:
863 Insight into the Hainan plume. *Tectonophysics* 633, 176-192.
- 864 Ivins, E.R., Dokka, R.K., Blom, R.G., 2007. Post-glacial sediment load and
865 subsidence in coastal Louisiana. *Geophysical Research Letters* 34, 5.
- 866 Ivins, E.R., James, T.S., Wahr, J., Schrama, E.J.O., Landerer, F.W., Simon, K.M.,
867 2013. Antarctic contribution to sea level rise observed by GRACE with improved
868 GIA correction. *Journal of Geophysical Research-Solid Earth* 118, 3126-3141.
- 869 James, T.S., Gowan, E.J., Wada, I., Wang, K.L., 2009. Viscosity of the asthenosphere
870 from glacial isostatic adjustment and subduction dynamics at the northern Cascadia
871 subduction zone, British Columbia, Canada. *Journal of Geophysical Research-Solid*
872 *Earth* 114.

873 Kendall, R.A., Mitrovica, J.X., Milne, G.A., 2005. On post-glacial sea level - II.
874 Numerical formulation and comparative results on spherically symmetric models.
875 *Geophys J Int* 161, 679-706.

876 Lambeck, K., Nakada, M., 1990. LATE PLEISTOCENE AND HOLOCENE SEA-
877 LEVEL CHANGE ALONG THE AUSTRALIAN COAST. *Global and Planetary*
878 *Change* 89, 143-176.

879 Lambeck, K., Purcell, A., Zhao, J., Svensson, N.O., 2010. The Scandinavian Ice
880 Sheet: from MIS 4 to the end of the Last Glacial Maximum. *Boreas* 39, 410-435.

881 Lambeck, K., Rouby, H., Purcell, A., Sun, Y.Y., Sambridge, M., 2014. Sea level and
882 global ice volumes from the Last Glacial Maximum to the Holocene. *Proceedings of*
883 *the National Academy of Sciences of the United States of America* 111, 15296-
884 15303.

885 Lange, H., Casassa, G., Ivins, E.R., Schroder, L., Fritsche, M., Richter, A., Groh, A.,
886 Dietrich, R., 2014. Observed crustal uplift near the Southern Patagonian Icefield
887 constrains improved viscoelastic Earth models. *Geophysical Research Letters* 41,
888 805-812.

889 Larter, R.D., Anderson, J.B., Graham, A.G.C., Gohl, K., Hillenbrand, C.D.,
890 Jakobsson, M., Johnson, J.S., Kuhn, G., Nitsche, F.O., Smith, J.A., Witus, A.E.,
891 Bentley, M.J., Dowdeswell, J.A., Ehrmann, W., Klages, J.P., Lindow, J., Cofaigh,
892 C.O., Spiegel, C., 2014. Reconstruction of changes in the Amundsen Sea and
893 Bellingshausen Sea sector of the West Antarctic Ice Sheet since the Last Glacial
894 Maximum. *Quaternary Science Reviews* 100, 55-86.

895 Lebedev, S., Nolet, G., 2003. Upper mantle beneath southeast Asia from S velocity
896 tomography. *Journal of Geophysical Research-Solid Earth* 108, 26.

897 Lecavalier, B.S., Milne, G.A., Simpson, M.J.R., Wake, L., Huybrechts, P., Tarasov,
898 L., Kjeldsen, K.K., Funder, S., Long, A.J., Woodroffe, S., Dyke, A.S., Larsen, N.K.,
899 2014. A model of Greenland ice sheet deglaciation constrained by observations of
900 relative sea level and ice extent. *Quaternary Science Reviews* 102, 54-84.

901 Lei, J.S., Zhao, D.P., Steinberger, B., Wu, B., Shen, F.L., Li, Z.X., 2009. New seismic
902 constraints on the upper mantle structure of the Hainan plume. *Physics of the Earth*
903 *and Planetary Interiors* 173, 33-50.

904 Lewis, S.E., Sloss, C.R., Murray-Wallace, C.V., Woodroffe, C.D., Smithers, S.G.,
905 2013. Post-glacial sea-level changes around the Australian margin: a review.
906 *Quaternary Science Reviews* 74, 115-138.

907 Li, C., van der Hilst, R.D., 2010. Structure of the upper mantle and transition zone
908 beneath Southeast Asia from traveltimes tomography. *Journal of Geophysical*
909 *Research-Solid Earth* 115, 19.

910 Lisiecki, L.E., Raymo, M.E., 2005. A Pliocene-Pleistocene stack of 57 globally
911 distributed benthic delta O-18 records. *Paleoceanography* 20, 17.

- 912 Mackintosh, A., Golledge, N., Domack, E., Dunbar, R., Leventer, A., White, D.,
913 Pollard, D., DeConto, R., Fink, D., Zwartz, D., Gore, D., Lavoie, C., 2011. Retreat of
914 the East Antarctic ice sheet during the last glacial termination. *Nature Geoscience* 4,
915 195-202.
- 916 Mauz, B., Ruggieri, G., Spada, G., 2015. Terminal Antarctic melting inferred from a
917 far-field coastal site. *Quaternary Science Reviews* 116, 122-132.
- 918 Milne, G.A., Gehrels, W.R., Hughes, C.W., Tamisiea, M.E., 2009. Identifying the
919 causes of sea-level change. *Nature Geoscience* 2, 471-478.
- 920 Milne, G.A., Long, A.J., Bassett, S.E., 2005. Modelling Holocene relative sea-level
921 observations from the Caribbean and South America. *Quaternary Science Reviews*
922 24, 1183-1202.
- 923 Milne, G.A., Mitrovica, J.X., 1998. Postglacial sea-level change on a rotating Earth.
924 *Geophys J Int* 133, 1-19.
- 925 Milne, G.A., Peros, M., 2013. Data-model comparison of Holocene sea-level change
926 in the circum-Caribbean region. *Global and Planetary Change* 107, 119-131.
- 927 Mitrovica, J.X., Milne, G.A., 2002. On the origin of late Holocene sea-level
928 highstands within equatorial ocean basins. *Quaternary Science Reviews* 21, 2179-
929 2190.
- 930 Mitrovica, J.X., Milne, G.A., 2003. On post-glacial sea level: I. General theory.
931 *Geophys J Int* 154, 253-267.
- 932 Mitrovica, J.X., Milne, G.A., Davis, J.L., 2001. Glacial isostatic adjustment on a
933 rotating earth. *Geophys J Int* 147, 562-578.
- 934 Mitrovica, J.X., Wahr, J., Matsuyama, I., Paulson, A., 2005. The rotational stability of
935 an ice-age earth. *Geophys J Int* 161, 491-506.
- 936 Pedoja, K., Shen, J.W., Kershaw, S., Tang, C., 2008. Coastal Quaternary
937 morphologies on the northern coast of the South China Sea, China, and their
938 implications for current tectonic models: A review and preliminary study. *Marine*
939 *Geology* 255, 103-117.
- 940 Peltier, W.R., 1974. IMPULSE RESPONSE OF A MAXWELL EARTH. *Reviews of*
941 *Geophysics* 12, 649-669.
- 942 Peltier, W.R., 2002. On eustatic sea level history: Last Glacial Maximum to
943 Holocene. *Quaternary Science Reviews* 21, 377-396.
- 944 Peltier, W.R., 2004. Global glacial isostasy and the surface of the ice-age earth: The
945 ice-5G (VM2) model and grace. *Annu Rev Earth Pl Sc* 32, 111-149.
- 946 Peltier, W.R., Fairbanks, R.G., 2006. Global glacial ice volume and Last Glacial
947 Maximum duration from an extended Barbados sea level record. *Quaternary Science*
948 *Reviews* 25, 3322-3337.

- 949 Preuss, H., 1979. Progress in computer evaluation of sea-level data within the IGCP
950 project no. 61, Proceedings of the 1978 international symposium of coastal evolution
951 in the Quaternary, Sao Paulo, Brazil.
- 952 Replumaz, A., Karason, H., van der Hilst, R.D., Besse, J., Tapponnier, P., 2004. 4-D
953 evolution of SE Asia's mantle from geological reconstructions and seismic
954 tomography. *Earth and Planetary Science Letters* 221, 103-115.
- 955 Rhein, M., Rintoul, S.R., Aoki, S., Campos, E., Chambers, D., Feely, R.A., Gulev, S.,
956 Johnson, G.C., Josey, S.A., Kostianoy, A., Mauritzen, C., Roemmich, D., Talley L.D.,
957 Wang, F., 2013. Observations: Ocean, in: Stocker, T.F., Qin, D., Plattner, G.-K.,
958 Tignor, M., Allen, S.K., Boschung, J., Nauels, A., Xia, Y., Bex, V., Midgley, P.M.
959 (Eds.), *Climate Change 2013: The Physical Science Basis. Contribution of Working*
960 *Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate*
961 *Change*. Cambridge University Press, Cambridge, United Kingdom and New York,
962 NY, USA.
- 963 Roy, K., Peltier, W.R., 2015. Glacial isostatic adjustment, relative sea level history
964 and mantle viscosity: reconciling relative sea level model predictions for the US East
965 coast with geological constraints. *Geophysical Journal International* 201, 1156-1181.
- 966 Shennan, I., 1982. Sea-level changes in the North Sea, in: Tooley, M.J., Shennan, I.
967 (Eds.), *Sea-Level Changes*. Basil Blackwell, Oxford, pp. 109-151.
- 968 Shennan, I., Long, A.J., B.P. Horton, 2015. *Handbook of Sea-Level Research*. John
969 Wiley & Sons
970 Chichester.
- 971 Sloss, C.R., Murray-Wallace, C.V., Jones, B.G., 2007. Holocene sea-level change on
972 the southeast coast of Australia: a review. *Holocene* 17, 999-1014.
- 973 Spada, G., Antonioli, A., Cianetti, S., Giunchi, C., 2006. Glacial isostatic adjustment
974 and relative sea-level changes: the role of lithospheric and upper mantle
975 heterogeneities in a 3-D spherical Earth. *Geophysical Journal International* 165, 692-
976 702.
- 977 Switzer, A.D., Sloss, C.R., Horton, B.P., Zong, Y.Q., 2012. Preparing for coastal
978 change Introduction. *Quaternary Science Reviews* 54, 1-3.
- 979 Tarasov, L., Dyke, A.S., Neal, R.M., Peltier, W.R., 2012. A data-calibrated
980 distribution of deglacial chronologies for the North American ice complex from
981 glaciological modeling. *Earth and Planetary Science Letters* 315, 30-40.
- 982 Teatini, P., Tosi, L., Strozzi, T., 2011. Quantitative evidence that compaction of
983 Holocene sediments drives the present land subsidence of the Po Delta, Italy. *Journal*
984 *of Geophysical Research-Solid Earth* 116, 10.
- 985 Tjia, H.D., 1996. Sea-level changes in the tectonically stable Malay-Thai Peninsula.
986 *Quaternary International* 31, 95-101.
- 987 Tooley, M.J., 1982. Sea-level changes in North England. *Proceedings of Geological*
988 *Association* 93.

989 Toscano, M.A., Peltier, W.R., Drummond, R., 2011. ICE-5G and ICE-6G models of
 990 postglacial relative sea-level history applied to the Holocene coral reef record of
 991 northeastern St Croix, U.S.V.I.: investigating the influence of rotational feedback on
 992 GIA processes at tropical latitudes. *Quaternary Science Reviews* 30, 3032-3042.

993 Van de Plassche, O., 1982. Sea-level change and water-level movements in
 994 the Netherlands during the Holocene. *Mededelingen Rijks Geologische Dienst* 36, 1-
 995 93.

996 Van de Plassche, O., 1986. Sea level research: a manual for the collection and the
 997 evaluation of data. Geobooks, Norwich.

998 Van De Plassche, O., Wright, A.J., Horton, B.P., Engelhart, S.E., Kemp, A.C.,
 999 Mallinson, D., Kopp, R.E., 2014. Estimating tectonic uplift of the Cape Fear Arch
 1000 (south-eastern United States) using reconstructions of Holocene relative sea level.
 1001 *Journal of Quaternary Science* 29, 749-759.

1002 Waelbroeck, C., Labeyrie, L., Michel, E., Duplessy, J.C., McManus, J.F., Lambeck,
 1003 K., Balbon, E., Labracherie, M., 2002. Sea-level and deep water temperature changes
 1004 derived from benthic foraminifera isotopic records. *Quaternary Science Reviews* 21,
 1005 295-305.

1006 Whitehouse, P.L., Bentley, M.J., Le Brocq, A.M., 2012. A deglacial model for
 1007 Antarctica: geological constraints and glaciological modelling as a basis for a new
 1008 model of Antarctic glacial isostatic adjustment. *Quaternary Science Reviews* 32, 1-24.

1009 Wolstencroft, M., Shen, Z.X., Toernqvist, T.E., Milne, G.A., Kulp, M., 2014.
 1010 Understanding subsidence in the Mississippi Delta region due to sediment, ice, and
 1011 ocean loading: Insights from geophysical modeling. *Journal of Geophysical Research-*
 1012 *Solid Earth* 119, 3838-3856.

1013 Woodroffe, S.A., 2009. Testing models of mid to late Holocene sea-level change,
 1014 North Queensland, Australia. *Quaternary Science Reviews* 28, 2474-2488.

1015 Woodroffe, S.A., Horton, B.P., 2005. Holocene sea-level changes in the Indo-Pacific.
 1016 *Journal of Asian Earth Sciences* 25, 29-43.

1017 Woodroffe, S.A., Long, A.J., Milne, G.A., Bryant, C.L., Thomas, A.L., 2015. New
 1018 constraints on late Holocene eustatic sea-level changes from Mahé, Seychelles.
 1019 *Quaternary Science Reviews* 115, 1-16.

1020 Wu, P., van der Wal, W., 2003. Postglacial sealevels on a spherical, self-gravitating
 1021 viscoelastic earth: effects of lateral viscosity variations in the upper mantle on the
 1022 inference of viscosity contrasts in the lower mantle. *Earth and Planetary Science*
 1023 *Letters* 211, 57-68.

1024 Yim, W.W.S., 1999. Radiocarbon dating and the reconstruction of late Quaternary
 1025 sea-level changes in Hong Kong. *Quaternary International* 55, 77-91.

1026 Yokoyama, Y., Okuno, J., Miyairi, Y., Obrochta, S., Demboya, N., Makino, Y.,
 1027 Kawahata, H., 2012. Holocene sea-level change and Antarctic melting history derived

1028 from geological observations and geophysical modeling along the Shimokita
1029 Peninsula, northern Japan. *Geophysical Research Letters* 39, 6.

1030 Yu, K.F., Liu, D.S., Shen, C.D., Zhao, J.X., Chen, T.G., Zhong, J.L., Zhao, H.T.,
1031 Song, C.J., 2002. High-frequency climatic oscillations recorded in a Holocene coral
1032 reef at Leizhou Peninsula, South China sea. *Science in China Series D-Earth Sciences*
1033 45, 1057-+.

1034 Zang, S.-x., Zhou, H.-l., Wei, R.-q., 2008. Progress in studies on the structure and
1035 physical properties of the mantle in China. *Acta Seismologica Sinica* 21, 526-534.

1036 Zong, Y.Q., 2004. Mid-holocene sea-level highstand along the southeast coast of
1037 China. *Quaternary International* 117, 55-67.
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 1045 Fig. 1: Location map of the data sites. The black circles indicate the location of each
 1046 SLIP and the red triangles the 16 localities for which RSL curves are generated (see
 1047 main text for details). The sites are: BB – Bohai Bay, JP - Jingasu Province, YRD –
 1048 Yangtze River Delta, FP – Fujian Province, HRD – Han River Delta, EG – East
 1049 Guangdong, PRD – Pearl River Delta, WG – West Guangdong, HP – Hainan

1050 Province, CP-Chao Phraya, PA- Paracub , PH- Phuket , ET- East coast Thailand, KG-
1051 Keelang, MAL- Strait of Malacca and TI-Tiomin Island. The black contours show the
1052 predicted relative sea level at 6 kyr BP using the starting model (see main text for
1053 details).

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1074 Fig. 2: Reconstructed SLIPs. Solid black circles mark each SLIP with 1-sigma
 1075 uncertainty in time and height indicated by the error bars

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Fig. 3: As Figure 2 but for the seven localities in Malay-Thailand.

1104 Fig. 4a: Reconstructed SLIPs and modelled RSL changes at the nine localities
 1105 considered in China. Solid black circles mark each SLIP with 1-sigma uncertainty in
 1106 time and height indicated by the error bars. Predictions generated using the starting
 1107 model (described in text) are shown by the solid black lines. Variations in the input
 1108 lithosphere thickness are shown by the dashed line: 71km (grey) and 120 km (black).
 1109 The dotted line shows changes in the upper mantle viscosity from 5×10^{19} Pa s (grey)
 1110 to 10^{22} Pa s (black). Changes in lower mantle viscosity are shown by the dot-dashed
 1111 line: 10^{21} Pa s (grey) and 5×10^{22} Pa s (black).
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1119 Fig. 4b: As Figure 4a but for the seven localities in Malay-Thailand.
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1122 Fig. 5: Comparison of the Ice-volume equivalent sea level (ESL) from the Last glacial
 1123 maximum to present day (a) and over the Holocene period (b) from the starting model
 1124 $((\Delta S_{esl}^{sm}(t), \text{Bassett et al., 2005})$ (grey lines)) and from the final model (black lines).
 1125 The solid lines represents the total contribution from all ice sheets (note the solid
 1126 black and grey lines plot on top of each other pre 10 kyr BP on (a)); the dotted lines
 1127 are the contribution from the Northern Hemisphere ice sheets only (Laurentide ice
 1128 sheet, Greenland Ice sheet, Scandinavian Ice sheet and British-Irish Ice sheet)) and
 1129 the dashed lines from the Antarctic Ice sheet only. Additionally show is the total ESL
 1130 from the final model when corrected for marine-based ice (black dashed-dotted).

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Fig. 6: An example subset of the ESL scenarios considered. The solid black line represents the ESL of the starting model ($\Delta S_{esl}^{sm}(t)$). Four alternative ESL scenarios (referred to as $\Delta S_{esl}^i(t)$ in the main text) are also illustrated: *melt1* (grey; solid line), *melt2* (black; dotted line), *melt3a* (black; dashed line) and *melt3b* (grey; dashed line). As described in the main text, these scenarios explore different aspects of the ESL history. Also illustrated are the inferred ESL histories for the CHINA1 (red) and MT1 (blue) models (when assuming the standard earth model).

1162 Fig. 7: A selection of the 13 melt models adopted to investigate the sensitivity of the
 1163 data-model misfit to the amount of melt partitioned before and after 7 kyr BP. The
 1164 amount of melt prior to 7 kyr BP was systematically reduced to alter the resultant
 1165 eustatic sea level at this time ranging from a minimum of 4.7 m (dotted-dashed) to a
 1166 maximum of 8 m (dashed). The dotted and solid lines illustrate two intermediate
 1167 models, with the grey line highlighting the median model. All models are based on
 1168 the timing of the ‘melt3’ model (Fig. 6).

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Fig. 8: Contour plots of the χ^2 value as a function of upper and lower mantle viscosity (ν_{um} and ν_{lm} , respectively) for a lithosphere thickness of 96 km using the CHINA2 model (a,b and c), CHINA3 model (d,ef) and the MT2 model (g,h and i). Frames a,d, and g show the results for China-complete data set; b,e and h the China-reduced data set, and the c,f and i the Malay-Thailand data. The black line represents the χ^2 value below which the quality of fit is equivalent at 95% confidence and the white triangle the location of the minimum χ^2 value.

1191 Fig. 9: (a) Comparison of the ESL models produced in this analysis: CHINA2 (dashed
 1192 –dotted red line), CHINA3 (dashed-dotted green line), MT2 (dashed-dotted blue line)
 1193 and the final model (solid black line). (b) Comparison of the final model (solid black
 1194 line) to two published models: ICE-5G (black dotted; Peltier, 2004) and the model by
 1195 Lambeck et al., (2014), referred to as LAM in the main text (black dashed line). Also
 1196 shown is the ESL contribution from the final model when taking into account marine-
 1197 based ice (grey line)
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1216 Fig. 10: The χ^2 misfits for the range of melt models considered for the China data
1217 (black) and Thailand data (grey). Note that the melt model is defined based on the
1218 ESL value (magnitude only) at 7 kyr BP. Results are shown for the China-complete
1219 data set (a) and the China-reduced data set (b). The solid horizontal line marks the
1220 location of the 95% confidence limit, (6.56 for China-complete (a) and 4.88 for
1221 China-reduced (b) (black line) and 25.2 for Malay-Thailand (grey line)) – melt
1222 models, which result in values below these limits, are statistically equivalent to 95%
1223 confidence. The shaded region indicates the melt models, which satisfy this criterion
1224 for both data sets. Note that different earth models were used to generate predictions
1225 for each region (China and Malay-Thailand; see text for details).

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Fig. 11: Contour plots of the χ^2 value as a function of upper and lower mantle viscosity (ν_{um} and ν_{lm} , respectively) for a lithosphere thickness of 96 km using the final model. Results are shown for the (a) China-complete data set, (b) China-reduced data set and (c) Malay-Thailand data. The black line represents the χ^2 value below which the quality fit is equivalent at 95% confidence and the white triangle the location of the minimum χ^2 value.

Fig. 12: Comparison of predicted and observed RSL at each of the nine localities in China. Solid black circles mark the individual SLIPs. The solid black line shows predictions for the starting ice-earth model. Predictions for the CHINA2 (dashed black) and the final model (red) ESL models were generated for an optimum earth structure inferred from the χ^2 results presented in Figure 11: lithosphere thickness of 96 km, upper mantle viscosity of 5×10^{19} Pa s, lower mantle viscosity of 1×10^{21} Pa s. The solid grey line shows predictions for ICE-5G combined with the VM2 earth model.

Fig. 13: Comparison of the predicted and observed RSL at each of the seven localities in Malay-Thailand. Solid black circles mark the location of each SLIP. The solid line shows predictions for the starting ice-earth model. Predictions for MT2 (dashed-black) and the final model (red) were generated for an optimum earth structure inferred from the χ^2 analysis in Figure 11: lithosphere thickness of 96 km, upper mantle viscosity of 2×10^{20} Pa s, lower mantle viscosity of 8×10^{21} Pa s. The solid grey line shows predictions for ICE-5G combined with the VM2 earth model