

Bayesian reconstruction of sea-level and hydroclimates from coastal landform inversion: application to Santa Cruz (US) and Gulf of Corinth

Gino de Gelder^{1,2,3}, Navid Hedjazian³, Laurent Husson¹, Thomas Bodin⁴, Anne-Morwenn Pastier⁵, Yannick Boucharat¹, Kevin Pedoja⁶, Tubagus Solihuddin², Sri Y. Cahyarini²

¹ISTerre, IRD, CNRS, Université Grenoble-Alpes, Saint-Martin d'Hères, 38400, France

²Res. Group of Paleoclimate & Paleoenvironment, Res. Centr. for Climate and Atmosphere, Res. Org. of Earth Sciences and Maritime, National Research and Innovation Agency, Bandung, 40135, Indonesia

³ENS de Lyon, CNRS, LGL-TPE, Université Claude Bernard Lyon1, Villeurbanne, 69100, France

⁴Instituto de Ciencias del Mar (ICM) - CSIC, Barcelona, Spain

⁵GeoForschungsZentrum, Potsdam, 14473, Germany

⁶Université de Caen Normandie, Caen, 14000, France

Correspondence to: Gino de Gelder (gino.de-gelder@univ-grenoble-alpes.fr)

Abstract. Quantifying Quaternary sea-level changes and hydroclimatic conditions is an important challenge given their intricate relation with paleo-climate, ice-sheets and geodynamics. The world's coastlines provide an enormous geomorphologic archive, from which forward landscape evolution modelling studies have shown their potential to unravel paleo sea-levels, albeit at the cost of assumptions on the genesis of these landforms. We take a next step, by applying a Bayesian approach to jointly invert the geometries of multiple coastal terrace sequences to paleo sea- and lake level variations and extract past hydroclimatic conditions. Using a Markov chain Monte Carlo sampling method, we first test our approach on synthetic marine terrace profiles as proof of concept and then benchmark our model on an observed marine terrace sequence in Santa Cruz (US). We successfully reproduce observed sequence morphologies and simultaneously obtain probabilistic estimates for past sea-level variations, as well as for other model parameters such as uplift and erosion rates. When applied to the semi-isolated Gulf of Corinth (Greece), our method allows deciphering the geomorphic Rosetta stone at an unprecedented resolution, revealing the connectivity between the Lake/Gulf of Corinth and the open sea for different hydroclimatic conditions. Eustatic sea-level and changing sill depths drive marine and transitional phases during interglacial and interstadial periods, whereas wetter and drier hydroclimates respectively over- and under-fill Lake Corinth during interstadial and glacial periods.

Short summary. Marine terrace sequences - staircase-shaped coastal landforms - record sea-level changes, vertical motions and erosional processes that are difficult to untangle. To do so we developed a numerical inversion approach: using the observed landscape as input, we constrain the ensemble of parameter ranges that could have created this landscape. We

32 apply the model to marine terrace sequences in Santa Cruz (US) and Corinth (Greece) to reveal past sea/lake levels, uplift
33 rates and hydro-climates.

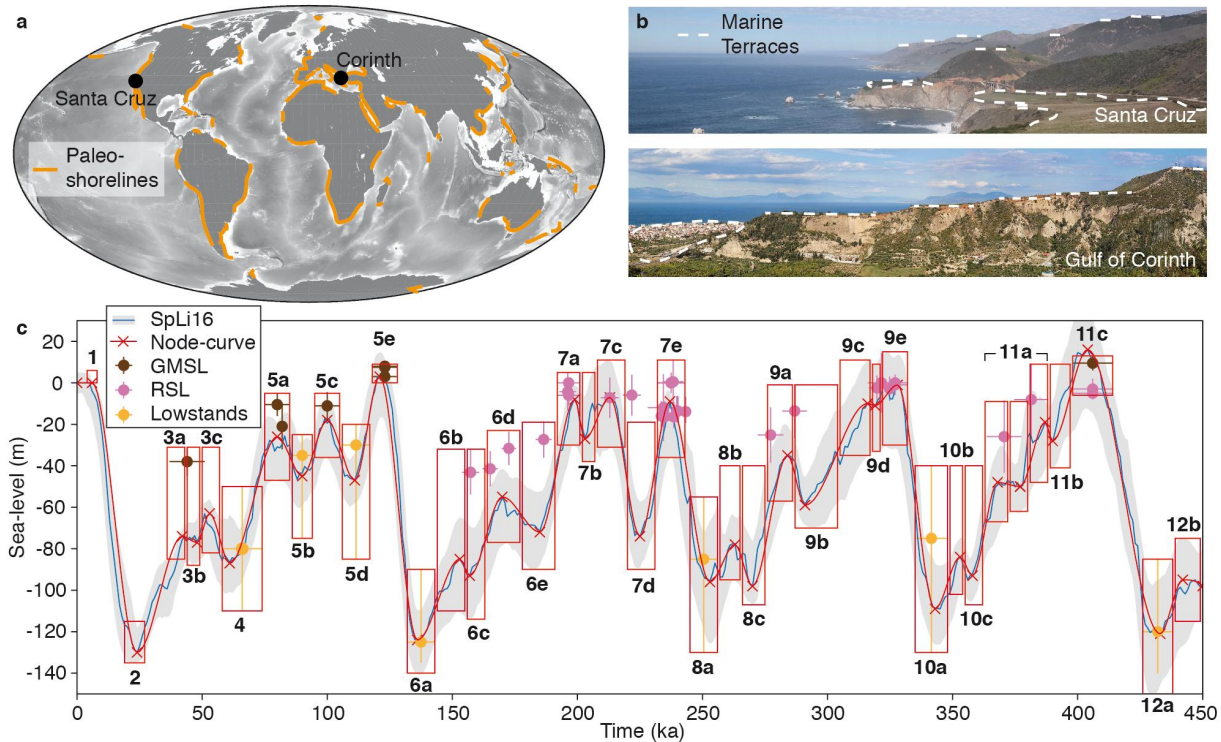
34 **1 Introduction**

35 Reconstructions of Quaternary sea-level variations provide crucial constraints on thresholds and feedbacks within climatic
36 and geodynamic systems that help understand how contemporary climate change may affect future sea level (Lambeck and
37 Chappell, 2001; Hay et al., 2014; Dutton et al., 2015; Shakun et al., 2015; Austermann et al., 2017). A key archive of past
38 sea-level is exposed within the geomorphology of most of the world's coastal areas in the form of paleo-shorelines
39 (Johnson and Libbey, 1997; Pedoja et al., 2011, 2014; Rovere et al., 2023; Fig. 1a), but it remains difficult to accurately
40 translate coastal observations and measurements into paleo-sea-level estimates, and to evaluate the uncertainties inherent to
41 these conversions. Major challenges include 1) dating of these landforms, as most paleo-shorelines are erosive in nature
42 (Pedoja et al., 2014) and absolute dating techniques themselves are complex and prone to large uncertainties (Strobl et al.,
43 2014; Hibbert et al., 2016; Ott et al., 2019), 2) observational bias, which are mostly restricted to the most recent glacial
44 cycle(s) and to periods where relative sea level was at similar elevations to present-day (Medina-Elizalde, 2013; Hibbert et
45 al., 2016), 3) absence of reciprocity between paleo-shorelines and sea-level stands, as not all highstands lead to paleo-
46 shorelines, and paleo-shorelines may have formed during one or many sea-level cycles (Guilcher, 1974; Malatesta et al.,
47 2021; Chauveau et al., 2023), and 4) separating the tectonic from the sea-level component within relative sea-level changes
48 (Pedoja et al., 2011).

49 Numerical models of landscape evolution started to overcome some of these limitations, by providing a means to
50 quantitatively interpret undated paleo-shorelines, incorporate full sea-level curves instead of highstands only, unravelling the
51 creation of paleo-shorelines formed over multiple glacial cycles, and considering multiple sea-level curves (e.g. Webster et
52 al., 2007; Jara-Muñoz et al., 2019; Leclerc and Feuillet, 2019; De Gelder et al., 2020; 2023). So far, such numerical models
53 have mainly been used for forward modelling approaches, where a number of proposed sea-level curves are used to predict
54 shorelines, which are then compared to actual observations. However, this only provides a limited way to explore the full
55 ensemble of possible sea-level histories and other model parameters like rock erosion rates or effective wave base depths,
56 which are difficult to estimate. It follows that uncertainties in sea-level estimates from marine terraces remain poorly known,
57 regardless of the method used, and in spite of uniformization attempts (Lorscheid and Rovere, 2019).

58 Semi-isolated marine basins, i.e. bodies of water that have been connected to the open sea in some intervals of their geologic
59 history, and little or disconnected from the sea in other intervals, develop in hydrodynamic settings for which it is
60 particularly complex to reconstruct sea and lake level. Such basins, like the Red Sea, Sea of Marmara (Turkey), Carioco
61 Basin (Venezuela) and Gulf of Corinth (Greece), have a special geologic interest, given the active tectono-sedimentary
62 processes driving their formation (e.g. Van Daele et al., 2011; McNeill et al., 2019), their sensitivity to rapid sea-level and
63 climatic changes (e.g. Aksu et al., 1999; Siddall et al., 2004), and their role in dispersion of species (e.g. Derricourt, 2005).

64 The main complexity in reconstructing sea-/lake-level fluctuations in such settings, is that 1) during disconnected phases
65 these basins may have been underfilled or overfilled depending on local hydroclimate, and 2) the structural highs (sills)
66 separating the basins from the sea can be simultaneously affected by tectonic vertical motion, sedimentation and erosion.
67 In this study, we intend to overcome common marine terrace analysis limitations, by using a Bayesian approach to invert the
68 geometry of paleo-shoreline sequences. Our approach provides probabilistic estimates of paleo sea-level, erosion rates, uplift
69 rates, wave-based depths and initial slopes. We focus on erosive marine terraces (Fig. 1b), which are both the most common
70 type of paleo-shoreline (Pedoja et al., 2014), and are simpler to model than their depositional and bio-constructed
71 equivalents (e.g. Pastier et al., 2019). We first apply our probabilistic inversion approach to a set of synthetic coastal profiles
72 to test and illustrate the method, after which we invert a well-studied marine terrace sequence in Santa Cruz (US) to
73 benchmark our model on a natural example. Finally, we use our approach on the semi-isolated Gulf of Corinth to decipher
74 the complex combination of tectonic uplift, sea- and lake-level fluctuations, local climatic drivers and sill dynamics. The
75 Gulf of Corinth has been considered as a natural rift laboratory and has therefore received a lot of attention as a primary
76 example for studying tectonic and surface processes in young rift systems (e.g. Armijo et al., 1996; Bernard et al., 2006;
77 Nixon et al., 2016). As a semi-isolated marine basin it has also been subjected to many studies on its environmental and
78 climatic evolution (e.g. Collier et al., 2000; Watkins et al., 2019), yet, so far, previous studies could not resolve to what
79 extent sea and lake levels have fluctuated.



81 **Figure 1. Paleo-shorelines and paleo-sea level.** **a)** Global compilation of paleoshoreline sequences, adjusted from Pedoja et al., 2014, **b)**
82 pictures of marine terrace sequences from Santa Cruz (US) and the Corinth Rift (Greece) and **c)** paleo-sea-level estimates for the past 450
83 ka, showing a sea-level curve (SpLi16, blue; Spratt and Lisiecki, 2016) derived from principal component analysis of 7 sea-level curves
84 with its 2.5% and 97.5% likelihood range (grey envelope), an approximation of that curve with nodes and a cubic spline interpolation (red),
85 global mean sea-level highstand estimates adjusted for glacio-istostatic adjustments (GMSL, brown dots; Kopp et al., 2009; Dutton et al.,
86 2015; Pico et al., 2016; Creveling et al., 2017; Dyer et al., 2021; Tawil-Morsink et al., 2022), selected relative sea-level highstand
87 estimates >130 ka (RSL, pink dots; Stirling et al., 2001; Murray-Wallace, 2002; Andersen et al., 2010; de Gelder et al., 2022; Marra et al.,
88 2023), global mean sea-level lowstand estimates from ice sheet data (orange dots; Batchelor et al., 2019), and red boxes that represent the
89 likely admissible range of relative sea-level elevations at locations far from the major ice-sheets (details in Supplementary Information and
90 Table S1) that we consider in this study. Marine Isotope Stages (MIS) are given in bold, and based on Railsback et al., 2015.

91 These case studies highlight how we can derive probabilistic estimates of past sea-level from marine terraces, and how the
92 natural archive of paleo-shorelines can be further utilized to improve both paleo sea-level estimates, and unravel complex
93 tectono-hydro-climatic interactions.

94 **2 Marine terrace sequence inversion**

95 Marine terraces are relatively flat surfaces of coastal origin, either horizontal or gently inclined seawards (Fig 1b; Pirazzoli,
96 2005). They are bounded inland by a fossil sea-cliff, and can be covered by a layer of coastal sediments. Here we model
97 erosive marine terraces, which are primarily formed by sea-cliff retreat in response to wave action. The superposition of
98 Quaternary sea-level variations (Fig. 1c) and vertical land movement typically leads to a staircase landscape exhibiting
99 marine terraces sequences (Fig. 1b). In the following sub-sections, we describe the inversion of marine terraces in 4 parts: 1)
100 the unknown model parameters and their relation to observed terraces, 2) the Bayesian formulation of the inverse problem, 3)
101 the Monte Carlo algorithm to approximate the probabilistic solution and 4) the bounds of the uniform prior ranges.

102 **2.1 Model parameters**

103 As a general formulation, we can consider:

$$104 \quad d = g(m) + \varepsilon \quad (1)$$

105 Where d describes the vector of observations, in our case the topographic profile of a terrace sequence, and m is the set of
106 unknown model parameters to be inverted for: uplift rate (U), erosion rate (E^*), wave-base depth (z_0), initial slope (α) and
107 sea-level history described by a set of nodes (see below). The function g describes the numerical erosion model that links
108 these model parameters to the topographic profile, here the REEF code (Husson et al., 2018; Pastier et al., 2019). Data errors
109 here are given by a random variable ε that describes the inability of the forward model $g(m)$ to explain the observations d .

110 REEF is a nonlinear model, in which wave erosion is based on the wave energy dissipation model developed by Anderson et
111 al. (1999). The model assumes that the vertical seabed erosion rate is a linear function of the rate of wave energy dissipation
112 against the seabed (Sunamura, 1992). Horizontal erosion rates depend on the energy available at the sea-cliff after
113 dissipation of the far-field wave energy (Anderson et al., 1999). The dissipation rate is dictated by the water depth profile,
114 which increases landwards exponentially with decreasing water depth. The initial erosion rate E^* is expressed as an effective
115

eroded volume per unit of time and coastal length, in which a fraction of erosional residual power E_r erodes the foundation at each location along the curvilinear profile s . This fraction E_r depends on the true local water depth along the profile $h(s)$, the water depth for wave base erosion z_0 , and a coefficient for sea bed erodibility K , so that:

$$\frac{\partial E_r}{\partial t} = K \times E_r \times \exp\left(-\frac{h(s)}{z_0}\right) \quad (2)$$

Then a residual power $E^* - \int_s \frac{\partial E_r}{\partial t} ds$ carves out a 1 m high notch and all its overhanging material to form a cliff. Following previous studies (Anderson et al., 1999; Pastier et al., 2019), for K we use 0.1 as bedrock erodibility and 1 for notch carving. Finally, the 2D model we use consists of a landmass with a seaward dipping linear initial slope (α ; Fig. 2a), an initial erosion rate (E^* ; Fig. 2a) that evolves as platforms are being carved, a wave base depth (z_0 ; Fig. 2a) that determines the vertical range over which erosion takes place, a land uplift rate (U ; Fig. 2a) and a sea-level history.

To invert the morphology of the marine terrace sequences, we parameterize sea-level history (sl) with a finite number of unknown parameters. We use nodes interpolated through a cubic spline scheme (Fig. 2b; light blue), in which each node has values for age and elevation, and the set of nodes can be expressed as:

$$v_i = (t_i, z_i | i = 1, 2, \dots, n) \quad (3)$$

This creates sea-level curves with similar characteristics to published sea-level curves (red line, Fig. 1c), in which the nodes represent sea-level minima (lowstands) and maxima (highstands) that are typically linked to even and odd-numbered marine isotope stages (MIS), respectively.

2.2 Bayesian formulation of the inverse problem

Reconstructing the sea-level history from present day observations of marine terrace sequences can be mathematically formulated as a highly non-linear inverse problem where the solution is non-unique. To embrace this non-uniqueness, the problem can be cast in a Bayesian (probabilistic) framework where the solution is a posterior probability distribution describing the probability of the model parameters (m), given the observed data (d).

One benefit of Bayesian inference is the ability to propagate uncertainty estimates from the observed measurements (ε in eq. 1) towards the unknown model parameters. For that, a likelihood probability distribution needs to be defined, based on the mathematical model for uncertainty estimates associated with observations. In this work, the data vector is defined as the horizontal position of a set of points measured on the shoreline with a vertical step size ($ipstep$; Fig. 2a). The misfit between this observed topography and the modelled paleo-shoreline sequences is calculated on the horizontal axis. As such, the elevation axis is divided in a regular grid, and the level of data fit is measured by comparing the horizontal distance between observed and simulated horizontal position at each grid point. The errors ε about the observed shorelines account for the inability of our numerical model $g(m)$ to explain observations d , and can be due to both observational (either instrumental or due to a naturally degrading morphology) or modelling errors. These errors are supposed random and normally distributed. Their statistics can therefore be described by a Covariance matrix C_d defined by a standard deviation (σ ; Fig. 2a) and a level

of spatial correlation (*corrI*; Fig 2a). In this way, the likelihood distribution defining the probability of observing the data d , given a set of model parameters m , writes:

$$p(d|m) \propto \exp(-\frac{1}{2}[d - g(m)]^T C_d^{-1}[d - g(m)]) \quad (4)$$

In order to probabilistically estimate the model parameters, Bayes' theorem can be used to combine this likelihood distribution (information given by the data), with the prior distribution $p(m)$ that represents independent information on model parameters.

$$p(m|d) \propto p(d|m) \times p(m) \quad (5)$$

The solution of the inverse problem $p(m|d)$ is the posterior solution and represents the probability of the model parameters given the data. The prior distribution for each parameter is defined as a uniform distribution between two bounds. That is, we assign a priori equal probability to all the values within the specified range. For example, the position of the nodes defining the sea level history can take a priori any value within the red boxes in Figure 2.

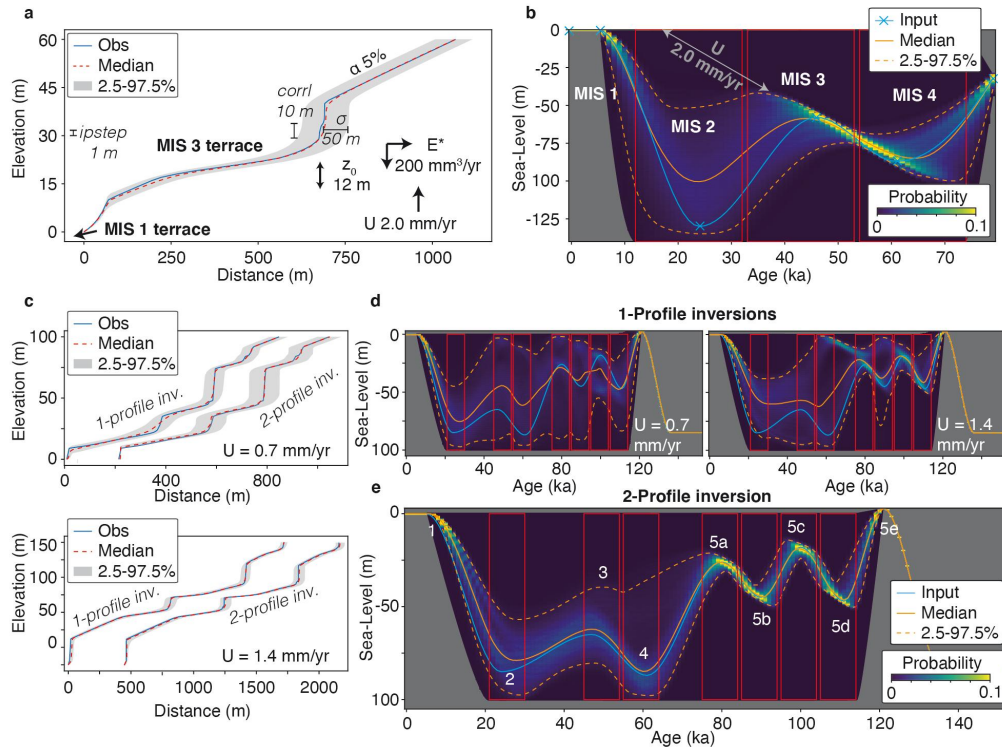
Note that the misfit can also be calculated along vertical axis (i.e. comparing elevations between observed and estimated profiles). We tested this (Fig. S1), but found it harder to reproduce realistic marine terrace sequences with a few m of terrace height variability, and a few hundred meter of terrace width variability (e.g. Regard et al., 2017; De Gelder et al., 2020). Regardless, we note that changing the misfit calculation from horizontal to vertical misfit does not change the paleo sea-level posterior distribution (Fig. S1).

2.3 Sampling the posterior solution

We use a Markov chain Monte Carlo algorithm to sample the posterior distribution and explore the range of models that can explain the observed topography within errors. For a review of Bayesian inference and Monte Carlo methods in the geosciences, we refer the reader to Mosegaard and Sambridge (2002), and Gallagher et al. (2009). In the Monte Carlo exploration of the model space, nodes can either be fixed at certain ages and elevations, or treated as unknown parameters, and left free to move within a prescribed range (e.g. red boxes in Fig. 2b,d,e). The 4 main erosion model parameters (α , E^* , z_0 , U) can also be fixed to chosen values, as done in the synthetic tests below, or left free within chosen ranges, as done for the Santa Cruz and Corinth examples below. The algorithm samples the parameter space as a random walk, where at each iteration a new sea level history model is proposed by perturbing the current one. The proposed model is then either accepted or rejected following an acceptance rule based on the level of data fit of the current and proposed models. The final solution is a large ensemble of paleo sea-level models that approximates the probabilistic solution. That is, the distribution of models follows the posterior probability solution. Of the 1 million model runs per figure, the first 50 000 models were discarded as burn-in models. To verify that the random walk samples the target distribution, we show a number of convergence diagnostics in Fig. S2, which includes parameter acceptance ratios and likelihood evolution for all model runs in this article.

2.4 Bounds of the uniform prior

181 For the different unknown model parameters, imposed prior constraints can be either restrictive, open or anything in between.
 182 Within the synthetic tests (section 3) we left the sea-level nodes open, and the other parameters either fixed (Fig. 2) or open
 183 (Fig. S3). For the Santa Cruz benchmark tests (section 4) we left the erosion rate, wave base depth and initial slope
 184 parameters relatively open, but placed a more restrictive range on the uplift rate of 1.3-1.65 mm/yr, so that the
 185 chronostratigraphy of the modeled terrace sequence would match published ages (Perg et al., 2001). We put soft prior
 186 constraints on the sea-level history, by restricting possible solutions to the red boxes in Fig. 1c, which represents a cautious
 187 interpretation of the ensemble of previous sea-level studies. We adopted a similar strategy for the Corinth terraces (section 5),
 188 with the erosion rate, wave base depth and initial slope parameters left relatively open, but stronger prior constraints on uplift
 189 rate ranges to respect previous findings on the chronostratigraphy. We tested models with the sea-/lake-level nodes either
 190 completely open between 15 and -150 meter elevation (Fig. S4), and with stronger prior constraints only on the highstands.
 191 During these highstands the Gulf of Corinth undoubtedly experienced marine conditions (McNeill et al., 2019), so the sea-
 192 level elevations can be expected to fall within the red boxes of eustatic sea-level defined in Fig. 1c, whereas lowstands are
 193 likely lacustrine.



194 **Figure 2. Inversion of synthetic marine terrace profiles.** a) Synthetic topography (Obs, blue) created from a forward model with known
 195 input parameters: α = Initial Slope, E^* = Erosion Rate, U = Uplift Rate and z_0 = wave-base depth. The posterior range of models that fit
 196 the observed topography with the given σ , $ipstep$ and $corrl$ (see text) is represented by the median (orange), and the 2.5 and 97.5

197 percentiles of the inverted models (grey envelope). **b)** Posterior probability distribution for the sea-level histories. Each individual paleo
198 sea-level history is described with 6 sea-level nodes linked with a cubic spline interpolation, of which the nodes at 78, 6 and 0 ka are fixed
199 in time and elevation, and the other three nodes can move within the three red boxes. The input (target) sea-level history is given in light
200 blue, and the probabilistic solution is depicted by the median (solid orange line) and the 2.5 and 97.5 percentiles (dashed orange lines). **c)**
201 Same as **a**, but with different uplift rates and sea-level histories. **d)** Sea-level histories for the inverted profiles in **c**, similar to **b** but with a
202 different input sea-level history including more nodes. **e)** Similar to **d**, but inverting the two profiles simultaneously to find a common sea-
203 level curve explaining both profiles. MIS are marked in white.

204 3 Synthetic marine terrace profiles

205 To test and illustrate the potential of the inversion approach, we inverted synthetic topographic profiles that were produced
206 by forward models with known input parameters (Fig. 2). To start with a relatively short and simple sea-level range, we
207 defined an 80 ka sea-level history consisting of 6 nodes (Fig. 2b; light blue). For the inversion, we fixed the elevation and
208 ages of the nodes at 0, 0 and -30 m and 78, 6 and 0 ka, respectively, and the positions of the other three nodes were left as
209 unknown model parameters to be recovered. In the Monte Carlo exploration of the model space, these three nodes were left
210 free to move within a prescribed range (red boxes in Fig. 2b). All other erosion model parameters (α , E^* , z_0 , U ; Fig. 2a)
211 were fixed during the inversion at the values used to produce the observed topographic profile. The parameters σ , $ipstep$ and
212 $corrl$ were set at 50, 1 and 10 m, respectively. We inverted the topographic profile between 0 and 60 m elevation by
213 sampling the parameter space with 1 million forward simulations. The solution is a large ensemble of sea-level histories that
214 reflect the probability of the paleo sea-level, given the synthetic coastline topography.

215 The resulting profiles show an MIS 3 terrace at an elevation range of ~15-30 m (Fig. 2a), whereas an MIS 1 terrace lies
216 below the present-day sea level, and is thus not considered in the inversion. As such, the range of sea-level histories that
217 could have created the MIS 3 terrace is narrower than for the MIS 1 terrace (Fig. 2b). This range is particularly limited for
218 the period of sea-level rise leading up to the MIS 3 peak, suggesting that uplifted marine terraces are more likely to form
219 during periods of relative sea-level rise. This is theoretically expected, as erosion scales with the total duration of sea-level
220 occupation (Malatesta et al., 2021), and simultaneous sea-level rise and land uplift implies favorable conditions for the
221 formation of marine terraces. Another notable feature is the distribution of possible sea-level histories along a diagonal line
222 that corresponds to the uplift rate. This line would reach the maximum terrace elevation when extrapolated to $t=0$ ka, in line
223 with classic graphical methods (Bloom and Yonekura, 1990). Although the MIS 1 terrace is not inverted, there are some
224 limitations to the magnitude and rate of sea-level rise between MIS 2 and MIS 1 (Fig. 2b), probably because this period
225 determines how much of the MIS 3 terrace is eroded at its distal edge.

226 For the inversion of every individual profile there should be a trade-off between younger, higher sea-level peaks and older,
227 lower sea-level peaks in line with the fixed uplift rate (as in Fig. 2b). These trade-off effects can be overcome through the
228 joint inversion of multiple profiles with different uplift rates, reducing the uncertainty in sea-level reconstructions. To show
229 this, we also inverted two different topographic profiles produced with different fixed uplift rates but with the same sea-level
230 history over a 135 ka timescale (the last glacial-interglacial cycle; Fig. 2c-e). When the two profiles are inverted individually,
231 the range of possible sea-level histories is relatively wide, and again the sea-level peaks would follow a diagonal line parallel

to the uplift rate (Fig. 2c, d). However, if we jointly invert both profiles, i.e. assuming that a unique sea-level history would have created both marine terrace staircase morphologies, the probability distribution for past sea-level narrows, and the median sea-level of the inversion better approximates the input curve (Fig. 2e). The range is particularly narrow for the transgressions leading up to the MIS 5a and 5c highstands, for which the corresponding terraces are well developed in the topographic profiles (Fig. 2c). Similar to the MIS 1 terrace in Fig. 2a, the MIS 1 and 3 terraces in Fig. 2c would be located below sea level for the given parameters, and thus the possible sea-level range is wider for the transgressions leading up to MIS 1 and 3 (Fig. 2d). Also for these highstands though, the sea-level is better constrained for the joint inversion (Fig. 2e) than with the individual inversions (Fig. 2d). This suggests that jointly inverting more profiles would increase even further our ability to constrain sea-level histories. To understand what happens in scenarios in which more parameters are unknown, we repeated the same tests (from Figures 2b-d) with broad prior ranges for the uplift rate, initial slope, erosion rate and wave base height (Fig. S3). Also in this case the joint inversion provides much narrower posterior ranges for paleo sea-level compared to the two individual profile inversions, and not too different from the cases with fixed model parameters (Figs. 2b-d). The posterior ranges for all parameters are consistently smaller for the joint inversions, compared to the individual inversions (Fig. S3).

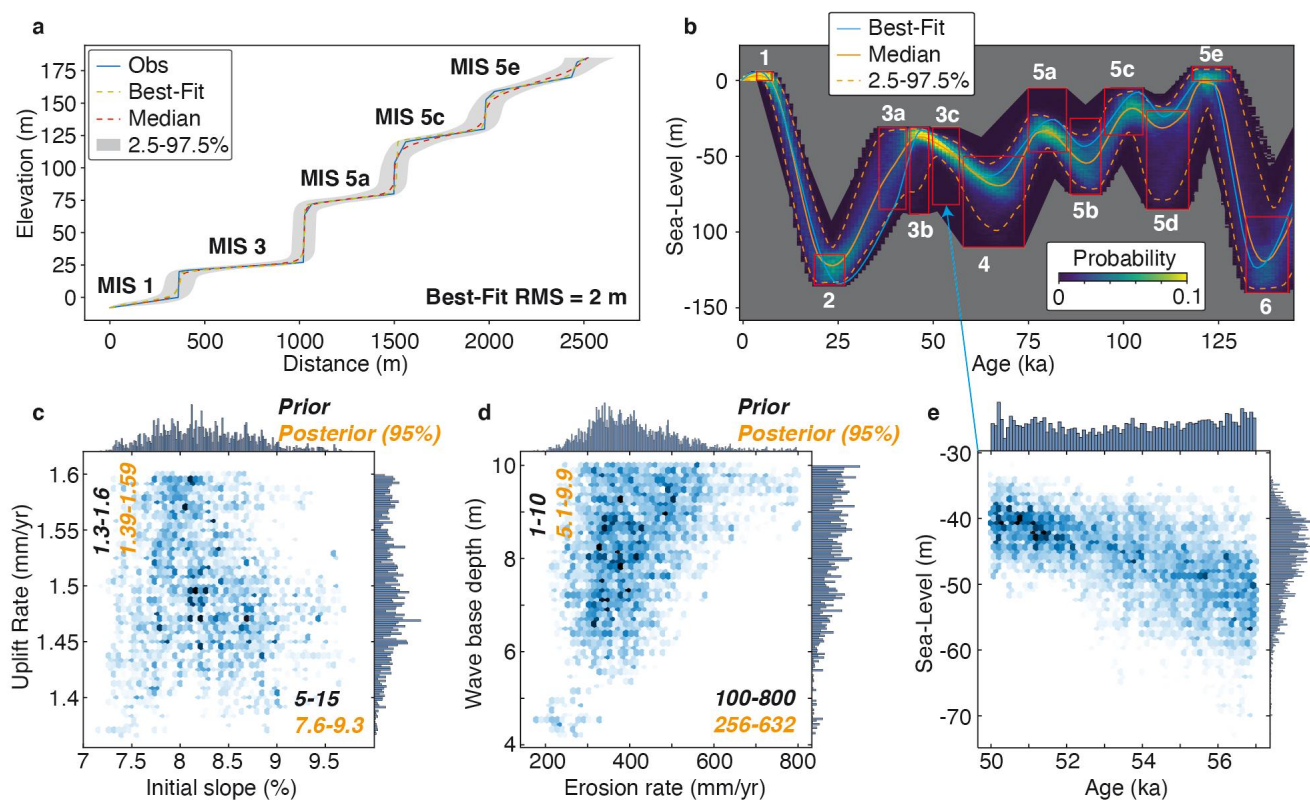
These synthetic tests imply that in natural examples, sea-level reconstruction should also benefit from the inversion of multiple marine terrace profiles if conditions change between those profiles. In this example we used two different uplift rates for the joint inversion, which lead to a range in different terrace sequence morphologies (Fig. 2c), but an approach where all parameters, including wave base, erosion rate or initial slope, are undefined a priori (or only within a given range), should lead to a more realistic range of possible sea-level histories.

To put this method to the test in real cases, we selected two well-documented yet contrasting cases, Santa Cruz and the Gulf of Corinth, each having their peculiarities that make them ideal to study the inversion of marine terraces. Santa Cruz is well documented and possible parametric windows are quite narrow, making it an ideal benchmark site for our method. On the other hand, the staircase sequence in the Gulf of Corinth, while equally well documented, is particularly relevant to take advantage of the predictive capacities of our method. There, we can decipher the complex interplay between vertical land motion and sea-/lake-level fluctuations in a semi-closed basin, and better reconstruct its hydrodynamic history where paleoclimatic data are insufficient.

4 Santa Cruz marine terrace sequence inversion

The marine terraces along the Santa Cruz coastline (Central California, US) formed through a combination of Quaternary sea-level oscillations and tectonic uplift by nearby active faults (e.g. Bradley, 1957; Anderson and Menking, 1994; Anderson et al., 1999; Perg et al., 2001; Matsumoto et al., 2022). We invert a topographic profile from Rosenbloom and Anderson (1994), who distinguished the original eroded bedrock surface, which we use, from its overlying colluvium for 5 marine terraces. We followed the age interpretation of Perg et al. (2001), suggesting these terraces were formed, from bottom to top,

264 during MIS 1, 3, 5a, 5c and 5e. Unlike in the synthetic tests, here we left uplift rate, erosion rate, wave base depth and initial
 265 slope parameters free within a prior range of values. We use the elevation (~ 170 m) and age (~ 118 -128 ka) of the upper
 266 terrace to derive a range of possible uplift rates (1.3-1.65 mm/yr), and simultaneously consider ranges for initial slope (5-
 267 15%), wave base depth (1-10 m) and erosion rates (100-800 mm³/yr) in the terrace inversion. We use the same inversion
 268 parameters as for the synthetic tests, running 1 million models over 450 ka with the sea-level high- and lowstands limited to
 269 the red boxes in Fig. 1 (See Supplementary Information and Table S1). We tested different inversion parameters (Fig. S5),
 270 and settled on values of 1 m *ipstep*, 100 m σ and 10 m *corrl*, as for these values the corresponding 95% posterior ranges for
 271 the variability in terrace height (~ 5 m; Fig. S5) and width (~ 150 m; Fig. S5), are along the same order of magnitude as the
 272 variability in a swath profile across the Santa Cruz terraces (Fig. S6). Higher values for σ and *corrl* result in both broader
 273 ranges for accepted terrace heights/widths, and consequently also in broader posterior ranges for paleo sea-level, although
 274 we note that the first order shape of the paleo sea-level curves remains similar irrespective of the tested *ipstep*, σ and *corrl*
 275 values (Fig. S5).



276 **Figure 3. Inversion of NW-Santa Cruz marine terrace sequence. a)** Observed topography (from Rosenbloom and Anderson, 1994; Obs,
 277 blue) with the age interpretation of Perg et al. (2001) marked in bold, together with the modeled best-fit, median, 2.5% and 97.5%
 278 percentile profiles. **b)** Probabilistic sea-level reconstruction for the profiles in **a**, MIS in white **c)** Posterior ranges for uplift rates and initial

279 slopes (histogram of the sampled models), **d**) Posterior ranges for wave base depths and erosion rates **e**) Posterior ranges for the MIS 3c
280 peak, i.e. distribution for the position of the 50-57 ka node within the paleo sea level curve.

281 The sampled sea-level histories successfully reproduce the terrace morphology, as evidenced by the low misfit of 2 m (Fig.
282 3a). As with the synthetic tests (Fig. 2), periods of sea-level rise are better constrained than periods of sea-level fall, and
283 highstands better constrained than lowstands (Fig. 3b). Also here there is a trade-off in sea-level peaks, in which younger,
284 higher sea-level peaks could result in similar shaped marine terraces as older, lower sea-level peaks (e.g. for MIS 3c, Fig. 3e).
285 The models limit the uplift rate to ~ 1.35 - 1.6 mm/yr, the initial slope to ~ 7 - 9.5% , the wave base depth to 4-10 m and the
286 erosion rate to 200-800 mm/yr (Figs. 3c,d). Notably there is a positive correlation between wave base depth and erosion rate
287 (Fig. 3d), suggesting a higher value for wave base depth would require a higher erosion rate to create the same marine
288 terrace sequence morphology.

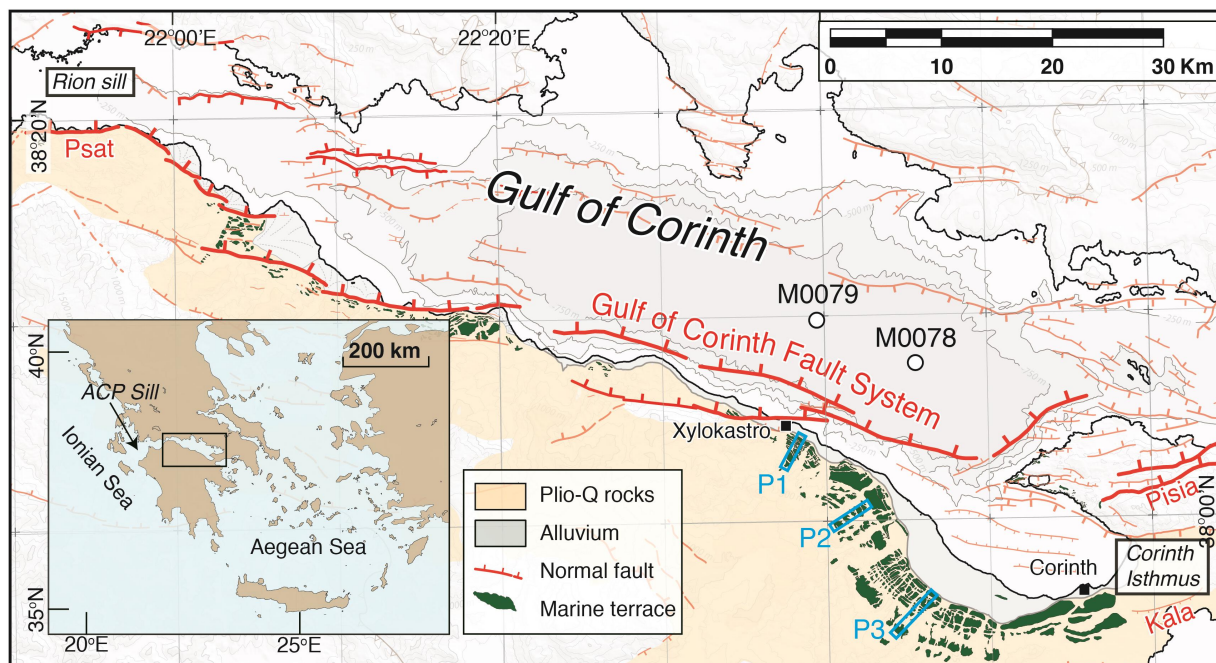
289 Compared to our proposed prior range of possible sea-level elevations for MIS 3 (-30 to -80 m; Fig. 3b), the posterior
290 distribution of the inversion suggests paleo sea-level values on the higher end of that spectrum. This is in agreement with a
291 growing number of studies suggesting oxygen-isotope derived sea-level curves underestimate sea-level for that period (Pico
292 et al., 2016; Dalton et al., 2019, 2022; Gowan et al., 2021; De Gelder et al., 2022). For MIS 5a on the other hand, the
293 posterior distribution of the inversion suggests a sea-level peak on the lower end of our proposed range of sea-level
294 elevations (Fig. 3b). Although the highstand posterior distributions still span a broad elevation range of ~ 25 m, these results
295 tend to align with studies proposing an overall decrease in sea-level between MIS 5e, 5c and 5a (e.g. Chappell and
296 Shackleton, 1986; Schellmann and Radtke, 2004; Tawil-Morsink et al., 2022).

297 We also tested additional uplift rate scenarios (Fig. S7), given that there has been concerns on the terrace chronology that we
298 adopted (Brown and Bourlès, 2002), and other studies have suggested the terrace at 27 m elevation might be formed during
299 MIS 5a, 5c or 5e instead of MIS 3 (Bradley and Addicott, 1968; Lajoie et al., 1975; Kennedy et al., 1982; Weber et al.,
300 1990). These uplift rates can fit the terrace sequence morphology equally well in terms of topographic misfit, but generally
301 imply a larger possible range of paleo sea level. This can be explained by the increased terrace re-occupation for lower uplift
302 rates (Malatesta et al., 2021), which also explains why the posterior ranges for the initial slope and wave base depth change
303 increase for lower uplift rate scenarios (Fig. S7), and erosion rate estimates decrease. These tests suggest that locations with
304 higher uplift rates will generally provide narrower constraints on paleo sea-level, while still providing realistic and unbiased
305 parameter estimates.

306 **5 Gulf of Corinth marine terrace sequence inversion**

307 The complexity of semi-isolated basins, connected to sea during some time intervals and isolated in others, make them
308 ultimate testing grounds for our modelling approach. We focus on one of such basins, the SE Gulf of Corinth, to derive a
309 sea-/lake-level history from terrace sequence geometries, and compare its outcomes to paleoclimate data, tectonic structures
310 and sill dynamics.

311 Natural interaction between the Gulf of Corinth and the open sea is currently restricted by the Rion and Acheloos-Cape
 312 Pappas sills at its W entrance at ~45-60 m depth (Fig. 4; Beckers et al., 2016). In the past there was an additional connection
 313 at its E end along the Corinth Isthmus, currently uplifted at ~80 m elevation but consisting of Quaternary marine sediments
 314 (Fig. 4; Caterina et al., 2023). These sills have controlled the Gulf's connection with the open sea over the past few
 315 hundred thousand years and lead to an alternation of marine and (semi-)isolated lake environments within the Gulf (McNeill
 316 et al., 2019). Although we approximately know the timing of these alternations, it remains unclear whether sill depths
 317 remained stable or fluctuated throughout the Quaternary (Roberts et al., 2009; McNeill et al., 2019), and in addition, whether
 318 lake levels were stable or fluctuating during periods with no connection to the sea. Hydroclimatic conditions, i.e. the balance
 319 between inflow, outflow, precipitation and evaporation, determine the level of the lake, and state-of-the-art paleo-
 320 environmental reconstructions (e.g. Kafetzidou et al., 2023) are insufficient to infer its fluctuations over time.



321
 322 **Figure 4. Tectonomorphology of the Gulf of Corinth.** Map showing the main features of the Gulf of Corinth, including the active faults,
 323 marine terraces, profile locations used in the inversion and connections to the Ionian Sea (Rion Sill) and Aegean Sea (Corinth Isthmus).
 324 M0078 and M0079 indicate the IODP-381 sedimentary coring locations (McNeill et al., 2019). Psat = Psathopyrgos Fault, Pisia = Pisia
 325 Fault, Kala = Kalamaki Fault. Modified from de Gelder et al. (2019).

326 Terrace sequences are well exposed in the SE of the Gulf, where the Gulf of Corinth Fault System (Fig. 4) has lead to
 327 differential coastal uplift rates (Armijo et al., 1996; De Gelder et al., 2019; Fig. S8). This peculiarity allows us to test on a
 328 natural example whether the joint inversion of multiple terrace sequence profiles with different uplift rates provides a better-
 329 constrained sea-/lake-level history (as in Fig. 2). To account for the unknown range of possible lake-level elevations, we
 330 carried out inversions with all nodes from (semi-)isolated periods broadly constrained between -15 and -150 m elevation. For
 331 the marine periods we follow the more restricted eustatic sea-level ranges defined in Fig. 1bc (red boxes), as the resulting

332 posterior sea-/lake-level ranges would remain similar to the prior ranges for tests in which we also gave a lot of freedom to
 333 nodes for the marine periods (Fig. S4). We selected three topographic profiles with little river incision and ~0.4-1.4 mm/yr
 334 uplift rates (Fig. S8), and avoided modelling the broad coastal plains at the base of all profiles that appear to have been
 335 modified by human presence (Fig. S8). We used the 90% percentile of 100-m wide swath profiles to obtain representative
 336 terrace sequence morphologies (Fig. S8). For the three profiles we assigned prior ranges of possible uplift rates of 1.25-1.4,
 337 0.7-0.9 and 0.4-0.55 mm/yr (De Gelder et al., 2019; Fig. S8), and broad prior ranges for erosion rate (100-1500 mm/yr),
 338 initial slope (1-20%) and wave base depth (1-12 m).

339 The individual profile inversions mostly constrain paleo sea/lake level for profile 1 (Fig. 5a), because it has the highest uplift
 340 rate and contains most terraces. The other two profiles provide limited constraints on paleo sea/lake level when inverted
 341 individually (Fig. 5b/c), but when jointly inverted with profile 1 they provide a much narrower range in terms of posterior
 342 distribution for sea-level (Fig. 5d). The cumulative misfit for the individual inversions (28 m) is slightly better than for the
 343 joint inversion (46 m), but there are no major visible differences between the terrace sequence profiles for the two inversions,
 344 and apart from the highest terrace of profile 2 (Fig. 5b) the terrace sequences are all near perfectly reconstructed. The three
 345 profiles show variations in initial slopes that are in line with the overall morphology, i.e. present-day profile 1 is steeper than
 346 2, which is steeper than 3, which is also what we find for their initial slopes. The three profiles do have similar posterior
 347 distributions for wave base depths and erosion rates (Fig. 5e). Although we might expect lateral differences in these rates
 348 given variability in sediment types, catchment area and coastal orientation, the broad ranges for the posterior distributions
 349 indicate we cannot quantify these lateral differences from the profile morphology alone. The posterior parameter ranges
 350 mostly remain the same between the individual and joint inversion, with exception of the uplift rates for profiles 2 and 3 that
 351 became a little lower for the joint inversion. As for the sea-/lake-level inversion, all the other posterior parameter ranges
 352 become narrower for the joint inversion (Fig. 5e).

353 The inverted sea-/lake-level history (Fig. 5d) shows a few notable features. To a first order, fluctuations resemble global sea-
 354 level trends, with relatively fast periods of sea-/lake-level rise prior to major sea-level highstands, followed by long periods
 355 of slow sea-/lake-level fall (Fig. 1c). Yet, unlike global sea-level trends, there are several periods of prolonged stability, in
 356 particular around 180-200 ka and 275-300 ka, and possibly also around 75-95 ka and 360-370 ka. In addition, glacial periods
 357 are often surprisingly poorly resolved, like during the period 20-75 ka or 160-180 ka. In the last section we discuss our
 358 interpretation of these trends, and show how they provide insightful arguments to decipher the relation between water level,
 359 fault activity, paleoclimate and tectonics.

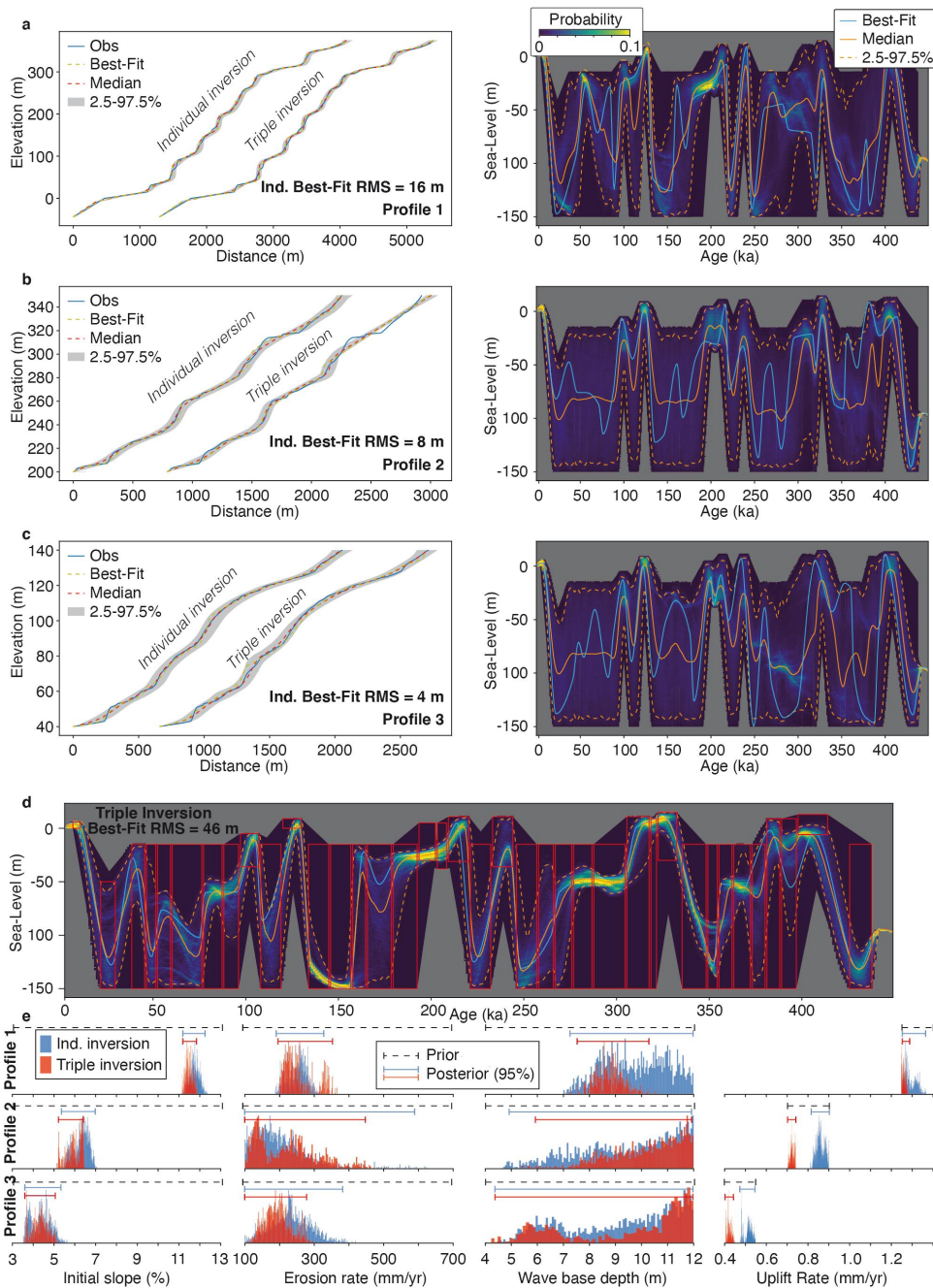


Figure 5. Inversion of SE Corinth Rift marine terrace sequence (previous page). a-c) Observed topography (left) from 3 different profiles in the SE Corinth Rift (locations see Fig. S8), together with the modeled best-fit, median, 2.5% and 97.5% percentile profiles for both an individual profile inversion and a joint inversion of the three profiles (horizontally offset by an arbitrary value). Corresponding probabilistic sea/lake-level ranges for the individually inverted profiles are given on the right, d) Posterior distribution for sea/lake-level history from joint inversion, with red boxes describing prior ranges e) Prior and posterior parameter ranges from both individual and joint profile inversions

367 6 Discussion

368 6.1 Beyond sea level: tectono-hydro-climatic processes in the Gulf of Corinth

369 While the results for the joint inversion of the 3 profiles in the Gulf of Corinth permit to unravel sea- and lake-levels through
370 time as expected, we also stress that such inversions yield more, when integrated within the broader environmental setting:
371 these reconstructions allow for a more detailed look into sea- and lake-level fluctuations within a (semi-)isolated basin.
372 Figure 6 compares our inverted sea-lake level to the stratigraphy, facies and pollen content within two sedimentary cores
373 from the sea floor of the central basin (McNeil et al., 2019; Gawthorpe et al., 2022; Kafetzidou et al., 2023). Based on those
374 combined datasets, we propose that the main hydroclimatic modes that have occurred in the Gulf of Corinth throughout the
375 past 450 ka, are 1) marine Gulf of Corinth, 2) transitional Gulf of Corinth or overfilled Lake Corinth and 3) underfilled Lake
376 Corinth. The first 2 of those have been proposed before based on sedimentary cores (McNeill et al., 2019; Gawthorpe et al.,
377 2022), whereas we base the occurrence of intervals with an underfilled Lake Corinth on our marine/lake terrace inversion, as
378 available paleoclimatic proxies are unable to uncover them.

379 The major peaks in our reconstructed sea/lake-level curve occurred during interglacial sea-level highstands, as direct
380 consequence of our choices in the ranges of the priors, when sea level in the Gulf of Corinth was similar to eustatic sea level
381 (marine mode M; Fig. 7a). Sedimentary cores indicate marine conditions (McNeill et al., 2019), the corresponding stratal
382 packages are bioturbated, and associated sedimentary facies are types FA1 and FA6 (Fig. 6; see caption for facies
383 description). From pollen records, the typical reconstructed biomes are cool mixed evergreen needleleaf and deciduous
384 broadleaf forests, indicating relatively warm and wet conditions with low amounts of steppic taxa (Kafetzidou et al., 2023;
385 Fig. 6).

386 We interpret the interstadial periods around 75-95 ka, 180-200 ka, 275-300 ka and 360-370 ka as periods with an overfilled
387 Lake Corinth, possibly with some marine incursions indicating a transitional Gulf of Corinth (T/O mode; Fig. 7b). This
388 would explain the prolonged sea-/lake-level stability, during interstadial periods when eustatic sea-level fluctuated by tens of
389 meters (e.g. Spratt and Lisiecki, 2016; De Gelder et al., 2022). In that case, sea-/lake-level elevations would correspond to
390 the paleo-sill depth of the Rion Sill and/or Corinth Isthmus (white line, Fig. 6). Within the sedimentary cores, these periods
391 are mostly characterized by laminated stratal packages, and associated sedimentary facies are types FA2, FA3 and FA4 (Fig.
392 6; see caption for facies description). The occurrence of marine incursions into Lake Corinth during these interstadial periods
393 is suggested by dated corals of ~76 ka, ~178 ka and ~201 ka (Roberts et al., 2009; Houghton et al., 2003) as well as the
394 white, aragonite-rich laminations of FA3 and FA4. In other locations such laminations have been linked to (seasonal) mixing
395 of marine and non-marine surface waters (Sondi and Juracic, 2010; Roeser et al., 2016).

396 The glacial periods are characterized by relatively low sea/lake-level elevations, possibly even down to the lower limit of -
397 150 m we used in the inversion. We interpret these periods as underfilled Lake Corinth conditions (U mode; Fig. 7c), during
398 which water inflow was lower than water evaporation within the lake, and lake level fell down to tens of meters below the
399 sill depth. Sedimentary cores indicate non-marine conditions (McNeill et al., 2019), the corresponding stratal packages are

mostly bedded and associated sedimentary facies are types FA5 and FA11 (Fig. 6; see caption for facies description). Reconstructed biomes from pollen suggest an increase in open vegetation such as grassland and steppe communities under colder and drier conditions (Kafetzidou et al., 2023; Fig. 6), matching reconstructed periods of lake underfilling. In general the inverted resolution of the lake-level elevation is much lower for these periods, with large probabilistic ranges. We attribute this to the occurrence of rapid lake-level fluctuations, like in other isolated E-Mediterranean water bodies such as the Dead Sea (Stein et al., 2010) and Lake Van (Turkey; Landmann et al., 1996). In such environments, the lake level is determined by the budget between runoff and evaporation, and quick variations are expected. Alternatively, this could also be due to the fact that terraces formed during low sea/lake level get increasingly eroded during transgressions. The interstadial periods with prolonged sea-/lake-level stability also allow for a possible reconstruction of sill depths through time (white lines, Fig. 6). The westernmost sill, the Acheloos-Cape Pappas Sill (Fig. 4), is currently at a depth of ~45-48 m. While there are no major faults there, we can't exclude slow subsidence or uplift at a few tenths of mm/yr, cumulating to a few tens of meters on the 100 kyr time scale. The Rion Sill, at the western entrance to the Gulf of Corinth, is currently at ~62 m depth (Perissoratis et al., 2000). As it is located in the hanging wall of the Psathopyrgos Fault (Fig. 5), active since at least the past ~200 ka (Houghton et al., 2003), the Rion Sill was unlikely deeper in the past. We reconstruct the Rion Sill depth assuming marine incursions around ~76 ka (Roberts et al., 2009) took place through this sill, and the Rion Sill was not lower than sea/lake level during the overfilled/transitional interval around ~200 ka. Extrapolating the trend, it would make sense for the older connections between Lake/Gulf of Corinth and the open sea to have occurred primarily through the Corinth Isthmus at the eastern end of the Gulf of Corinth. The Corinth Isthmus is currently at an elevation of ~80 m, and has been uplifted through the Pisira Fault, Kalamaki Fault and/or a regional uplift (Armijo et al., 1996; Roberts et al., 2009; Caterina et al., 2022). We reconstruct the Corinth Isthmus depth assuming lake/sea level during overfilled/transitional intervals around ~290 and ~360 ka correspond to the Corinth Isthmus depth, and the isthmus was not lower than sea/lake level during the overfilled/transitional interval around ~200 ka. Extrapolating this trend fits with the current Corinth Isthmus elevation of ~80 m. The isthmus elevation before ~360 ka is difficult to constrain from our data, but was possibly shallower before, given the small amount of marine sediments deposited around the ~400 ka interglacial period, and the lack of deposits within the Isthmus stratigraphy older than ~350 ka (Collier and Dart, 1991; Caterina et al., 2022). Our reconstruction of both Rion Sill and Corinth Isthmus fits with the sedimentary interpretation of a tidal strait around ~300 ka at the isthmus, with marine connections on both ends of the Gulf of Corinth (Caterina et al., 2022)

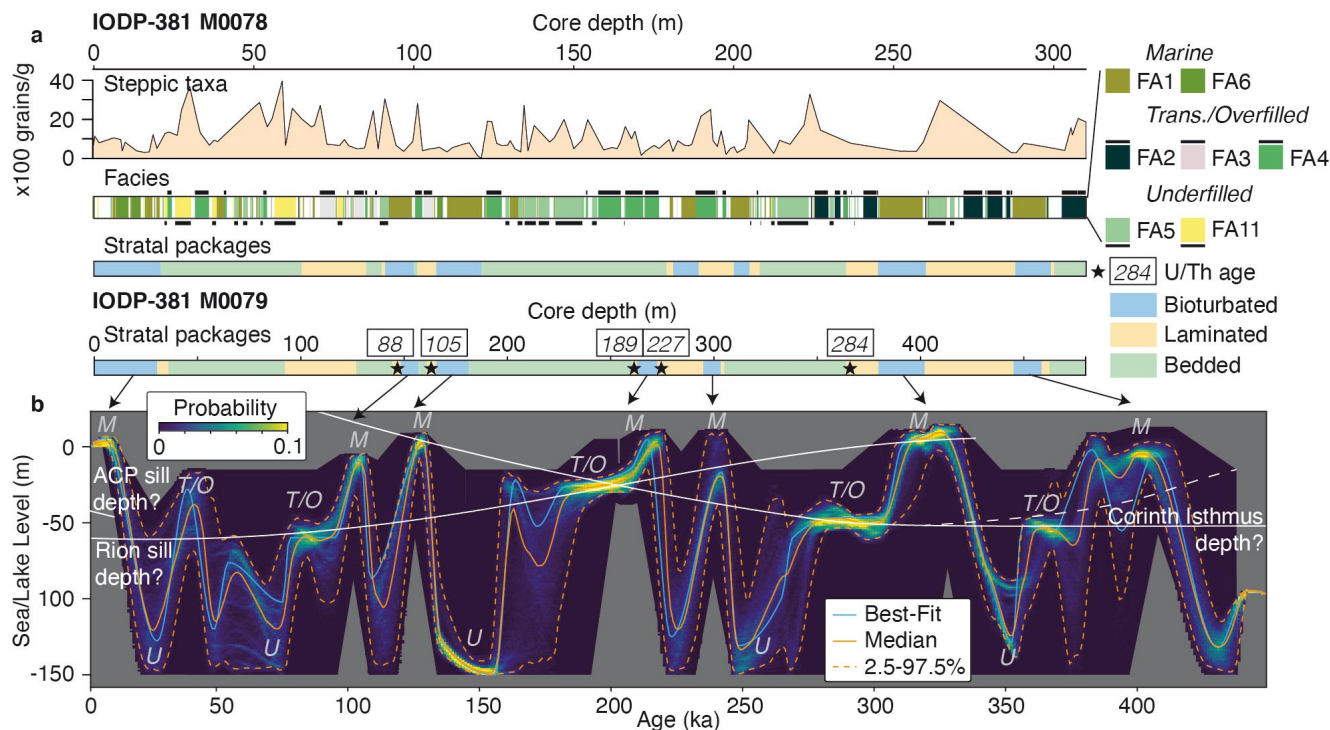


Figure 6. Comparison of sea-/lake-level reconstruction to other datasets. a) Comparison to IODP-381 cores M0078 and M0079, with steppe taxa from Kafetzidou et al. (2023), facies from McNeill et al. (2019) and stratal packages with U/Th ages from Gawthorpe et al. (2022). FA1: homogenous mud, FA2: greenish gray mud with dark gray to black silty-to-sandy beds (cm-scale), FA3: light gray to white sub-mm laminations (cc or aragonite) alternating with mud-silt beds, FA4: laminated greenish gray to gray mud with muddy beds, FA5: greenish gray mud with homogeneous cm thick gray mud beds, FA6: green bedded partly bioturbated mud, silt and sand, FA11: interbedded mud/silt and cm thick sand beds. **b)** Inversion result from Fig. 5d with proposed sill/isthmus elevations, marking periods with marine (M), transitional or overfilled (T/O) and underfilled (U) hydroclimatic modes.

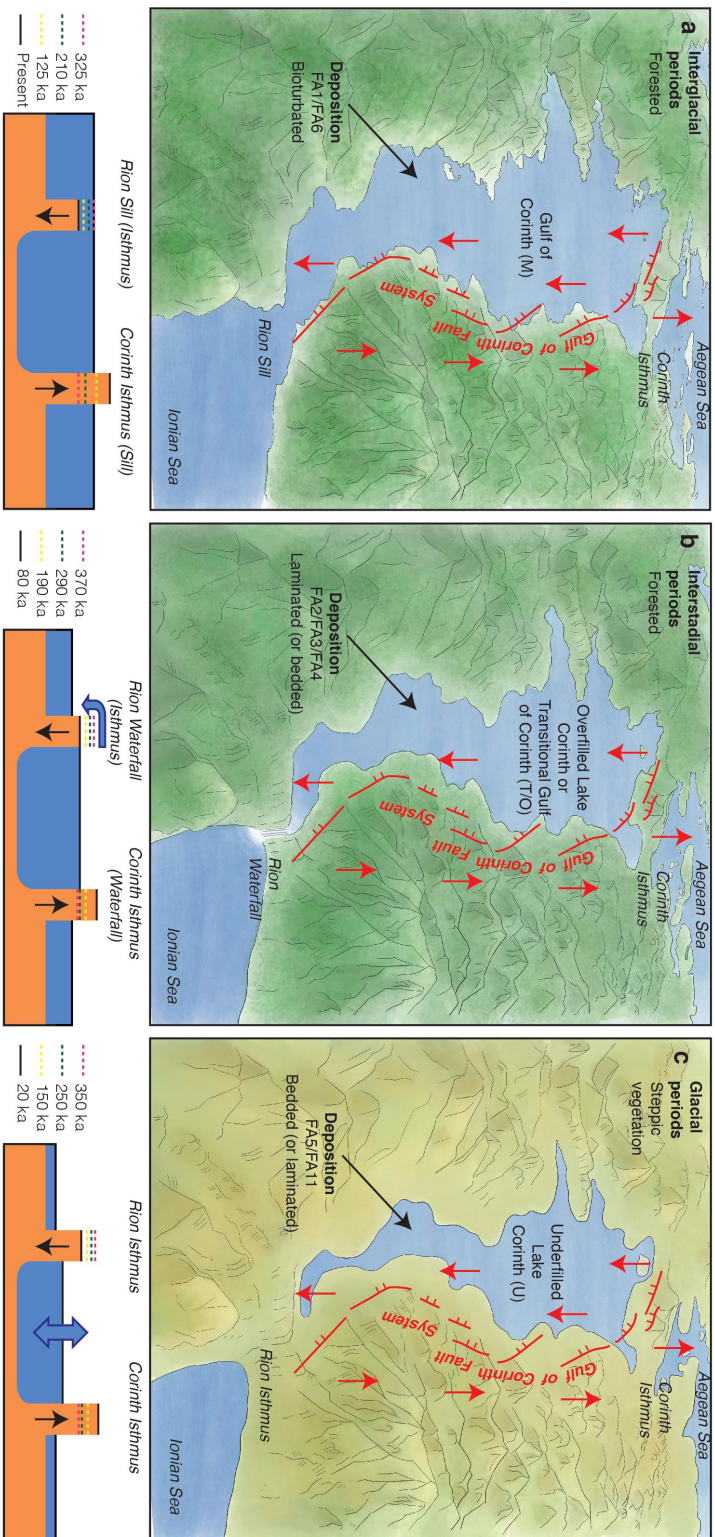


Figure 7 (Previous page). Schematic illustrations of the different hydroclimatic modes in the Lake/Gulf of Corinth. Illustrations of a) the interglacial marine (M) hydroclimatic mode as it is today. Around MIS 5e (~125 ka) the marine connection also occurred above the Rion Sill, with the Corinth Isthmus a land bridge like it is today. Around MIS 7a-c (~210-240 ka) and MIS 9 (~325 ka) the marine connection occurred above both the Rion and Corinth Sills. Around MIS 11 (~400 ka) the marine connection would have occurred above the Corinth Sill, with a Rion Isthmus on the W end of the Gulf of Corinth. b) The transitional or overfilled (T/O) hydroclimatic mode at ~80 ka, with a narrow connection or waterfall to the open sea at the Rion Sill. Around ~190 ka a narrow connection or waterfall to the open sea could have occurred at both the Rion and/or Corinth Sill, and at ~290 ka and ~370 ka only at the Corinth Sill, with a Rion Isthmus on the W end of the Gulf of Corinth. c) The underfilled (U) hydroclimatic mode at ~20 ka. As for the older glacial periods (~150, ~250 and ~350 ka), land bridges occur on both side of Lake Corinth, at the Rion and Corinth Isthmus. In all illustrations red arrows indicate uplift and subsidence in the footwall and hanging wall of the Gulf of Corinth Fault System, respectively.

Our exploration of sea-/lake-level variations in the Gulf of Corinth demonstrates the strength of using marine/lacustrine terrace sequence inversion. Although several questions remain – like the effects of erosion and sedimentation on sill evolution, or the effects of non-constant uplift rates of the marine terrace sequences – we are able to provide a solid framework that can explain several different tectonic and hydro-climatic processes simultaneously. Furthermore, our results may provide a window to explore the links between paleogeographic and biologic evolution, with our proposed history of sills, waterfalls and land bridges (Fig. 7) theoretically impacting temporal isolation and mixing of species between the Peloponnese and the Greek mainland. We distinguish 4 different hydroclimatic modes of for the Lake/Gulf of Corinth, that have probably also occurred in other (semi-)isolated basins like the Sea of Marmara or the Carioco Basin. Marine and transitional modes will most likely depend on eustatic sea-level elevations and sill depths, whereas over- or underfilled lakes likely depend on sill depths and local climatic conditions. For the Corinth Lake we show that this transition from over- to underfilled lakes occurs during changes from interstadial to glacial periods, and is accompanied by changes in vegetation that imply drier conditions.

6.2 Inversion of marine terrace sequences

In the examples above, we showed how to assess paleo sea-level variations, and simultaneously extract quantified metrics for morphotectonics and hydrodynamics, from the geometry of marine terrace sequences. Using a probabilistic inversion methodology set in a Bayesian framework, we avoid the simplifications of bijective approaches in which a single marine terrace is always linked to a single sea-level highstand and vice-versa (e.g. Pastier et al., 2019; Malatesta et al., 2021). By considering a full sea-level curve and its possible variability, it is possible to provide quantitative constraints on highstands, lowstands, sea-level rise and fall, filling the observational gap for time periods for which field measurements are scarce. We admit that some model simplifications and approximations may alter our interpretations. In particular, we neglect subaerial erosion, and kept uplift rate, erosion rate, initial slope and wave base depth parameters time-constant for each individual sampled paleo sea level curve. Both could be fine-tuned in future developments.

To apply this model procedure to other sites, we can make several recommendations based on our findings in this paper. Any coastal area with marine terraces will have a lateral variability in terms of terrace width and terrace elevation. Given that the posterior ranges for the model parameters (paleo sea-level, uplift rate, etc.) will directly depend on the chosen inversion parameters (*ipstep*, σ and *corr1*), it is important to adjust those inversion parameters to the naturally observed variability in

473 the geometry of a marine terrace sequence. In the Santa Cruz example the similarity in terrace width/height variability
 474 between the modeled posterior ranges (Fig. S5), and the observed variability within a section of the marine terrace sequence
 475 (Fig. S6) justifies our choice of inversion parameters. But, if we wanted to characterize a larger area along the Santa Cruz
 476 coastline, we would either have to increase σ and/or *corr1* to account for the higher degree of variability, or jointly invert
 477 multiple cross-sections in which every cross-section has a similar variability in terrace height/width.

478 The prior information required to obtain reliable results, will also be dependent on the context of a specific marine terrace
 479 sequence, and on the parameter cut-off choices deemed realistic. For Santa Cruz and Corinth, the posterior distributions for
 480 wave base depth suggest that, purely based on marine terrace sequence morphology, this cut-off could have been deeper than
 481 10-12 m (Figs. 3 and 5). However, based on observations and models of cliff erosion it seems unlikely that the wave base for
 482 bedrock erosion is more than 10 m for Santa Cruz (Kline et al., 2014), and wave base depths/heights should be smaller for
 483 the calmer Gulf of Corinth. This choice in cut-off in turn affects the posterior distribution of other parameters, and should
 484 thus be chosen carefully.

485 In the cases of Corinth and Santa Cruz we assigned relatively narrow windows for the uplift rate, based on chronological
 486 information that was already available. As such we only obtained refinements, rather than ‘new’ information about the uplift
 487 rate. For the Santa Cruz case, we tested four uplift rate scenarios matching different possible chrono-stratigraphies (Fig. S7)
 488 that could all explain the morphology equally well, albeit with different ranges for the other parameters. We also inverted the
 489 Santa Cruz terrace sequence morphology with a broad range between 0.3 and 1.5 mm/yr, but the inversion would converge
 490 on only one of the four scenarios in Fig. S7. This suggests that at least an approximate prior idea on the uplift rate is a
 491 prerequisite for a reliable inversion of a marine terrace sequence morphology.

492 Many paleo sea-level studies that use geomorphic/geologic observations tend to have a confirmation bias regarding sea-level
 493 curves, and propose refinements of paleo sea-level estimates to sub-m scale (e.g. Murray-Wallace, 2002; Roberts et al., 2012)
 494 or uplift rates to precisions of ~0.01 mm/yr (e.g. Pedoja et al., 2018; Meschis et al., 2022). In this study, we take a step back
 495 by allowing more freedom to possible paleo sea-level variations, as well as uplift rate, erosion rate, initial slope and wave
 496 base depth, to provide a more reliable way to translate morphologic observations to paleo sea-level constraints. For instance,
 497 the low uplift rate examples from the Corinth Rift (Fig. 5b, c) and Santa Cruz (Fig. S7) reveal very little about paleo sea/lake
 498 levels, even if the uplift rate is roughly known. As a marine terrace is formed over several sea-level cycles, the resulting
 499 terrace width and height will depend on all those cycles, as well as wave base depths, erosion rates and initial slopes, all of
 500 which are generally poorly constrained. Even in the hypothetical case that these parameters are known (Fig. 2), there is still a
 501 wide spectrum of sea-level histories that could have created the specific morphology of a marine terrace sequence. It
 502 suggests that estimating paleo sea-level based on the comparison of a present-day landform to a paleo-landform (Rovere et
 503 al., 2016), may be too simplistic in many cases, at least for erosive marine terraces. Although uncertainties that we provide
 504 on paleo sea-level are much larger than what calculations based on hydrodynamic ranges would suggest (Lorchsteid and
 505 Rovere, 2019), we do consider them to be reliable as they take in a large number of unknowns. To reduce uncertainties, we
 506 demonstrated in this paper that a joint inversion of multiple profiles is a powerful way forward.

For simplicity, as the aim of this paper was to test the model on increasingly complex settings, we only inverted 1 profile for Santa Cruz and 3 for Corinth, but it is easily possible to explore with more profiles at those locations in future work. In addition, although here we focused on erosive marine terraces to develop a proof of concept, another promising avenue is to apply this inversion method to bio-constructed (coral reef) terraces, which tend to be better dated (e.g. Pedoja et al., 2014; Hibbert et al., 2016) and for which modelling routines also exist (e.g. Toomey et al., 2013; Pastier et al., 2019). One of our key findings is that inverting multiple profiles simultaneously provides much better paleo sea-level constraints than focusing on individual profiles (Figs. 2, 5). The global archive of paleo-shorelines (Fig. 1a) presents a huge potential for such multi-profile marine terrace inversions. This massive inversion would not only lead to improved estimates of local relative sea-level histories, but may also complement studies on glacio-isostatic adjustments that are relevant to a global sea-level perspective.

7 Conclusions

In this study we demonstrated the use of a probabilistic inversion approach to decipher the formation of marine terrace sequences in general, and the tectonic and hydro-climatic evolution of (semi-)isolated basins in particular. With this approach, we provide the tools (see below) to simultaneously estimate past sea-level variations, uplift rates, erosion rates, initial slopes and wave heights.

From synthetic tests, benchmarking on a terrace sequence near Santa Cruz marine terrace sequence, and application to the sequence in the SE Gulf of Corinth, our results bring a theoretical advance by showing that : 1) Paleo sea-level and other parameter ranges can be better constrained from sequences that are uplifting at higher rates compared to lower rates, and better constrained from a joint inversion of multiple profiles than from inversion of a single profile. 2) Uplift rates, sea-level variations and wave erosion parameters are intricately linked. By allowing more freedom to possible ranges of all the relevant parameters, we provide a more reliable way to translate morphologic observations to paleo sea-level constraints. Resulting uncertainties may be higher compared to ‘classic’ approaches of comparing present to past shoreline elevations, but are more realistic. 3) Probabilistic inversion of marine terrace sequences is a powerful method, applicable to a large portion of the world’s coastlines to disentangle tectonic and hydro-climatic processes.

Beyond the methodological achievement, by applying our method to a complex case -the semi-isolated Gulf of Corinth (Greece), we show that this method can be a powerful tool to explore subtle environmental forcings, like the balance between precipitation and evaporation, that may have had a prime importance in setting the lake level during certain periods of time. We found that eustatic sea-level and tectonically changing sill depths drive marine and transitional phases during interglacial and interstadial periods, respectively. Wetter and drier conditions drive over- and underfilling of Lake Corinth during interstadial and glacial periods, respectively. We expect such transitions to be different for each unique tectono-hydro-climatic setting, with our inversion approach providing a new way to decipher such geomorphic Rosetta stones.

539 **Code availability**

540 The marine terrace inversion code used in this study can be found at <https://github.com/ginodegelder/Rosetta>.

541 **Supplement link**

542 The supplementary material to this study can be found at <https://doi.org/10.31223/X5B117>.

543 **Author contribution**

544 **Conceptualization** by GdG, LH and TB, **Formal analysis** by GdG, **Funding acquisition** by TB, **Investigation** by GdG,
545 NH, LH and TB, **Methodology** by GdG, NH, AMP, TB and YB, **Visualization** by GdG, NH and YB, **Writing - original**
546 **draft** by GdG, **Writing - review & editing** by all authors.

547 **Competing interests**

548 The authors declare that they have no conflict of interest.

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556 **References**

557 Aksu, A. E., Hiscott, R. N., & Dogan, Y. (1999). Oscillating Quaternary water levels of the Marmara Sea and vigorous
558 outflow into the Aegean Sea from the Marmara Sea–Black Sea drainage corridor. *Marine Geology*, 153(1-4), 275-302.
559 Andersen, M. B., Stirling, C. H., Potter, E. K., Halliday, A. N., Blake, S. G., McCulloch, M. T., ... & O’ Leary, M. J. (2010).
560 The timing of sea-level high-stands during Marine Isotope Stages 7.5 and 9: Constraints from the uranium-series dating of
561 fossil corals from Henderson Island. *Geochimica et Cosmochimica Acta*, 74(12), 3598-3620.
562 Anderson, R.S., Densmore, A.L., Ellis, M.A., 1999. The generation and degradation of marine terraces. *Basin Res.* 11 (1),
563 7e19.

Anderson, R. S., & Menking, K. M. (1994). The Quaternary marine terraces of Santa Cruz, California: Evidence for coseismic uplift on two faults. *Geological Society of America Bulletin*, 106(5), 649-664.

Armijo, R., Meyer, B. G. C. P., King, G. C. P., Rigo, A., & Papanastassiou, D. (1996). Quaternary evolution of the Corinth Rift and its implications for the Late Cenozoic evolution of the Aegean. *Geophysical Journal International*, 126(1), 11-53.

Austermann, J., Mitrovica, J. X., Huybers, P., & Rovere, A. (2017). Detection of a dynamic topography signal in last interglacial sea-level records. *Science Advances*, 3(7), e1700457.

Batchelor, C. L., Margold, M., Krapp, M., Murton, D. K., Dalton, A. S., Gibbard, P. L., ... & Manica, A. (2019). The configuration of Northern Hemisphere ice sheets through the Quaternary. *Nature communications*, 10(1), 3713.

Beckers, A., Beck, C., Hubert-Ferrari, A., Tripsanas, E., Crouzet, C., Sakellariou, D., ... & De Batist, M. (2016). Influence of bottom currents on the sedimentary processes at the western tip of the Gulf of Corinth, Greece. *Marine Geology*, 378, 312-332.

Bernard, P., Lyon-Caen, H., Briole, P., Deschamps, A., Boudin, F., Makropoulos, K., ... & Linde, A. (2006). Seismicity, deformation and seismic hazard in the western rift of Corinth: New insights from the Corinth Rift Laboratory (CRL). *Tectonophysics*, 426(1-2), 7-30.

Bloom, A. L., & Yonekura, N. (1990). Graphic analysis of dislocated Quaternary shorelines. *Sea-level change*, 104-115.

Bradley, W. C. (1957). Origin of marine-terrace deposits in the Santa Cruz area, California. *Geological Society of America Bulletin*, 68(4), 421-444.

Bradley, W. C., & Addicott, W. O. (1968). Age of first marine terrace near Santa Cruz, California. *Geological Society of America Bulletin*, 79(9), 1203-1210.

Brown, E. T., & Bourles, D. L. (2002). Use of a new ^{10}Be and ^{26}Al inventory method to date marine terraces, Santa Cruz, California, USA: Comment and Reply: COMMENT. *Geology*, 30(12), 1147-1148.

Caterina, B., Rubi, R., & Hubert-Ferrari, A. (2023). Stratigraphic architecture, sedimentology and structure of the Middle Pleistocene Corinth Canal (Greece). *Geological Society, London, Special Publications*, 523(1), 279-304.

Chappell, J., & Shackleton, N. (1986). Oxygen isotopes and sea level. *Nature*, 324(6093), 137-140.

Chauveau, D., Pastier, A.-M., de Gelder, G., Husson, L., Authemayou, C., Pedoja, K. et al. (2024) Unravelling the morphogenesis of coastal terraces at Cape Laundi (Sumba Island, Indonesia): Insights from numerical models. *Earth Surface Processes and Landforms*, 49(2), 549–566.

Collier, R. L., & Dart, C. J. (1991). Neogene to Quaternary rifting, sedimentation and uplift in the Corinth Basin, Greece. *Journal of the Geological Society*, 148(6), 1049-1065.

Collier, R. E., Leeder, M. R., Trout, M., Ferentinos, G., Lyberis, E., & Papatheodorou, G. (2000). High sediment yields and cool, wet winters: Test of last glacial paleoclimates in the northern Mediterranean. *Geology*, 28(11), 999-1002.

Creveling, J. R., Mitrovica, J. X., Clark, P. U., Waelbroeck, C., & Pico, T. (2017). Predicted bounds on peak global mean sea level during marine isotope stages 5a and 5c. *Quaternary Science Reviews*, 163, 193-208.

597 Dalton, A. S., Finkelstein, S. A., Forman, S. L., Barnett, P. J., Pico, T., & Mitrovica, J. X. (2019). Was the Laurentide Ice
598 Sheet significantly reduced during marine isotope stage 3?. *Geology*, 47(2), 111-114.

599 Dalton, A. S., Pico, T., Gowan, E. J., Clague, J. J., Forman, S. L., McMartin, I., ... & Helmens, K. F. (2022). The marine
600 $\delta^{18}\text{O}$ record overestimates continental ice volume during Marine Isotope Stage 3. *Global and Planetary Change*, 212,
601 103814.

602 De Gelder, G., Fernández-Blanco, D., Melnick, D., Duclaux, G., Bell, R. E., Jara-Muñoz, J., ... & Lacassin, R. (2019).
603 Lithospheric flexure and rheology determined by climate cycle markers in the Corinth Rift. *Scientific Reports*, 9(1), 4260.

604 De Gelder, G., Jara-Munoz, J., Melnick, D., Fernández-Blanco, D., Rouby, H., Pedoja, K., ... & Lacassin, R. (2020). How do
605 sea-level curves influence modeled marine terrace sequences?. *Quaternary Science Reviews*, 229, 106132.

606 De Gelder, G., Husson, L., Pastier, A. M., Fernández-Blanco, D., Pico, T., Chauveau, D., ... & Pedoja, K. (2022). High
607 interstadial sea levels over the past 420ka from the Huon Peninsula, Papua New Guinea. *Communications Earth &*
608 *Environment*, 3(1), 256.

609 De Gelder, G., Solihuddin, T., Utami, D. A., Hendrizan, M., Rachmayani, R., Chauveau, D., ... & Cahyarini, S. Y. (2023).
610 Geodynamic control on Pleistocene coral reef development: insights from northwest Sumba Island (Indonesia). *Earth*
611 *Surface Processes and Landforms*, 48(13), 2536-2553.

612 Derricourt, R. (2005). Getting “Out of Africa”: sea crossings, land crossings and culture in the hominin migrations. *Journal*
613 *of world prehistory*, 19, 119-132.

614 Dutton, A., Carlson, A. E., Long, A. J., Milne, G. A., Clark, P. U., DeConto, R., ... & Raymo, M. E. (2015). Sea-level rise
615 due to polar ice-sheet mass loss during past warm periods. *science*, 349(6244), aaa4019.

616 Dyer, B., Austermann, J., D’ Andrea, W. J., Creel, R. C., Sandstrom, M. R., Cashman, M., ... & Raymo, M. E. (2021). Sea-
617 level trends across The Bahamas constrain peak last interglacial ice melt. *Proceedings of the National Academy of*
618 *Sciences*, 118(33), e2026839118.

619 Fernández-Blanco, D., de Gelder, G., Lacassin, R., & Armijo, R. (2019). A new crustal fault formed the modern Corinth
620 Rift. *Earth-Science Reviews*, 199, 102919.

621 Gallagher, K., Charvin, K., Nielsen, S., Sambridge, M., & Stephenson, J. (2009). Markov chain Monte Carlo (MCMC)
622 sampling methods to determine optimal models, model resolution and model choice for Earth Science problems. *Marine and*
623 *Petroleum Geology*, 26(4), 525-535.

624 Gallen, S. F., & Fernández - Blanco, D. (2021). A new data - driven Bayesian inversion of fluvial topography clarifies the
625 tectonic history of the corinth rift and reveals a channel steepness threshold. *Journal of Geophysical Research: Earth*
626 *Surface*, 126(3), e2020JF005651.

627 Gawthorpe, R. L., Fabregas, N., Pechlivanidou, S., Ford, M., Collier, R. E. L., Carter, G. D., ... & Shillington, D. J. (2022).
628 Late Quaternary mud - dominated, basin - floor sedimentation of the Gulf of Corinth, Greece: Implications for deep - water
629 depositional processes and controls on syn - rift sedimentation. *Basin Research*, 34(5), 1567-1600.

Gowan, E. J., Zhang, X., Khosravi, S., Rovere, A., Stocchi, P., Hughes, A. L., ... & Lohmann, G. (2021). A new global ice sheet reconstruction for the past 80 000 years. *Nature communications*, 12(1), 1199.

Guilcher, A., 1974. Les «rasas»: un problème de morphologie littorale générale. *Annales de Géographie*, 83, 1–33.

Hay, C., Mitrovica, J. X., Gomez, N., Creveling, J. R., Austermann, J., & Kopp, R. E. (2014). The sea-level fingerprints of ice-sheet collapse during interglacial periods. *Quaternary Science Reviews*, 87, 60-69.

Hibbert, F. D., Rohling, E. J., Dutton, A., Williams, F. H., Chutcharavan, P. M., Zhao, C., & Tamisiea, M. E. (2016). Coral indicators of past sea-level change: A global repository of U-series dated benchmarks. *Quaternary Science Reviews*, 145, 1-56.

Houghton, S. L., Roberts, G. P., Papanikolaou, I. D., McArthur, J. M., & Gilmour, M. A. (2003). New ²³⁴U - ²³⁰Th coral dates from the western Gulf of Corinth: Implications for extensional tectonics. *Geophysical Research Letters*, 30(19).

Husson, L., Pastier, A.-M., Pedoja, K., Elliot, M., Paillard, D., Authemayou, C., et al., 2018. Reef carbonate productivity during quaternary sea level oscillations. *Geochem. Geophys. Geosyst.* 19 (4), 1148e1164.

Jara-Muñoz, J., Melnick, D., Pedoja, K., & Strecker, M. R. (2019). TerraceM-2: A Matlab® interface for mapping and modeling marine and lacustrine terraces. *Frontiers in Earth Science*, 255.

Johnson, M. E., & Libbey, L. K. (1997). Global review of upper Pleistocene (substage 5e) rocky shores: tectonic segregation, substrate variation, and biological diversity. *Journal of Coastal Research*, 297-307.

Kafetzidou, A., Fatourou, E., Panagiotopoulos, K., Marret, F., & Kouli, K. (2023). Vegetation Composition in a Typical Mediterranean Setting (Gulf of Corinth, Greece) during Successive Quaternary Climatic Cycles. *Quaternary*, 6(2), 30.

Kennedy, G. L., Lajoie, K. R., & Wehmiller, J. F. (1982). Aminostratigraphy and faunal correlations of late Quaternary marine terraces, Pacific Coast, USA. *Nature*, 299(5883), 545-547.

Kline, S. W., Adams, P. N., & Limber, P. W. (2014). The unsteady nature of sea cliff retreat due to mechanical abrasion, failure and comminution feedbacks. *Geomorphology*, 219, 53-67.

Kopp, R. E., Simons, F. J., Mitrovica, J. X., Maloof, A. C., & Oppenheimer, M. (2009). Probabilistic assessment of sea level during the last interglacial stage. *Nature*, 462(7275), 863-867.

Landmann, G., Reimer, A., & Kempe, S. (1996). Climatically induced lake level changes at Lake Van, Turkey, during the Pleistocene/Holocene transition. *Global Biogeochemical Cycles*, 10(4), 797-808.

Lajoie, K. R., Wehmiller, J. F., Kvenvolden, K. A., Peterson, E., & White, R. H. (1975). Correlation of California marine terraces by amino acid stereochemistry. In *Geological Society of America Abstracts with Programs* (Vol. 7, No. 3, pp. 338-339).

Lajoie, K.R., 1986. Coastal tectonics. In: Press, N.A. (Ed.), *Active Tectonics*. National Academic Press, Washington DC, pp. 95e124.

Lambeck, K., & Chappell, J. (2001). Sea level change through the last glacial cycle. *Science*, 292(5517), 679-686.

662 Leclerc, F., & Feuillet, N. (2019). Quaternary coral reef complexes as powerful markers of long-term subsidence related to
 663 deep processes at subduction zones: Insights from Les Saintes (Guadeloupe, French West Indies). *Geosphere*, 15(4), 983-
 664 1007.

665 Lorscheid, T., & Rovere, A. (2019). The indicative meaning calculator—quantification of paleo sea-level relationships by
 666 using global wave and tide datasets. *Open Geospatial Data, Software and Standards*, 4, 1-8.

667 Malatesta, L. C., Finnegan, N. J., Huppert, K. L., & Carreño, E. I. (2022). The influence of rock uplift rate on the formation
 668 and preservation of individual marine terraces during multiple sea-level stands. *Geology*, 50(1), 101-105.

669 Marra, F., Sevink, J., Tolomei, C., Vannoli, P., Florindo, F., Jicha, B. R., & La Rosa, M. (2023). New age constraints on the
 670 MIS 9–MIS 5.3 marine terraces of the Pontine Plain (central Italy) and implications for global sea levels. *Quaternary Science*
 671 *Reviews*, 300, 107866.

672 Matsumoto, H., Young, A. P., & Carilli, J. E. (2022). Modeling the relative influence of environmental controls on marine
 673 terrace widths. *Geomorphology*, 396, 107986.

674 McNeill, L. C., Shillington, D. J., Carter, G. D., Everest, J. D., Gawthorpe, R. L., Miller, C., ... & Green, S. (2019). High-
 675 resolution record reveals climate-driven environmental and sedimentary changes in an active rift. *Scientific Reports*, 9(1),
 676 3116.

677 Medina-Elizalde, M. (2013). A global compilation of coral sea-level benchmarks: implications and new challenges. *Earth*
 678 *and Planetary Science Letters*, 362, 310-318.

679 Meschis, M., Roberts, G. P., Robertson, J., Mildon, Z. K., Sahy, D., Goswami, R., ... & Iezzi, F. (2022). Out of phase
 680 Quaternary uplift-rate changes reveal normal fault interaction, implied by deformed marine
 681 palaeoshorelines. *Geomorphology*, 416, 108432.

682 Mosegaard, K., & Sambridge, M. (2002). Monte Carlo analysis of inverse problems. *Inverse problems*, 18(3), R29.

683 Murray - Wallace, C. V. (2002). Pleistocene coastal stratigraphy, sea - level highstands and neotectonism of the southern
 684 Australian passive continental margin—a review. *Journal of Quaternary Science: Published for the Quaternary Research*
 685 *Association*, 17(5 - 6), 469-489.

686 Nixon, C. W., McNeill, L. C., Bull, J. M., Bell, R. E., Gawthorpe, R. L., Henstock, T. J., ... & Kranis, H. (2016). Rapid
 687 spatiotemporal variations in rift structure during development of the Corinth Rift, central Greece. *Tectonics*, 35(5), 1225-
 688 1248.

689 Ott, R. F., Gallen, S. F., Wegmann, K. W., Biswas, R. H., Herman, F., & Willett, S. D. (2019). Pleistocene terrace formation,
 690 Quaternary rock uplift rates and geodynamics of the Hellenic Subduction Zone revealed from dating of paleoshorelines on
 691 Crete, Greece. *Earth and Planetary Science Letters*, 525, 115757.

692 Pastier, A.-M., Husson, L., Pedoja, K., Bezos, A., Authemayou, C., Arias-Ruiz, C., Cahyarini, S.Y. (2019). Genesis and
 693 architecture of sequences of quaternary coral reef terraces: Insights from numerical models. *Geochem. Geophys. Geosyst.*, 20
 694 (8), 4248e4272.

695 Pedoja, K., Husson, L., Regard, V., Cobbold, P. R., Ostanciaux, E., Johnson, M. E., ... & Delcaillau, B. (2011). Relative sea-
696 level fall since the last interglacial stage: are coasts uplifting worldwide?. *Earth-Science Reviews*, 108(1-2), 1-15.

697 Pedoja, K., Husson, L., Johnson, M. E., Melnick, D., Witt, C., Pochat, S., ... & Garestier, F. (2014). Coastal staircase
698 sequences reflecting sea-level oscillations and tectonic uplift during the Quaternary and Neogene. *Earth-Science*
699 *Reviews*, 132, 13-38.

700 Pedoja, K., Jara-Muñoz, J., De Gelder, G., Robertson, J., Meschis, M., Fernández-Blanco, D., ... & Pinel, B. (2018).
701 Neogene-Quaternary slow coastal uplift of Western Europe through the perspective of sequences of strandlines from the
702 Cotentin Peninsula (Normandy, France). *Geomorphology*, 303, 338-356.

703 Perg, L. A., Anderson, R. S., & Finkel, R. C. (2001). Use of a new ^{10}Be and ^{26}Al inventory method to date marine terraces,
704 Santa Cruz, California, USA. *Geology*, 29(10), 879-882.

705 Perissoratis, C., Piper, D. J. W., & Lykousis, V. (2000). Alternating marine and lacustrine sedimentation during late
706 Quaternary in the Gulf of Corinth rift basin, central Greece. *Marine Geology*, 167(3-4), 391-411.

707 Pico, T., Mitrovica, J. X., Ferrier, K. L., & Braun, J. (2016). Global ice volume during MIS 3 inferred from a sea-level
708 analysis of sedimentary core records in the Yellow River Delta. *Quaternary Science Reviews*, 152, 72-79.

709 Pirazzoli, P.A., 2005. Marine terraces. In: Schwartz, M.L. (Ed.), *Encyclopedia of Coastal Science*. Springer Netherlands,
710 Dordrecht, pp. 632-633.

711 Railsback, L. B., Gibbard, P. L., Head, M. J., Voarintsoa, N. R. G., & Toucanne, S. (2015). An optimized scheme of lettered
712 marine isotope substages for the last 1.0 million years, and the climatostratigraphic nature of isotope stages and
713 substages. *Quaternary Science Reviews*, 111, 94-106.

714 Regard, V., Pedoja, K., De La Torre, I., Saillard, M., Corte_s-Aranda, J., Nexer, M., 2017. Geometrical trends within
715 sequences of Pleistocene marine terraces: selected examples from California, Peru, Chile and New-Zealand. *Zeitschrift Fur*
716 *Geomorphologie* 61 (1), 53e73.

717 Roberts, G. P., Houghton, S. L., Underwood, C., Papanikolaou, I., Cowie, P. A., van Calsteren, P., ... & McArthur, J. M.
718 (2009). Localization of Quaternary slip rates in an active rift in 105 years: An example from central Greece constrained by
719 ^{234}U - ^{230}Th coral dates from uplifted paleoshorelines. *Journal of Geophysical Research: Solid Earth*, 114(B10).

720 Roberts, D. L., Karkanas, P., Jacobs, Z., Marean, C. W., & Roberts, R. G. (2012). Melting ice sheets 400,000 yr ago raised
721 sea level by 13 m: Past analogue for future trends. *Earth and Planetary Science Letters*, 357, 226-237.

722 Roeser, P., Franz, S. O., & Litt, T. (2016). Aragonite and calcite preservation in sediments from Lake Iznik related to bottom
723 lake oxygenation and water column depth. *Sedimentology*, 63(7), 2253-2277.

724 Rosenbloom, N. A., & Anderson, R. S. (1994). Hillslope and channel evolution in a marine terraced landscape, Santa Cruz,
725 California. *Journal of Geophysical Research: Solid Earth*, 99(B7), 14013-14029.

726 Rovere, A., Raymo, M. E., Vacchi, M., Lorscheid, T., Stocchi, P., Gomez-Pujol, L., ... & Hearty, P. J. (2016). The analysis
727 of Last Interglacial (MIS 5e) relative sea-level indicators: Reconstructing sea-level in a warmer world. *Earth-Science*
728 *Reviews*, 159, 404-427.

729 Rovere, A., Ryan, D. D., Vacchi, M., Dutton, A., Simms, A. R., & Murray-Wallace, C. V. (2023). The World Atlas of Last
 730 Interglacial Shorelines (version 1.0). *Earth System Science Data*, 15(1), 1-23.

731 Schellmann, G., & Radtke, U. (2004). A revised morpho-and chronostratigraphy of the Late and Middle Pleistocene coral
 732 reef terraces on Southern Barbados (West Indies). *Earth-Science Reviews*, 64(3-4), 157-187.

733 Shakun, J. D., Lea, D. W., Lisiecki, L. E., & Raymo, M. E. (2015). An 800-kyr record of global surface ocean $\delta^{18}\text{O}$ and
 734 implications for ice volume-temperature coupling. *Earth and Planetary Science Letters*, 426, 58-68.

735 Siddall, M., Smeed, D. A., Hemleben, C., Rohling, E. J., Schmelzer, I., & Peltier, W. R. (2004). Understanding the Red Sea
 736 response to sea level. *Earth and Planetary Science Letters*, 225(3-4), 421-434.

737 Sonđi, I., & Juraćić, M. (2010). Whiting events and the formation of aragonite in Mediterranean Karstic Marine Lakes: new
 738 evidence on its biologically induced inorganic origin. *Sedimentology*, 57(1), 85-95.

739 Spratt, R. M., & Lisiecki, L. E. (2016). A Late Pleistocene sea level stack. *Climate of the Past*, 12(4), 1079-1092.

740 Stein, M., Torfstein, A., Gavrieli, I., & Yechieli, Y. (2010). Abrupt aridities and salt deposition in the post-glacial Dead Sea
 741 and their North Atlantic connection. *Quaternary Science Reviews*, 29(3-4), 567-575.

742 Stirling, C. H., Esat, T. M., Lambeck, K., McCulloch, M. T., Blake, S. G., Lee, D. C., & Halliday, A. N. (2001). Orbital
 743 forcing of the marine isotope stage 9 interglacial. *Science*, 291(5502), 290-293.

744 Strobl, M., Hetzel, R., Fassoulas, C., & Kubik, P. W. (2014). A long-term rock uplift rate for eastern Crete and geodynamic
 745 implications for the Hellenic subduction zone. *Journal of Geodynamics*, 78, 21-31.

746 Sunamura, T. (1992). *Geomorphology of Rocky Coasts*, vol. 3. John Wiley & Son Ltd.

747 Tawil-Morsink, K., Austermann, J., Dyer, B., Dumitru, O. A., Precht, W. F., Cashman, M., ... & Raymo, M. E. (2022).
 748 Probabilistic investigation of global mean sea level during MIS 5a based on observations from Cave Hill,
 749 Barbados. *Quaternary Science Reviews*, 295, 107783.

750 Toomey, M., Ashton, A. D., & Perron, J. T. (2013). Profiles of ocean island coral reefs controlled by sea-level history and
 751 carbonate accumulation rates. *Geology*, 41(7), 731-734.

752 Van Daele, M., van Welden, A., Moernaut, J., Beck, C., Audemard, F., Sanchez, J., ... & De Batist, M. (2011).
 753 Reconstruction of Late-Quaternary sea-and lake-level changes in a tectonically active marginal basin using seismic
 754 stratigraphy: The Gulf of Cariaco, NE Venezuela. *Marine Geology*, 279(1-4), 37-51.

755 Waelbroeck, C., Labeyrie, L., Michel, E., Duplessy, J. C., Mcmanus, J. F., Lambeck, K., ... & Labracherie, M. (2002). Sea-
 756 level and deep water temperature changes derived from benthic foraminifera isotopic records. *Quaternary science
 757 reviews*, 21(1-3), 295-305.

758 Watkins, S. E., Whittaker, A. C., Bell, R. E., McNeill, L. C., Gawthorpe, R. L., Brooke, S. A., & Nixon, C. W. (2019). Are
 759 landscapes buffered to high-frequency climate change? A comparison of sediment fluxes and depositional volumes in the
 760 Corinth Rift, central Greece, over the past 130 ky. *GSA Bulletin*, 131(3-4), 372-388.

761 Weber, G.E. (1990). Late Pleistocene slip rates on the San Gregorio fault zone at Point Ano Nuevo, San Mateo County,
762 California, in Garrison, R.E., et al., eds., *Geology and tectonics of coastal California, San Francisco to Monterey* (volume
763 and guidebook): Bakersfield, California, Pacific Section, American Association of Petroleum Geologists, p. 193–203.
764 Webster, J. M., Wallace, L. M., Clague, D. A., & Braga, J. C. (2007). Numerical modeling of the growth and drowning of
765 Hawaiian coral reefs during the last two glacial cycles (0–250 kyr). *Geochemistry, Geophysics, Geosystems*, 8(3).