Continental-Scale Surface Water Distribution:

Coupled groundwater and surface water modelling to visualise lake extent and total terrestrial water storage under a changing climate

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Abstract

Large-scale flow-routing algorithms efficiently route water to the ocean, neglecting both inland basins that may be able to form lakes, and changing groundwater storage. We add these elements of reality in a simplified and computationally-efficient way, combining groundwater and surface-water routing to simulate changing groundwater levels, surface-water flow pathways, and lake locations and extents through time. The groundwater component is based upon a linear-diffusive model for an unconfined aquifer developed by Reinfelder et al (2013), and surface water is routed through a simple downslope-flow algorithm that differs from most flow-routing algorithms in that it takes into account the elevation of the water surface, and not just the land surface. Our model requires as inputs topography, climatic data (P-ET and winter temperature), and an approximation of hydraulic conductivity based on topographic slope and mapped soils. The model outputs grids of depth to water table and thickness of surface water; the latter depicts any lakes that would form under the topographic and climatic conditions. The model can be run to equilibrium, or, if a starting depth to water table input is provided, for any user-selected length of time. Such solutions are transient only with respect to groundwater movement: surface-water flow is significantly faster, so it is always run to equilibrium. The model allows infiltration when surface water flows across cells that are not fully saturated in the groundwater, and it allows exfiltration and the formation of groundwater-fed lakes where convergent groundwater flow raises the water table above the land surface. We show sample results from this model on continental test regions. Future work using this model will include global runs since the Last Glacial Maximum, with ground truthing possible using past lake shoreline data. Changing depth to water table plus the surface water storage computed using this model allows computation of changing terrestrial water storage volume through time.

Groundwater Component

The model comprises two separate components: groundwater and surface water. The groundwater component has been adapted from Fan Reinfelder et al (2013), whose model has proven successful in estimation of groundwater levels on a global scale. Groundwater moves down hydraulic gradient and exfiltrates when the water table rises above the land surface (Figure 1).

Inputs:

Climate data - we have used TraCE-21K model runs and are experimenting with HADCM3 as an alternative. We use both winter temperature and recharge as P-ET + icemelt (from ICE-6G).

Topography - changing topography based on the ICE6G/VM-5a ice and Earth rheology models. Slope also plays a role in the model calculations (deeper water tables under steeper slopes).

Soil properties - we use broad soil classes as an estimator for regional hydraulic conductivity.

Outputs:

Depth to water table. While we do not get the absolute volume of groundwater storage, running the model at multiple dates allows us to calculate change in groundwater storage through time.

This groundwater model comes with a caveat: it does not include a surface water component, and in fact, all surface water is assumed to evaporate and/or run off immediately. While this may be realistic for rivers on our time scales, this ignores the existence of lakes, including the large proglacial and pluvial lakes that are known to have existed in the past. Any lakes which were groundwater-fed would be a major water sink in this model design, artificially lowering the surrounding water table to a potentially large degree.

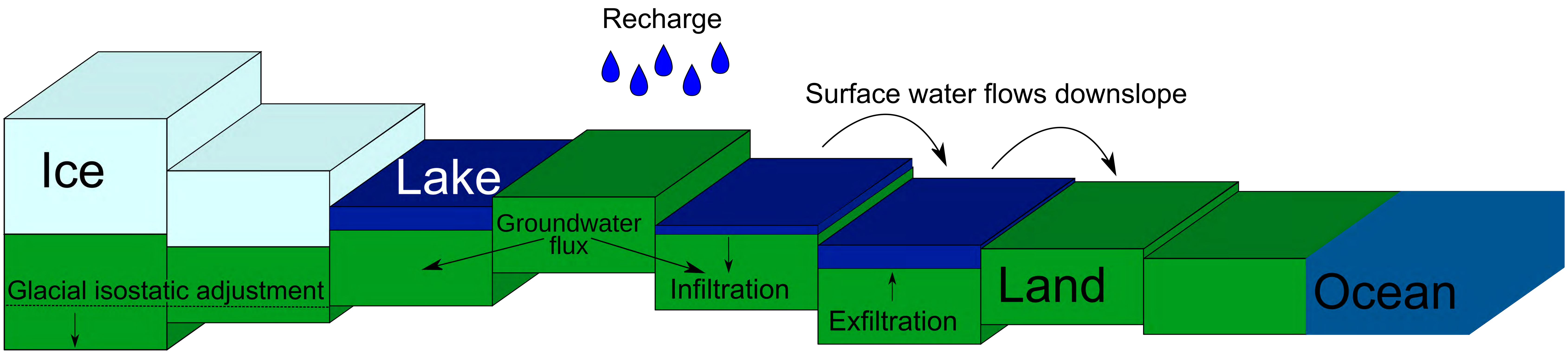

Figure 1: Model schematic showing both the groundwater and surface water components of the model. Inputs are recharge and topography, adjusted according to glacial isostatic adjustment and changing sea level. Groundwater moves down hydraulic gradient and exfiltrates when the water table rises above the land surface in a given cell. Surface water flows downslope and infiltrates when it enters a cell that is not fully saturated in the groundwater. Lakes form in local basins, and additional water runs off to the ocean. Groundwater levels fluctuate with changing recharge and topography through time.

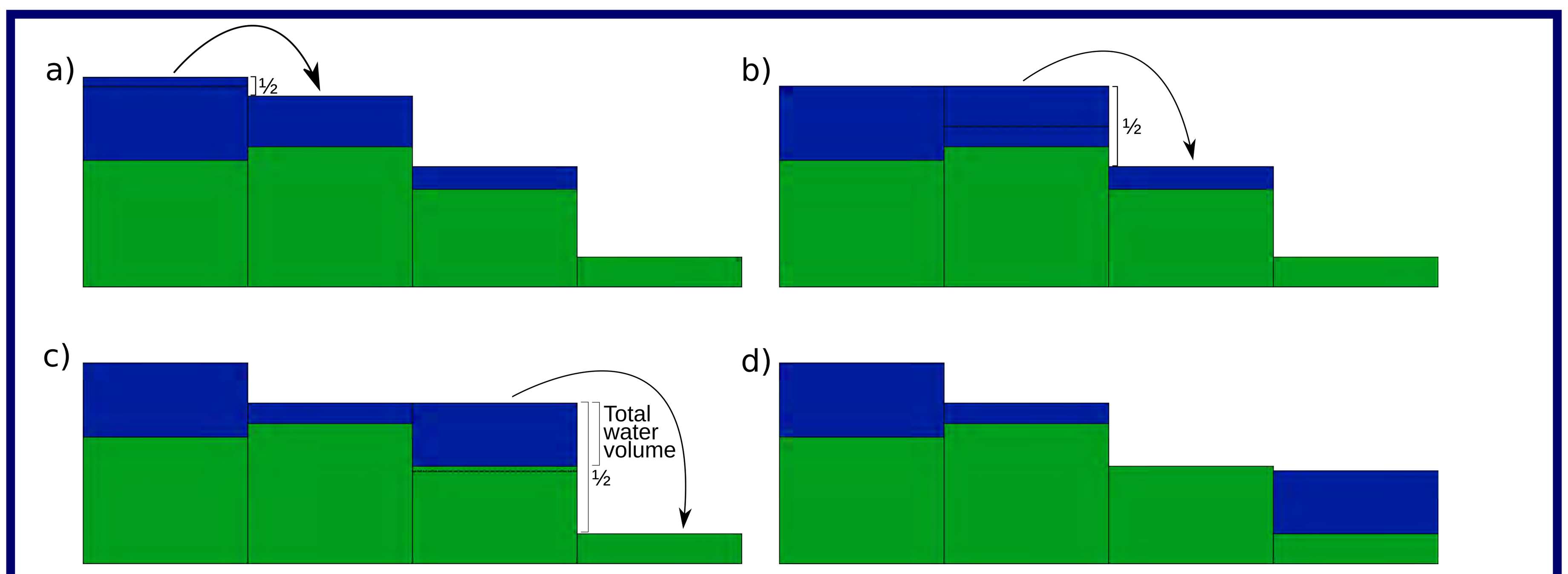

Figure 2: A visual representation of the surface water algorithm. Cell by cell, water is moved downslope, where slope includes both land and water. The steepest direction is selected, and then max(half the difference between the two cells, all the water in the cell) is moved to the downslope cell. From a to d, one iteration through the surface water array is shown. In the next iteration, water will continue to move downslope in the same way until basins are filled and ridges have exported all of their surface water.

Surface Water Component

In order to correctly model depths to water table through time, and to compute changing total terrestrial water storage on a global scale, we add a surface water component to accommodate the existence of groundwater-fed lakes. Pre-existing large-scale flow-routing algorithms efficiently route water to the ocean, neglecting inland basins that may be able to form lakes. We add this element in a simplified way, suitable for large-scale model runs (Figure 1).

Inputs:

Topography - the same used for the groundwater portion of the model (ICE6G for isostatic adjustment).

Starting depth to water table and surface water thickness - obtained from the groundwater component of the model.

Outputs:

New depths to water table and surface water thickness, to be fed back into the groundwater component.

The surface water model component works by simple down-slope flow of water, but takes into account the thickness of the water itself as well as the topography (Figure 2). Water infiltrates and changes the depth to groundwater whenever it enters a cell in which the water table is not at the land surface; conversely, water is able to exfiltrate in the groundwater component of the model. Due to the length of the timescales we are interested in (most model runs are performed for 500 years, with one month being the shortest step taken in any model run), we assume that surface water reaches equilibrium every time this model component is used.

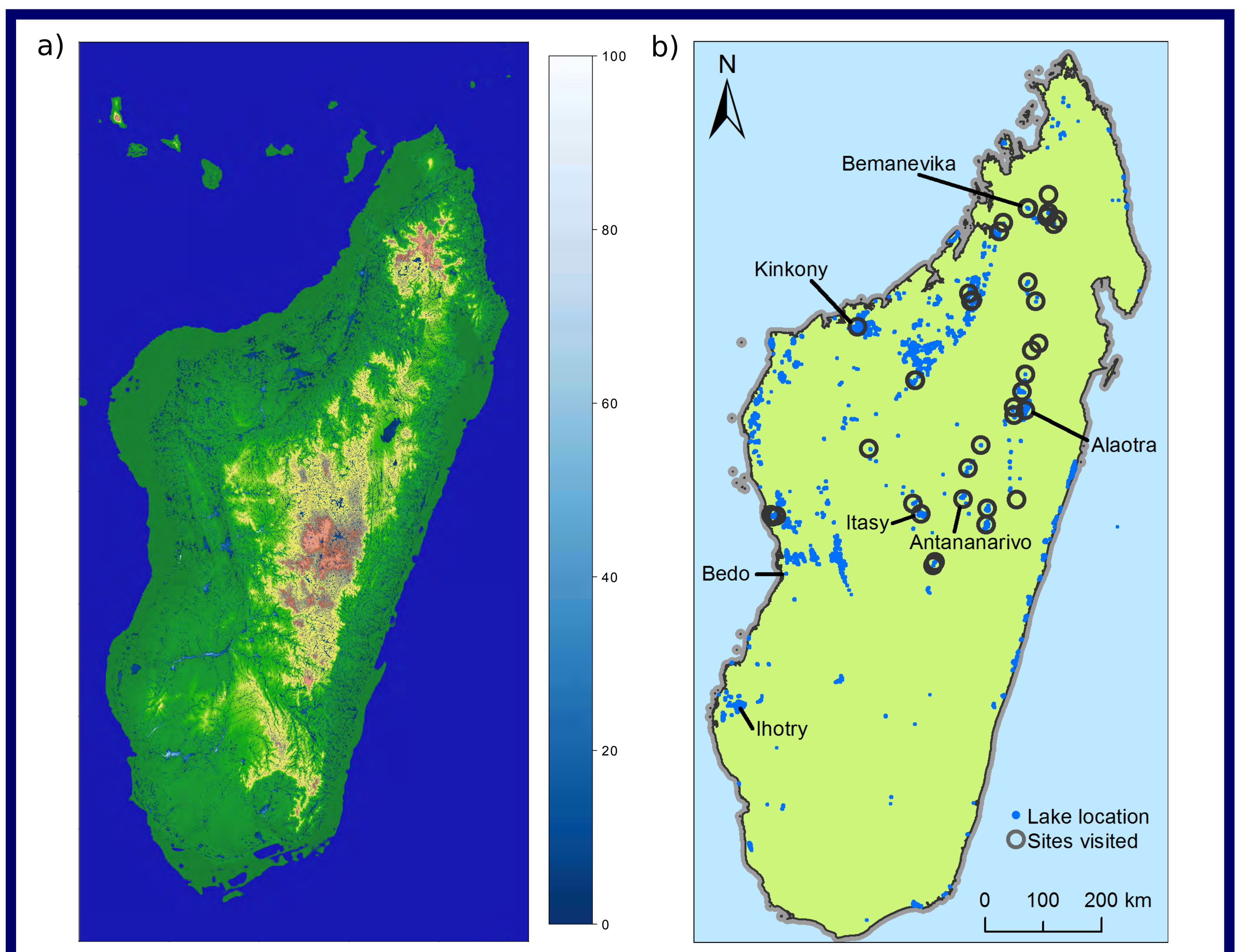

Figure 3: a) Results of the surface water component of the model for modern-day Madagascar (metres of water). Madagascar is a useful test case since it is a relatively small land mass surrounded by ocean. This figure shows the result of a scenario where the ground is impermeable (no infiltration) and starts with a metre of surface water everywhere. This results in an excess of surface water compared to reality, but makes all of the island's enclosed basins clearly visible. b) some of Madagascar's lakes, from Bamford et al (2017).

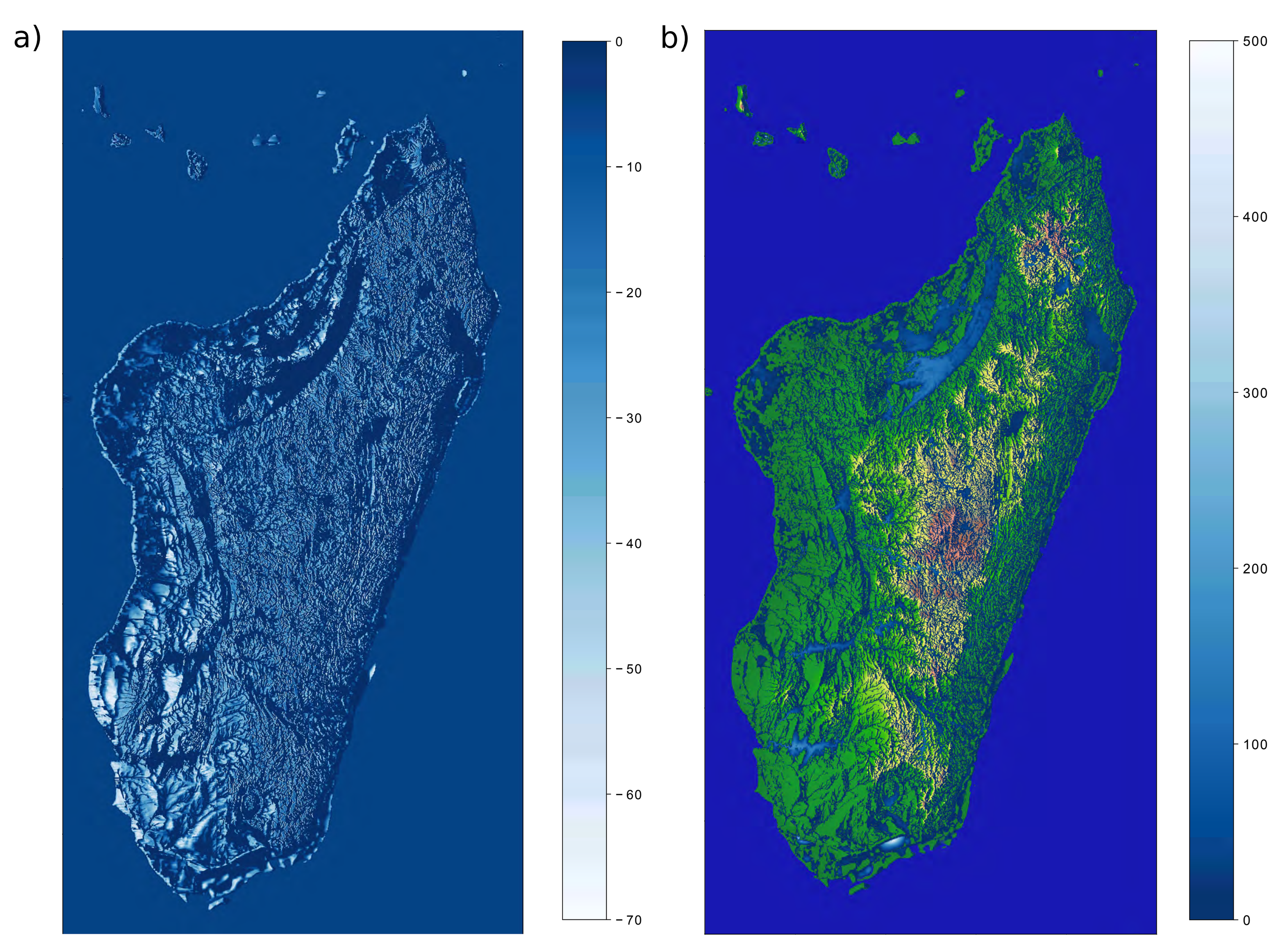

Figure 4: First attempts at model coupling. The coupled model is still under construction, but some preliminary results have been obtained. This test model run for LGM Madagascar (21 ka) assumed a starting depth to water table of 5 m throughout Madagascar. Future transient model runs will be started with a more realistic water table depth obtained from an equilibrium model run. a) depth to groundwater in metres, and b) thickness of surface water in metres.

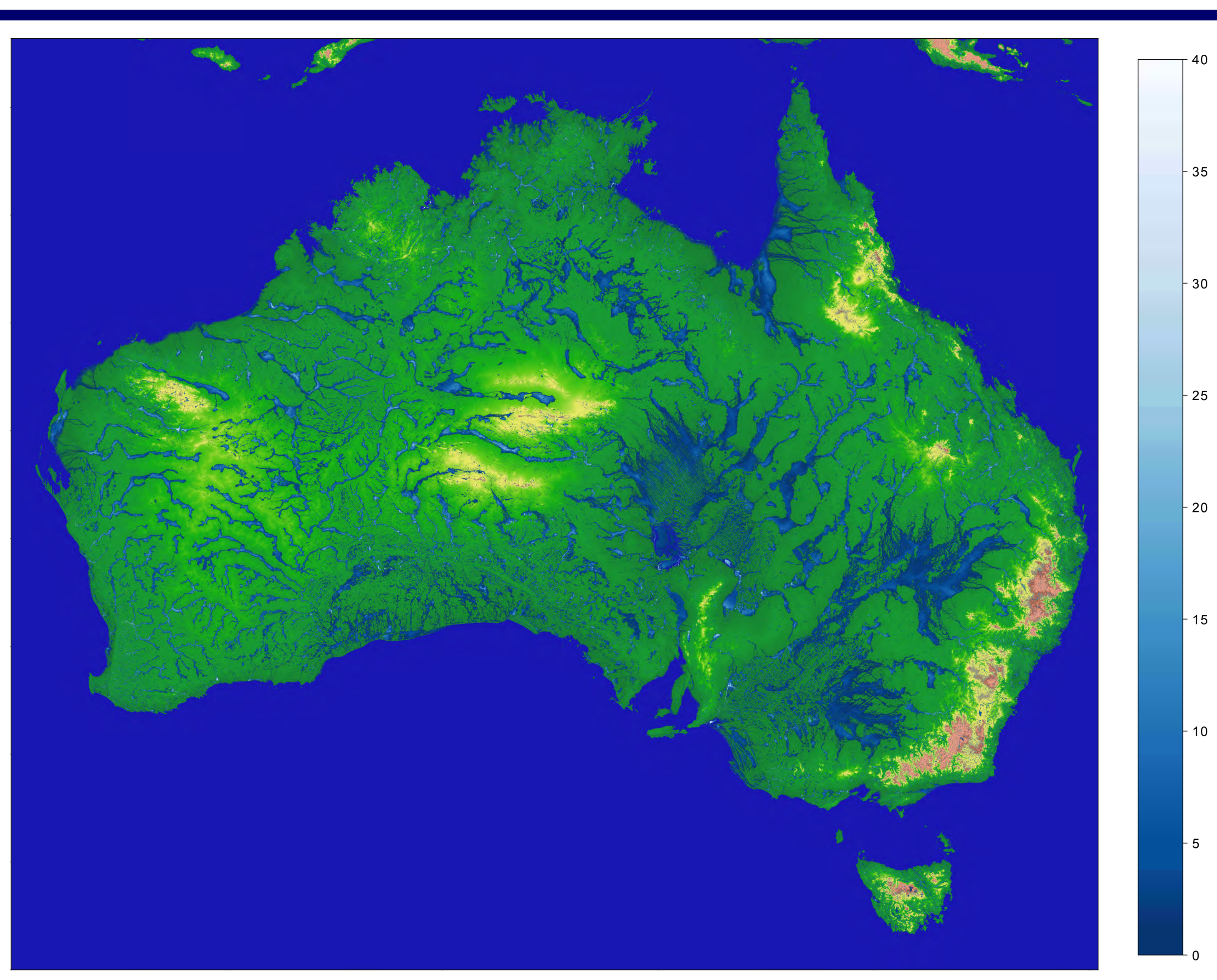

Figure 5: Results of the surface water component of the model for Australia (metres of water). This model can be run on a continental or global scale. While the results of the surface water component alone may be unrealistic - in this case, a combination of an overly large water input and a lack of infiltration and evaporation has resulted in an overestimate of Australian surface water - it does provide an interesting look at Australia's enclosed basins.

Computational Design

Both groundwater and surface water components are intended for continental or global-scale work. In the results shown here (Figures 3, 4, 5), I have performed model runs at a 30 arcsecond resolution. When fully coupled, most model runs will be performed for 500 years (for the transient groundwater component) with monthly timesteps.

For the surface water component, the reduced-complexity approach discussed here was selected after it was found to be more efficient and cause fewer computational issues than a semi-implicit method based on Manning's equation.

However, this approach does come with its own problems. One issue is the increasing slowness with which the algorithm nears equilibrium; improving the result comes at the cost of rapidly extending processing times as flow from one cell to another cascades up the network, and the total volume of water moved in a single iteration becomes ever smaller.

This brings up some issues as to efficiency and deciding on show to threshold the equilibrium of the surface water. A more sensible thresholding scheme still needs to be implemented: at this time, we threshold based on the maximum difference between a before and after-iteration array; the derivative of this difference may be a more useful metric.

Coupling the two model components involves only a few changes to the groundwater module - a small adjustment allows exfiltration to occur, and this can be stored in an array to pass to the surface water component. The surface water component then runs to equilibrium and passes back a groundwater level array adjusted due to any infiltration which has occurred. Efficiency is also a consideration at this point: the surface water model can be run after every single iteration of groundwater movement (i.e. once a month with my settings), but, as the slower component of the model, it can also be run less frequently to allow for longer timescales. For the preliminary result shown in Figure 4, surface water was set to compute once per year in an 80-year model run.