Simulating post-LGM riverine fluxes to the coastal zone: The Waipaoa River System, New Zealand

Phaedra Upton a,*, Albert J. Kettner b, Basil Gomez c, Alan R. Orpin d, Nicola Litchfield a, Michael J. Page a

a GNS Science, Lower Hutt, New Zealand
b CSDMS, INSTAAR, University of Colorado Boulder, CO, USA
c Department of Geography, University of Hawai‘i at Mānoa, HI, USA
d NIWA, Wellington, New Zealand

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A B S T R A C T

HydroTrend, a climate-driven hydrologic transport model, is used to simulate the suspended sediment discharge of the Waipaoa River System (WRS) over the last 5.5 kyr. We constrain the precipitation input with a paleo-rainfall index derived from the high-resolution Lake Tutira storm sediment record. The simulation is extended to 22 ka using a lower resolution version of the model, constrained by terrestrial and marine paleoenvironment indicators and a simulated model of northeast New Zealand’s climate at the Last Glacial Maximum (LGM). Comparison of the 5.5 kyr simulation with the shelf sediment core MD97-2122 suggests that the sediment flux variations observed on the shelf primarily reflect changes in rainfall associated with wetter and drier periods of centuries to millennia duration. Storage of sediment on the Waipaoa River floodplain (Poverty Bay Flats) moderates the signal by reducing the sediment flux reaching the coast. During the LGM conditions were more erosive than the Holocene with tussock and grass dominated vegetation. For erodibility four times the Holocene’s and half today’s, the LGM Waipaoa River System would have generated approximately half the current sediment yield and about 3 times the amount generated when the catchment was fully forested.

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1. Introduction

Steepland rivers draining tectonically active continental margins are known to be particularly sensitive to environmental change, driving variations in the sediment flux to the ocean over time scales of centuries to millennia [e.g., Dadson et al., 2005, 2003; Gomez et al., 2004; Syvitski et al., 2005]. Unlike their longer and larger counterparts that drain passive margins through low gradient and extensive muddy deltas, these short and steep high-yield rivers typically discharge directly to the ocean [e.g., Milliman and Syvitski, 1992]. Thus, the terrestrial sediment sources and marine sinks are closely coupled. In steepland river basins the dispersal of erosion products is also amplified by the severity and flashy behaviour of floods, short river courses, and narrow continental shelves [e.g. Dadson et al., 2005; Gomez et al., 2007a; Sommerfield and Nittroer, 1999].

The Waipaoa River System (WRS) in northeastern New Zealand is a prototypical, non-glaciated steepland fluvial system in which the balance of environmental drivers has been the subject of detailed investigation [Gomez et al., 2007a; Kettner et al., 2007]. It forms the terrestrial component of the Waipaoa Sedimentary System (WSS) (Fig. 1), a focus site of the MARGINS Source-to-Sink initiative [Carter et al., 2010]. As such the WSS is a well characterised and ideal location to quantitatively test the application of sediment models and their ability to hindcast changing catchment yields in response to environmental drivers. Here, we used HydroTrend (Kettner and Syvitski, 2008a), a climate driven hydrological model, along with best estimates of climate variation through time to interpret environmental factors such as vegetation cover and unravel the important drivers of sediment delivery from the Waipaoa River mouth since the LGM.

Using a sediment core from Lake Tutira to reconstruct the trend in regional rainfall [Orpin et al., 2010; Page et al., 1994, 2010] and a reconstruction of shoreline progradation across the Poverty Bay Flats [Wolinsky et al., 2010] we extend the 3000 year model of Kettner et al. (2007) back to the mid Holocene (5.5 cal ka, note that all ages used in this manuscript are calibrated). We also use terrestrial and marine paleoenvironmental indicators in conjunction with a simulated model of northeast New Zealand’s climate at the LGM [Drost et al., 2007] to tune a lower resolution version of the model and extend the simulation back to 22 ka, at the Last Glacial Maximum (LGM).


2. Waipaoa Sedimentary System (WSS)

The WSS (Fig. 1) is located on the tectonically active and geologically complex northern Hikurangi Subduction margin of northeastern New Zealand, where oceanic crust of the Pacific Plate is being subducted obliquely beneath the Raukumara Peninsula along its eastern margin (Lewis and Pettinga, 1993). Subduction-related underplating is elevating the North Island axial ranges (Litchfield et al., 2007; Reyners et al., 1999; Upton et al., 2003; Walcott, 1987), including the catchment of the Waipaoa River, at rates of 1–4 mm yr\(^{-1}\) (Berryman et al., 2003; Walcott, 1987), including the catchment of the upper catchment rank among the highest recorded on Earth (Gomez et al., 2001; Hicks et al., 2000) and sediment yields from 1060–1992 at three gauging stations within the catchment, including the Waipaoa River, at rates of 1–4 mm yr\(^{-1}\), with the highest uplift focused on the headwaters (Berryman et al., 2000; Litchfield and Berryman, 2006). Here, the strongly jointed and clay-rich lithology results in a highly unstable landscape, characterised by earthflows, landslides, and extensive gully systems (DeRose et al., 1991, 1998; Trustrum et al., 1999), which sustains the intense gully systems (DeRose et al., 1991, 1998; Trustrum et al., 1999), which sustains the highly erodible uplands of the Waipaoa River (based on data from 2205 km\(^2\) catchment’s very high suspended sediment yield of 6750 t km\(^{-2}\) yr\(^{-1}\) (Hicks et al., 2000). The modern river channel long profile has an exponential form to within ~25 km of the coast (Gomez et al., 2001), where regional neotectonism causes a transition from uplift to subsidence (Berryman et al., 2000; Brown, 1995). Fluvial sediment storage in the lower, subsiding, floodplain (Poverty Bay Flats, Fig. 1) varies with changes in sea level, but Wolinsky et al. (2010) found that at least during the last 7000 years changes in floodplain storage were counterbalanced by storage in the nearshore areas (Poverty Bay, such that the delivery of mud to the shelf remained roughly constant.

The WRS delivers ~15 Mt yr\(^{-1}\) of suspended sediment and ~0.1 Mt yr\(^{-1}\) of bedload to Poverty Bay (based on data from 1060–1992 at three gauging stations within the catchment, Gomez et al., 2001; Hicks et al., 2000) and sediment yields from the upper catchment rank among the highest recorded on Earth (Walling and Webb, 1996). Bankfull flows of between ~800 m\(^3\) s\(^{-1}\) and 2300 m\(^3\) s\(^{-1}\) have a reoccurrence interval of 4 ± 2 years on the annual maximum series (Gomez et al., 2007b) and can occur in any season, although they are most common in the autumn and early winter months, from April to June (Kettner et al., 2008; Reid and Page, 2002). Hicks et al. (2004) suggest that under modern conditions flood discharges from the Waipaoa River approach or exceed the critical suspended sediment concentration (40,000 mg L\(^{-1}\); Mulder and Svytotski, 1995) for generating hyperpycnal plumes at the river mouth every ~40 years. Such events are of particular relevance when quantifying terrestrial fluxes as they have the potential to episodically deliver considerable quantities of mud. For example during cyclone Bola in March 1988, rainfall locally reached 900 mm over 3 days in the highly erodable uplands of the Waipaoa catchment. Wide-spread landsliding and intense gully erosion caused pronounced aggradation of headwater tributaries. The river level rose by ~10 m and inundated the flood plains of Poverty Bay Flats. At the coast, the Waipaoa River discharged ~36 Mt of suspended sediment over just three days, more than twice its mean annual discharge of 15 Mt yr\(^{-1}\) (Hicks et al., 2004), and anecdotal evidence suggests a fluid-mud layer, ~2 m thick, was deposited on the adjacent continental shelf of Poverty Bay (Foster and Carter, 1997). In contrast to the modern period, except under exceptional circumstances, due to the much lower catchment erodibility there appears to be only limited potential to generate such hyperpycnal plumes when the catchment was fully forested (Gomez et al., 2007a; Kettner et al., 2007), suggesting that riverine sediment yields to the coast were considerably lower and more consistent.

The long-term (millennial) evolution of the WSS is relatively well constrained from geomorphology (e.g., Berryman et al., 2000; Marden et al., 2008) and marine geology (e.g., Gerber et al., 2010; Orpin et al., 2006) studies. One of the key reasons is the presence of multiple tephra layers (e.g., Eden et al., 2001; Gerber et al., 2010) recording eruptions of the Taupo Volcanic Zone to the west (Fig. 1), which sustain the long-term (millennial) evolution of the WSS is relatively well constrained from geomorphology (e.g., Berryman et al., 2000; Marden et al., 2008) and marine geology (e.g., Gerber et al., 2010; Orpin et al., 2006) studies. One of the key reasons is the presence of multiple tephra layers (e.g., Eden et al., 2001; Gerber et al., 2010) recording eruptions of the Taupo Volcanic Zone to the west (Fig. 1), which sustain the modern river channel long profile has an exponential form to within ~25 km of the coast (Gomez et al., 2001), where regional neotectonism causes a transition from uplift to subsidence (Berryman et al., 2000; Brown, 1995). Fluvial sediment storage in the lower, subsiding, floodplain (Poverty Bay Flats, Fig. 1) var
3. The hydrological transport model: HydroTrend

The climate driven hydrological transport model, HydroTrend, simulates water discharge and sediment load at the river mouth on a daily basis (Kettner and Syvitski, 2008a). The model is designed to simulate the fluvial fluxes of rivers for decades to 10,000's of years, provided that the input parameters represent the river basin characteristics over that period (Kettner and Syvitski, 2008a, b, 2009). Input parameters incorporate drainage basin properties (hypsometry, relief, lakes, lithology) together with biophysical parameters (temperature, precipitation, evapotranspiration, and glacier characteristics) and anthropogenic factors (vegetation/human influenced soil erosion factor, dams and reservoirs). Simulated water discharge is slightly simplified for this study as glacier accumulation and melt are not significant. It is therefore based on the water balance module such that the amount of water discharge at the river mouth \( Q \) is similar to the amount of precipitation that a basin receives as rain \( Q_r \) and or snowmelt \( Q_s \), reduced by the evapo-transpiration either from the groundwater pool or from canopy interception \( Q_{\text{evap}} \) and modified by water storage or release of the groundwater \( Q_{\text{gr}} \):\

\[
Q = Q_r + Q_s + Q_{\text{gr}} - Q_{\text{evap}}
\] (1)

Suspended sediment load is computed in two steps. The first step involves computing the average long-term suspended sediment load, \( \bar{Q}_s \) (\( > 30 \) years) which is simulated by implementing the empirical relation \( B\text{QART} \) (Syvitski and Milliman, 2007, Eq. (7a)). The relation is based on a global dataset from 488 rivers that in total deliver 63% of the world’s freshwater to the ocean:

\[
\bar{Q}_s = \omega BQ^{0.31} A^{1.5} R T \quad \text{for} \quad T \geq 2 \, ^\circ \text{C}
\] (2)

where \( \omega \) is a coefficient of proportionality that equals 0.02, \( B \) is the long term water discharge derived as an average from Eq. (1), \( A \) is the drainage basin area of a specific river system, \( R \) is the drainage basin relief and \( T \) is the average long-term temperature of the drainage basin. The \( B \) term expands to:

\[
B = L(1-T_s)E_h
\] (3)

where \( L \) is the average basin lithology factor, which typically ranges between 0.5 and 3, depending on the erodibility of the rock. \( T_s \) represents the trapping efficiency as a fraction (between 0 and 1) due to lakes or reservoirs of the basin, which is defined by the Brune equation for large reservoirs or Brown equation for small reservoirs (see Kettner and Syvitski, 2008a). \( E_h \) is a basin disturbance factor, originally designed to describe the anthropogenic impact in a global context based on Gross National Product (GNP) per capita and population density (Syvitski and Milliman, 2007). However, Kettner et al. (2007, 2010) successfully applied a modified disturbance factor where \( E_h \) is based on changes in vegetation cover over time (a vegetation-erosion index), which will be applied to this study. \( E_h \) generally varies between 0.2 and 2 but for a few rivers can increase to 8 (Kettner et al., 2007; Syvitski and Milliman, 2007).

The second step involves computing short-term suspended sediment load, which can be simulated once the long-term sediment load trend is determined. Morehead et al. (2003) developed a stochastic relationship (i.e. between sediment load and discharge:

\[
\frac{Q_{s,H}}{Q_H} = \frac{Q_{s,0}}{Q_{0}} C_{10}
\] (4)

where \( Q_{s,0} \) is the daily sediment load, \( Q_{s,H} \) is a daily log-normal random distribution to mimic short-term scatter in the sediment load (inter-annual variability), and \( C_{10} \) is a normal distribution annual rating coefficient (intra-annual variability). Daily variation is typically of less interest for studies like ours, which focuses on changes over centuries to millennia. Short-term processes including: (1) a reduction in the rate of increase in decline in sediment concentration with discharge at higher discharges (Hicks et al., 2000, 2004), mimicked by imposing a upper boundary condition (Kettner et al., 2007), and (2) floodplain storage (currently accounting for 16% of the suspended sediment load transported during events that exceed bankfull conditions), incorporated by making a proportionate reduction of suspended sediment when the simulated water discharge exceeds 1550 m³ s⁻¹, do influence the simulated sediment budget and are therefore incorporated (Kettner et al., 2007).

Model source code is freely available from the Community Surface Dynamics Modeling System (CSDMS) repository (http://csdms.colorado.edu/wiki/Model:HydroTrend). A more detailed description of the model structure and the governing equations is provided by Kettner and Syvitski (2008a).

4. Input and boundary conditions

4.1. Sealevel and changing catchment area, LGM to present

The evolution of the WRS since the LGM has been reconstructed by estimating river mouth positions on a digital elevation model, which is a combination of the present-day (onshore) 20 m digital elevation and the (offshore) post-glacial marine transgression surface (W1) of Orpin et al. (2006) (Fig. 2). In doing so, we have ignored river incision as well as tectonic uplift and subsidence of the lowstand coastal plain. The timing of the post-LGM phase of incision through the W1 (glacial) gravels and into bedrock is not well constrained and it should be noted that because the simulation ignores the effect incision had on sediment fluxes the modelled sediment fluxes are minimums.

The location of the river mouth is unknown prior to ~1.7 ka, and from 1.7 ka to 9.5 ka it is assumed to lie along the present day river. Prior to 10 ka the paleo-river mouth is assumed to be directly offshore of the current river mouth (Fig. 2) at elevations taken from localised on the New Zealand sea level curve of Carter et al. (1986), calibrated by Lamarche et al. (2006). After 10 ka the river mouth was positioned on the reconstructed shorelines beneath the Poverty Bay Flats by Pullar and Penhale (1970) and Brown (1995), calibrated by Wolinsky et al. (2010). Once the river mouth positions were estimated, the catchment boundaries were inferred to incorporate various rivers (e.g., the Waimata) and coastal streams (Fig. 2). Based on these reconstructions, the LGM river mouth was close to the shelf break, and the river course was a minimum of 32 km longer than it is at present (134 cf 102 km) and the WRS drained approximately twice the area it does today (4019 m² of 2140 m² (Fig. 2 and 3B). From the LGM to approximately 7.3 ka, the river length and catchment area decreased, possibly in a step-wise fashion in response to both stillstands in the post-glacial sea level rise and as tributaries disconnected to drain directly into the ocean. At approximately 7.3 ka, when sea level reached its maximum value (Clement et al., 2010; Gibb, 1986) the river length and catchment area attained minima of 82 km and 1730 km² respectively (Fig. 3B). Subsequently, as a result of sea level stabilisation and continuing rapid
progradation, the river length and catchment area increased, capturing rivers such as the Te Arai, to the present day (Brown, 1995; Wolinsky et al., 2010).

4.2. Climate and storms

Precipitation patterns in the modern Waipaoa catchment are spatially diverse, ranging from ~1000 mm in the coastal zone to ~3000 mm in the upland headwaters, with an average of ~1590 mm across the catchment as a whole. The catchment is highly sensitive to intense rainfall caused by deep mid-latitude depressions and extra-tropical cyclones, which can produce as much as 800 mm of precipitation in a 72 h period (Hastings, 1990; Reid and Page, 2002). In the historic record the ~1:100 yr March, 1988 Cyclone Bola storm caused wide-spread landsliding and intensified gully erosion in the catchment and generated ~36 million tonnes of suspended sediment in a 3 day period (Hicks et al., 2004). A hind-cast relationship based on the storm sediment record from Lake Tutira suggests that there may have been ~53 pre-historic storms with a magnitude similar to the Cyclone Bola event in the past ~7 ka, and perhaps 7 even larger catastrophic storm events (Orpin et al., 2010). Although under modern pastoral land use events of low and intermediate magnitude dominate suspended sediment yield (Hicks et al., 2000, 2004), these large-magnitude events dominated in the period prior to human arrival (~0.5 ka), when the indigenous vegetation substantially decreased landscape sensitivity (Gomez et al., 2007a; Trustrum et al., 1999).

4.3. Climate and climate proxies to LGM

Running HydroTrend requires climate parameters including annual and seasonal temperature and precipitation as well as their statistical variability over the time period of the model runs. These are not available over the time period of interest and so we must approximate them. For the past 22 ka we use two pinning points for the Waipaoa, today and an LGM climate model (Drost et al., 2007). Seasonal climate statistics at the LGM are estimated using the climate model of Drost et al. (2007) for the eastern North Island. Their model uses the United Kingdom Met Office global model (HadAM3H), a regional model (HadRM3H) and glacial conditions as specified by the Paleoclimate Modelling Intercomparison Project (PMIP). Climate models suggest the region was colder and drier than present (Drost et al., 2007), consistent with the presence of loess and recycled pollen in offshore core P69 (Fig. 1, McGlone, 2001). Between these points we looked for a proxy that could provide us with a continuous climate related record from 22 ka to the present day. This method is similar to that used by Kettner and Syvitski (2008b) when modelling the Po River catchment, they forced their climate statistics using a normalised $\delta^{18}O$ GRIP curve (Dansgaard et al., 1993). The only record in the vicinity of the WRS that provides quantitative paleotemperature estimates over the past 22 ka is Pahnke and Sach’s (2006) sea surface temperature (SST) record derived from the giant piston core, MD97-2121 (Fig. 1, 2314 m water depth, Carter et al., 2008). The period 3–0 ka is missing from this record and we approximate the SST over this time period by a linear increase to reach present day values
We extrapolated to 500 year timesteps and calculated the change in SST over each timestep. Although present day, short-term (daily or monthly) sea-surface temperatures vary less than land temperatures, the long term sea surface temperature record is consistent with other fragmentary temperature records (Litchfield et al., 2011).

In the absence of any long term paleo-precipitation record, we use the SST as a proxy for both paleo-temperature and paleo-rainfall (e.g., Kettner and Syvitski, 2008b). The modelled seasonal and annual climate statistics at LGM (Drost et al., 2007) were combined with the present day climate statistics and interpolated over time by pinning the ends and using the SST as a forcing factor. The values of annual and seasonal paleo-precipitation were derived assuming that precipitation has varied uniformly with paleo-temperature between the LGM to now (Fig. 3D and E).

4.4. The Lake Tutira proxy rainfall record

A mid-to-late Holocene storm sediment record from Lake Tutira provides a continuous, high resolution record of erosional events in the surrounding 32 km² watershed (Page et al., 2010). The lake lies 110 km to the southwest of the Waipaoa River basin, and variations in the southern Pacific atmospheric circulation that are linked to El Niño-Southern Oscillation (ENSO) and the Southern Annular Mode (SAM) have been shown to influence the frequency of storms throughout the middle and late Holocene (Gomez et al., 2011). New Zealand rainfall varies regionally due to the interaction of the atmospheric circulation and topography (Kidson and Renwick, 2002; Kidson et al., 2009; Ummenhofer and England, 2007), and a high-resolution sediment record from the WSS also contains a strong ENSO signal (Gomez et al., 2004). We show elsewhere that there is a coherent response in the sediment records from WSS and Lake Tutira which we use herein as a regional proxy for Eastern North Island rainfall.

Quantitative New Zealand climate proxies are of limited length but the storm sediment record from Lake Tutira does provide a record of precipitation for the past 7 kyr (Page et al., 2010). Mean annual rainfall in the Tutira Catchment is 1358 mm, and (for the > 100 yr period of record) minimum and maximum annual rainfall totals are 0.5 and ~2.0 times the mean. Overall, New Zealand rainfall is a product of interactions between the country’s north-south trending mountain ranges and the zonal atmospheric circulation, which is modulated by the SAM and ENSO teleconnections (Kidson and Renwick, 2002; Kidson et al., 2009; Ummenhofer and England, 2007). During the past 7 kyr, the changing relationship between these two leading southern hemisphere climate modes has given rise to periods of a century or more when storm activity recorded in Lake Tutira was either enhanced or subdued, compared to the modern level of ~1 sediment producing storm per decade (Gomez et al., 2011; Page et al., 2010). Variations in storm activity seen in the Lake Tutira record have been equated with wetter and, by analogy, drier periods in the Waipaoa River basin and across the central North Island (Gomez et al., 2011).

Here we used the Lake Tutira storm sediment to create a rainfall record for the last ~5.5 kyr (Fig. 4A), by scaling the number of storms in each 100 yr period to the modern rainfall record (Litchfield et al., 2011). To do this we assumed that the number of storms recorded in the historic period (11 per 100 yr) reflects the contemporary mean annual rainfall and that those periods with the maximum (87 per 100 yr) and minimum (0 per

![Fig. 4. (A) Scaled rainfall derived from Lake Tutira (see text for details); (B) Modelled suspended sediment discharge from the Waipaoa over the past 5.5 ka based on scaled rainfall in (A). Figures on the right hand side represent average Qs for the same time periods as sediment accumulation rates from Phillips and Gomez (2007). Numbers in parentheses are from a simulation without floodplain storage. (C) Observed rate of terrigenous mass accumulation on the middle shelf at core site MD97-2122 (Gomez et al., 2007a; Phillips and Gomez, 2007), figures in italics are sediment accumulation rates (mm yr⁻¹) over the indicated time periods. (D) Modelled suspended sediment discharge from the Waipaoa over the past 5.5 ka using climate derived from the SST record of MD97-2121. Average Qs, are shown for the same periods as in (B) and (C).](#)
100 yr) number of storms equate with 2.0 and 0.5 times the contemporary mean annual rainfall, respectively. Rainfall in all other (100 yr) periods was estimated by linear interpolation between these values. These values were then scaled to the Waipaoa using the present day average rainfall of 1590 mm yr$^{-1}$. Having determined the century mean rainfall estimates they were input into HydroTrend which then uses a statistical model to generate standard deviations and hence daily precipitation values from observed statistical variability. Present day climate variability was used for the whole model time as we had no way to estimate climatic variability in the past (Kettner and Syvitski, 2008a; Morehead et al., 2003).

4.5. Vegetation and climate proxies

Within the BQARET equation, $E_h$, a modified disturbance factor, is used to take into account changes in vegetation cover over time (Kettner et al., 2007, 2010; Syvitski and Milliman, 2007). In order to approximate the influence of vegetation on erosion in the WRS through the LGM and as reforestation occurred with warming, we use the pollen record as a proxy for the degree of forest cover which we then convert to an $E_h$ value (Fig. 3F). Today, under conditions of almost full deforestation, short introduced grasses and sheep farming, $E_h$ has been determined to be $\sim 8$ in the WRS (Kettner et al., 2007; Syvitski and Milliman, 2007). The pollen record shows maximum deforestation at 20 ka with only small, scattered forest refugia along the coast (McGlone, 2001; Newnham and Lowe, 2000). The majority of the catchment was covered by shrubs and tussock. The catchment was once more fully reforested by $\sim 17.5$ ka (McGlone, 2001), consisting of beech forest in the montane areas and emergent podocarps over a continuous canopy in lowland areas. Periodic disturbances occurred due to volcanic eruptions, fires and severe storms, but the largest disturbance was the forest clearance starting with Maori arrival approximately 700 years ago (Wilmshurst et al., 1999), and increasing rapidly with European arrival starting in 1830 AD (MacKay, 1982). The conversion to grass pasture was almost complete by 1920 AD.

It is difficult to estimate the effect of vegetation on erosion in the LGM as we have no similar catchment with tussocks and shrublands to use as an analogue. Over several glacial cycles, non-carbonate mass accumulation rates typically increase during cold periods and reduce in warm periods (Carter et al., 2000). The only constraint we have for our period of interest is that Carter et al. (2002) estimate LGM sediment flux to a slope basin recorded in core MD97-2121 was approximately twice that of the Holocene, within the limitations of dating based on the tephra record. Using this constraint, we assume that the maximum value of $E_h$ during the LGM, corresponding to maximum deforestation at 20 ka, was between 3 and 5, approximately 4 times as erosive as during times of full forest cover when $E_h = 1$ (Kettner et al., 2007) and half as erosive as today. We ran a series of simulations with varying $E_h$ between 3 and 5 to explore its effect on sediment flux from the LGM catchment.

4.6. Floodplain storage on the Poverty Bay

Since $\sim 7$ ka the shoreline at Poverty Bay has migrated 12 km seaward, building the Poverty Bay Flats (Wolinsky et al., 2010). Kettner et al. (2007) included a proportionate reduction in the suspended sediment load at the river mouth whenever the simulated discharge of the Waipaoa River exceeded 1550 m$^3$ s$^{-1}$ based on measurements of modern day floodplain storage during flood events (Gomez et al., 1999). That sediment was lost to the storage for the entire duration of the model run. We assumed this to be the case for $t = 0–3$ ka and prior to $t = 3$ ka we reduced the amount of floodplain storage proportionally assuming it was zero at 7 ka.

5. Model results

5.1. Simulated riverine suspended sediment flux over the past 5.5 ka

Our simulation was run over the past 5.5 ka (Fig. 4B) for a comparison of the HydroTrend results to the reconstruction of the continental shelf core MD97-2122 (Fig. 4C) (Gomez et al., 2004, 2007a; Phillips and Gomez, 2007). Precipitation in this simulation was driven by the Lake Tutira storm record. Suspended sediment yield tracks precipitation and we see spikes in simulated sediment flux during wetter periods from $\sim 4.8$ ka to 4.4 ka and 2 ka to 1.8 ka. Average annual sediment fluxes during these periods are $2.7 \pm 0.5$ (mean $\pm 2\sigma$) Mt yr$^{-1}$ and $2.8 \pm 0.5$ Mt yr$^{-1}$, respectively. Drier periods produce sediment fluxes between 1.8 Mt yr$^{-1}$ and 2.2 Mt yr$^{-1}$.

The long-term variations in sediment flux predicted by the HydroTrend model agree with those observed in MD97-2122 (Gomez et al. and C) (Gomez et al., 2004, 2007a; Phillips and Gomez, 2007). The increase in terrigenous flux and sedimentation rate centred on 2 ka is particularly obvious in the model and it implies that this sediment peak within the record is a result of the regionally wetter period previously documented by Lorrey et al. (2008) and Page et al. (2010). The slight decrease in terrigenous flux at $\sim 3.4$ ka (from 1.4 g cm$^{-2}$ yr$^{-1}$ to 1.3 g cm$^{-2}$ yr$^{-1}$) is manifest only when floodplain storage is included in our simulation. A second simulation, driven by the SST paleoclimate model, predicts a more uniform rate of sediment delivery with average $Q_{ss} = 2.3–2.4$ Mt yr$^{-1}$ over the whole 5.5 ka time period (Fig. 4D).

Whereas neither model produces an exact match, we can compare trends and note differences between the two models. The emphasis of the two simulations is different. Precipitation is the major driving of discharge and suspended sediment load, thus wetter and drier periods will stand out in the simulation driven by the LT storm record. The simulation driven by SST will reflect general climatic variability.

5.2. Simulated riverine suspended sediment flux from LGM for a constant vegetation conditions

A first order model of suspended sediment flux ($Q_{ss}$) from the WRS back to the LGM uses the SST derived from MD97-2122 as a climate driver. Here, the model uses constant $E_h = 1$, as if the catchment had been fully forested throughout its post-LGM history (Fig. 5A). The simulation indicates a slight decrease in sediment flux from 22 ka to present day and reflects two competing influences. As subsequent catchment size halves from $\sim 4000$ km$^2$ to $2000$ km$^2$ with sea-level rise and shoreline transgression, sediment supply reduces and the riverine capacity to transport sediment reduces sympathetically. Meanwhile, since the LGM the watershed climate changed from cold and dry conditions to warmer and wetter conditions, increasing the production of sediment through enhanced erodibility. Two climatic events are captured within this time range, the Antarctic Cold Reversal (ACR, $\sim 14.5–12.5$ ka) (Wilmshurst et al., 2007) and the Holocene Climate Optimum (HCO, $\sim 9.0–7.0$ ka) (Wilmshurst et al., 2007). Following the ACR, as the climate warmed into the HCO, the predicted sediment load increases. At the HCO simulated $Q_{ss}$ drops significantly despite increased temperature and precipitation. This reflects the reduction in the catchment size with maximum marine inundation of the coastal plain. Following the HCO, sea level stabilises and modelled $Q_{ss}$ increases slightly to the present day as the temperature rose slowly. In order to disentangle the variability in $Q_{ss}$ resulting from climate alone, we ran a second simulation with the catchment size fixed at that of today (Fig. 5A, grey curve). By the way of comparison, the colder, drier conditions at the LGM produce about two thirds of the today’s sediment yield. The most productive time is from 12 ka to 5 ka, following the ACR, as the climate warmed into and through the HCO.
Simulated by HydroTrend using a $E_h$ variable vegetation conditions.

6. Discussion and interpretations

Previous studies (Gomez et al., 2007a; Kettner et al., 2007, 2009) have indicated that in the last 3000 years changes to the vegetation cover (particularly those caused by human activities and volcanic eruptions) and the associated changes in the erodibility of the catchment have exerted the dominant control on sediment flux. The present simulation, which extends the record back to 5.5 ka, suggests that lower frequency variations – the result of regional changes in rainfall patterns that give rise to extended wetter and drier periods of centuries or millennia duration – may also exert an important influence on suspended sediment flux. Using the Lake Tutira storm sediment record (calibrated to account for human impacts upon the landscape over the past 500 years) as a proxy for the regional rainfall, we are able to replicate the long-term variations in the terrigenous flux and sedimentation rate recorded on the middle Poverty shelf, in core MD97-2122. This suggests that, in the absence of profound disturbances to the vegetation cover, climate exerts the dominant control on sediment fluxes. The close comparison between our climate-driven predicted sediment fluxes and the record on the shelf also helps to place a constraint on the timing of post-glacial incision within the Waipaoa catchment (Marden et al., 2008). The implication being that the impact of the rapid post-glacial incision must have attenuated and the signal passed through the catchment by 5.5 ka.

6.1. Erosion conditions at the LGM

We attempted to estimate erosion conditions within the Waipaoa catchment at the LGM. Lack of significant forest cover in the catchment at LGM is expected to have led to increased sediment loads (Orpin et al., 2010; Page and Trustrum, 1997; Page et al., 2010). Unlike today, the dominant ground cover was tussocks and native grasses with some shrublands. While not as protective of the landscape as full forest cover, the LGM vegetation likely afforded greater protection from erosion than today’s short pastoral grasses and the erosive impact of grazing stock. An $E_h$ of ~4 at LGM doubles the sediment flux compared to the pre-Anthropogenic Holocene and is taken as a reasonable approximation to conditions at LGM. Higher erodibility at LGM is not restricted to the Waipaoa catchment. Dosseto et al. (2010) use paleochannels of the Murrumbidgee River in Australia to show that erosion rates increased during glacial periods due to a change in vegetation. They conclude that there is an indirect influence of climate on erosion in a catchment via the influence of climate on the type of ecosystem present. In their example, shrubs rather than trees with deep root systems were coincident with higher erosion rates (Dosseto et al., 2010).

6.2. Future modelling of the WSS from a Source-to-Sink perspective

The terrestrial and marine components of the WSS portrayed as a Source-to-Sink cartoon in Fig. 6. Our HydroTrend simulations address the terrestrial delivery of sediment to the coastal and shallow marine transfer/storage zone. With increasing depth, further seaward transfer and storage occurs on the continental shelf and slope basins and beyond (Carter et al., 2010). Through each seaward step away from the terrestrial source, terrestrial signals are overprinted, modified or removed. Using our model, we show that since 5.5 ka precipitation may exert some control on the vertical sedimentation rate at core site MD97-2122 on the shelf. However, providing appropriate physical proxies (e.g., precipitation, ages) to constrain simulations further back in time to also include the last-glacial transgression presents major challenges despite the WSS being among the best characterised source-to-sink study areas in the world. We are limited by the resolution of the age model in cores such as MD97-2122 where we have four dates over 5.5 ka. To be able to more directly compare our model results with the depositional record of the cores, an age model with tied points every 2–300 years would be required. Furthermore, over these longer time scales the impact of marine overprinting...
through bioturbation, mixing and erosion of the slope record, tectonic deformation, and changes in ocean circulation become important factors. An important contribution of our modelling effort has been to unravel strong source drivers, but an outstanding challenge remains to model and balance sink deposition and responses over glacio-eustatic timescales. Currently, HydroTrend is critically limited by the reliable high resolution temperature and precipitation records as well as the challenge of modelling complex transitions between various system components. Coupling of the predicted water and sediment fluxes to a sediment model such as Sedflux (e.g., Kettner et al., 2009) that takes into account subsidence and uplift within the Poverty Bay floodplain as well as tectonics offshore could address some of these issues.

7. Conclusions

Previous studies have indicated that in the last 3000 years vegetation change and associated erodibility are the dominate control on sediment flux from the Waipaoa River (Gomez et al., 2007a; Kettner et al., 2007, 2009). This study suggests that the sediment flux alterations over the past 5.5 ka have higher frequency variations driven by precipitation. The simulation is consistent with observations of the continental shelf core MD97-2122 including a slight reduction in sediment flux at ~3.4 ka associated with increased storage on the Poverty Bay floodplain. The sediment flux from the WRS is dominated by climate with increasing storage superimposed upon it.

Further back to 22 ka the source model is more difficult to constrain as are the links to the seaward transport and storage components of the Source-to-Sink system. An LGM sediment flux of approximately twice the Holocene flux is predicted for a catchment disturbance factor, of 4 during glacial times. Following re-vegetation of the WRS at about 17.5 ka, increasing rainfall and decreasing catchment size compete from LGM to the Holocene to constrain as are the links to the seaward transport and storage components of the Source-to-Sink system. An LGM sediment flux of approximately twice the Holocene flux is predicted for a catchment disturbance factor, of 4 during glacial times. Following re-vegetation of the WRS at about 17.5 ka, increasing rainfall and decreasing catchment size compete from LGM to the Holocene to maintain the simulated water and sediment discharge approximately constant. It remains a challenge to model system behaviour back in time and to predict sink responses over glacio-eustatic timescales.

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