Controls on sediment export from the Waipaoa River basin, New Zealand

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ABSTRACT

A stability model of drainage basin mass balance is used to interpret historic and prehistoric patterns of sediment production, storage and output from the Waipaoa River basin, New Zealand and assess the sensitivity of basin sediment yield to land use change in the historic period. Climate and vegetation cover changed during the late Holocene, but the drainage basin mass balance system was stable before the basin was deforested by European colonists in the late 19th and early 20th centuries. In this meso-scale dispersal system sediment sources and sinks are closely linked, and before that time there was also little variability in the rate of terrigenous mass accumulation on the adjacent continental shelf. However, despite strong first-order geologic controls on erosion and extensive alluvial storage, sediment delivery to the continental shelf is sensitive and highly responsive to historic hillslope destabilization driven by land use change. Alluvial buffering can mask the effects of variations in sediment production within a basin on sediment yield at the outlet, but this is most likely to occur in basins where alluvial storage is large relative to yield and where the residence time of alluvial sediment is long relative to the time scale of environmental change. At present, neither situation applies to the Waipaoa River basin. Thus, the strength of the contemporary depositional signal may not only be due to the intensity of the erosion processes involved, but also to the fact that land use change in the historic period destabilized the drainage basin mass balance system.

INTRODUCTION

The sediment flux from rivers to Earth’s oceans is influenced by a broad suite of factors which determine the supply from uplands to the fluvial system, and transport and storage within channels. Thus, it is not unreasonable to assume that any change in the dynamics of sediment production and dispersal within a drainage basin will trigger a corresponding change in output. Accordingly, changes in climate and land use are believed to have greatly modified sediment fluxes in geologically recent and historic times (e.g. Milliman & Meade, 1983; Hay, 1994; Milliman, 2001; Syvitski & Milliman, 2007). Nevertheless, sediment fluxes often appear to be relatively constant over large areas and lengthy periods of time. This may be because the landscape is in a state of dynamic equilibrium, sensu Hack (1960), whereby the rate of sediment production is approximately equal to sediment yield, in which case the denudation rate derived from present-day mass fluxes should not differ greatly from geological estimates of the long-term rate (cf. Renneau & Dietrich, 1991; Granger et al., 1996; Meigs et al., 1999). Constancy also may be maintained because the influence of some dominant factor (e.g. relief or tectonics) is so pronounced relative to that of other factors (e.g. climate and vegetation change) that the latter are not readily detectable, or because storage mutes the impact of climate or land use change on sediment yield (Summerfield & Hulton, 1994; Gunnell, 1998; Métivier & Gaudemar, 1999; Phillips, 2003). In the latter case, the effects of changes in erosion rates may primarily be reflected by changes in the volume of storage within the drainage basin, rather than by changes in sediment yield at the basin outlet (Trimble, 1977; Church & Slaymaker, 1989). The net result is that there is substantial variation in the extent to which sediment yield is responsive to changing conditions within the drainage basin. Isolating the factors that determine this responsiveness is important for identifying the portions or aspects of geomorphic systems which respond to environmental change, and for evaluating its effect on the land–ocean transfer of terrigenous sediment and associated bio-limiting nutrients (Syvitski & Morehead, 1999; Sommerfield et al., 2002; Gomez et al., 2003; Leithold et al., 2005).

Phillips (2003) presented and tested a stability model of drainage basin mass balance, which specifies the geomorphic conditions for dynamical stability, as reflected in relative constancy of sediment yield. Here, we use this model to assess the sensitivity of sediment yield to land use change that occurred within the Waipaoa River basin, New Zealand, in the historic period. Specifically, we use the mass balance stability framework to interpret historic...
and prehistoric patterns of sediment production, storage and output. In so doing we acknowledge that considerable spatial and temporal variability exists in fluvial sediment transport and storage, and we are concerned only with the general relationship between the recent (historic) and longer-term (late Holocene) output of this meso-scale fluvial system. The Waipaoa River basin is an ideal setting to explore this issue because it is characterized by active tectonics, steep hillslopes, and extensive alluvial storage, and changes in climate, vegetation cover and land use have been documented during the time period in question (Gomez et al., 2004b, in press).

STUDY AREA

The unglaciated, 2205 km² Waipaoa River basin lies on the East coast of New Zealand's North Island (Fig. 1), within the zone of active deformation associated with the obliquely converging Australian and Pacific lithospheric plates, and surface uplift in the headwaters is ~4 mm year⁻¹ (Berryman et al., 2000; Mazengarb & Speden, 2000). Rocks in the drainage basin comprise two major units: (i) a structurally complex suite of Cretaceous and lower Tertiary sandstone, argillite, mudstone, marl and limestone, that in the north-northwestern part of the headwaters forms thrust sheets of the East Coast Allochthon; and (ii) a cover sequence of poorly consolidated Miocene and Pliocene sandstone, siltstone, mudstone and limestone. Airfall tephras derived from the Taupo and Okataina volcanic centres are dispersed throughout the basin (Vucetich & Pullar, 1964), and unconsolidated Pleistocene, Holocene and recent alluvium is preserved as terrace remnants throughout the headwaters and in the lower reaches of the major tributaries, as well as on the Poverty Bay Flats (Pullar & Penhale, 1970; Brown, 1995; Berryman et al., 2000; Phillips et al., 2007). The last major phase of downcutting ceased ca. 5.58 ka calendar years BP, and much of the basin has not yet adjusted to the new base level and retains its relict, pre-incision form (Berryman et al., 2000; Crosby & Whipple, 2006). The presence of air-fall tephras within soil profiles throughout the headwaters also suggests portions of the weathered mantle have remained stable for thousands of years (Gage & Black, 1979). Soil development throughout the headwaters is strongly related to rock type, tephra cover and rainfall, and the soil profiles...
on many hillslopes adjacent to river channels have been truncated by erosion and contain no topsoil (O’Byrne, 1967; Jessen et al., 1999).

During the middle Holocene winters in the eastern part of the North Island became wetter, and the prevailing El Niño Southern Oscillation (ENSO) climatic regime was established by ca. 4000 calendar years BP (Froggatt & Rogers, 1990; McGlone et al., 1993; Gomez et al., 2004b). Mean annual precipitation currently ranges from ~1000 mm year$^{-1}$ at the coast to ~3000 mm year$^{-1}$ in the headwaters, where the maximum elevation is ~1200 m, and cyclonic storms are an important component of the local rainfall regime (Hessell, 1980; Hastings, 1990). The primary forests in the headwaters and on the Poverty Bay Flats were established in the mid-Holocene (Norton et al., 1986; Froggatt & Rogers, 1990; McGlone et al., 1993). Before the arrival of Polynesian settlers in the 13th century AD, the forests periodically were disturbed by volcanic eruptions, fire and severe storms (Grant, 1985; Wilmshurst et al., 1999). Cultural activities, involving land clearance by fire, impacted the native vegetation from ~650 $^{14}$C year BP onwards (Wilmshurst et al., 1999), but wholesale destruction of the forest cover did not begin until after the arrival of European colonists in the 1820s (Pullar, 1962). Today only ~2.5% of the basin is covered by old growth forest.

Different combinations of lithology, structure and topography in the Waipaoa River basin give rise to ‘terrain types’ with varying degrees of susceptibility to erosion and slope failure (Fig. 1). Hills surrounding the Poverty Bay Flats and in the headwaters underlain by the Miocene–Pliocene cover sequence are prone to shallow landsliding, whereas the deformed rocks in the headwaters are susceptible to gully erosion (Reid & Page, 2002; Marden et al., 2005). At present, shallow landsliding is an important erosion process in 67% of the basin. Before deforestation less of the basin would have been susceptible to landsliding, but much larger, deep-seated failures probably occurred under the native forest cover and on steep, riparian hillslopes in the headwaters that only possessed a skeletal soil cover (Henderson & Ongley, 1920). After deforestation, the tempo of erosion increased dramatically (Trustrum et al., 1999), and the Waipaoa River and its tributaries have been aggravating in response to the increased sediment supply since the early decades of the 20th century (Gage & Black, 1979; Gomez et al., 1999, 2001).

Both the very high contemporary rate of erosion (6750 t km$^{-2}$ year$^{-1}$) and the Waipaoa River’s correspondingly high suspended sediment yield (15 ± 6.7 Mt) are driven by gully erosion (Hicks et al., 2000). Gully erosion can occur under the native forest cover (cf. Liebault et al., 2005), but most of the gullies were initiated during the first quarter of the 20th century after deforestation changed the soil moisture status and pattern of hillslope runoff, and lowered the topographic threshold for erosion (Marden et al., 2005; Parker et al., 2006). Recent (post-1960) exotic reforestation of the most severely eroded land in critical headwater regions has helped stabilize many smaller gullies, but ~420 gullies in the headwaters remain active and erosion in several large gully complexes continues largely unabated (DeRose et al., 1998; Marden et al., 2005).

### THE DRAINAGE BASIN DENUDATION SYSTEM

The drainage basin mass balance model is based on a conception of geomorphic systems as $n$-dimensional systems with components $x_i, i = 1, 2, \ldots, n$, such that

$$\frac{dx_i}{dt} = f(dx_i/dt)$$  \hspace{1cm} (1)

where $x$ indicates the vector of all $x_i$. Thus the components of the system potentially affect and are potentially affected by one another. The system state at time $t$ is given by

$$x(t) = Cx(o)e^{\lambda t}$$  \hspace{1cm} (2)

where $x(o)$ is the initial state (at the onset of landscape evolution or at the time of a change or disturbance); and $C$ is a vector constant related to the initial conditions. The $\lambda$ values are the $n$ Lyapunov exponents of the system (equivalent to the real parts of the complex eigenvalues of a Jacobian interaction matrix of the system), where $\lambda_1 > \lambda_2 > \ldots > \lambda_n$. If all values of $\lambda < 0$, the effects of minor initial variations or of small perturbations are damped over time, and the system is stable. One implication of stability in a fluvial sediment dispersal system is that, in spite of environmental change, sediment export will vary little over time. Any $\lambda > 0$ indicates exponential divergence over finite time, and dynamical instability.

Sediment yield is expected to vary over time in unstable fluvial systems subject to environmental change. In this case the interacting components are the parts of a mass balance system, wherein those parts may influence each other as depicted in Eqn. (1). A mass balance for any drainage basin is

$$W = Y + A + R$$  \hspace{1cm} (3)

where $W$ (weathering) is broadly defined to include all processes that convert bedrock to transportable debris; $Y$ (yield) represents all mass transported out of the basin; $A$ (alluvium) is the mass stored in the channel or on the floodplain; and $R$ (regolith) is the mass of weathered material stored as regolith on hillslopes (Phillips, 2003). All quantities in Eqn. (3) can be defined in terms of units of mass or volume per unit time. It is assumed all weathering products either remain within the drainage basin ($A, R$), or are exported by river flow ($Y$). Magnitudes and rates of $W, Y, A$ and $R$ are temporarily variable and difficult to measure, but the effects of an increase or decrease in any component on the other components qualitatively can be identified and are portrayed graphically in Fig. 2. These relationships can be analysed to determine the dynamical stability of the system using qualitative asymptotic stability analysis, as described by Phillips (1999).
Phillips (2003) summarized the interactions between weathering rates and the allocation of weathering products among regolith, alluvial storage and sediment yield, and determined the qualitative stability of the drainage basin mass balance model by employing the Routh–Hurwitz criteria (Cesari, 1971). These criteria make it possible to determine whether or not any $\lambda > 0$. The present analysis was accomplished by translating the relationships shown in Fig. 2 into an interaction matrix, the characteristic equation of which can be written in terms of feedback (Puccia & Levins, 1985). Feedback at level $k$ ($F_k$) represents the mutual influences of system components on each other for all disjunct loops with $k$ components.

$$F_k = \sum (-1)^{m+1}Z(m, k)$$  \hspace{1cm} (4)

$Z(m, k)$ is the product of $m$ disjunct loops with $k$ components. $F_0 = -1$ by convention. The characteristic equation is

$$F_0\lambda^m + F_1\lambda^{m-1} + F_2\lambda^{m-2} + \cdots + F_{m-1}\lambda + F_0 = 0$$  \hspace{1cm} (5)

The Routh–Hurwitz criteria give the necessary and sufficient conditions for all real parts of all eigenvalues to be negative (and thus for all Lyapunov exponents to be negative). These are:

- $F_i < 0$, for all $i$.
- Successive Hurwitz determinants are positive.

The interaction matrix in Table 1 is derived from Fig. 2, where each matrix entry represents the positive, negative or negligible influence of the column component on the row component. Positive entries indicate that an increase or decrease in one component triggers a corresponding directional change in the other. Thus, for example, an increase in weathering causes regolith mass and sediment yield to increase. Other positive relationships connect regolith availability to alluvial storage and yield. Negative entries in the interaction matrix denote that a change in the component leads to a change in the row component in the opposite direction. The table includes a negative link from yield to alluvium, and self-limiting effects for weathering. In fact, the regolith-to-weathering and alluvium-to-yield links can be positive or negative, depending on the thickness of the regolith and on the extent to which alluvial storage is available to buffer effects of changes in sediment supply on yield. These interrelationships are discussed in more detail by Phillips (2003).

The coefficients ($F_i$) of the characteristic Eqn. (5) can be determined from the entries of the interaction matrix ($z_{ij}$) using Eqn. (4). Phillips (2003) determined the conditions under which the first and second Routh–Hurwitz criteria will be met, and demonstrated that the stability of the drainage basin mass balance system is contingent on two requirements: first, regolith thickness must exert a negative influence on weathering rates by isolating the bedrock from the surface environment; and second, there must be a positive link from alluvium to yield, such that at times when inputs from uplands are reduced, material in or adjacent to the channel enhances sediment transport and yield. The regolith-weathering feedback is predicated on the notion that starting with a bare rock surface, a developing regolith cover enhances weathering by increasing moisture storage and contact time and facilitating biological activity (positive feedback). At some critical thickness, moisture and biological activity are no longer limiting, or are less so than exposure to $O_2$ and other surface phenomena. Thereafter the regolith-weathering feedback is negative, as a thicker regolith increasingly isolates the weathering front from the surface and slows the weathering rate. These relationships are discussed in more detail elsewhere (e.g. Minasny & McBratney, 1999; Heimsath et al., 2000; Anderson, 2002; Phillips et al., 2005). However, where negative feedback dominates it not only slows the rate of increase in regolith cover thickness, but also implies that regolith removal by erosion stimulates weathering which, in turn, partly offsets the depletion of the supply of erodible material.

External influences (e.g. tectonism, eustasy, climate and vegetation cover) are also important in the drainage basin

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**Table 1. Interaction matrix derived from Fig. 2.**

<table>
<thead>
<tr>
<th></th>
<th>$W$</th>
<th>$R$</th>
<th>$A$</th>
<th>$Y$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Weathering ($W$)</td>
<td>$-a_{11}$</td>
<td>$a_{12}$</td>
<td>0</td>
<td>$a_{14}$</td>
</tr>
<tr>
<td>Regolith ($R$)</td>
<td>$\pm a_{21}$</td>
<td>0</td>
<td>$a_{23}$</td>
<td>$a_{24}$</td>
</tr>
<tr>
<td>Alluvium ($A$)</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>$\pm a_{41}$</td>
</tr>
<tr>
<td>Yield ($Y$)</td>
<td>0</td>
<td>0</td>
<td>$-a_{41}$</td>
<td>0</td>
</tr>
</tbody>
</table>
mass balance. If these factors are not directly influenced by weathering, regolith accumulation, alluvial storage or sediment yield, or operate at time scales that are several orders of magnitude longer than the interactions included within the model, they have no effect on system stability (Schaffer, 1981; Phillips, 1986). However, their independence from the stability of the mass balance also allows the response of the basin to perturbations in these parameters to be assessed. We now assess these contingencies in the context of the Waipaoa River basin, in concert with the evidence for the (in)stability of basin sediment yield.

MODEL APPLICATION TO THE WAIPAOA RIVER BASIN

A variety of field evidence gives an indication of the likely status of the qualitative relationships between the different components of the drainage basin mass balance model as it is applied herein, and also permits us to modify Fig. 2 and derive the interaction matrix for conditions specific to the post-deforestation Waipaoa River basin.

Sediment production

The Waipaoa River’s high measured suspended sediment concentration (mean = 21.611 mg L\(^{-1}\)) and the sediment’s low intensity of chemical weathering (Sm : Na = 1.7; where Sm and Na are normalized to the composition of upper continental crust) are in accord with the global inverse relation between weathering intensity and sediment concentration reported by Gaillardet \(\text{et al.}\) (1999) (Fig. 3). This suggests that the suspended sediment currently in transport in the Waipaoa River at low to intermediate flows represents material generated by a denudational regime where physical as well as chemical weathering is prevalent, and only limited chemical weathering is a necessary precursor to mass removal. The largest amphitheatre-shaped gully complexes occur in association with crush zones along major faults and with argillite rocks that are readily weathered upon surface exposure and are especially susceptible to acid sulphate weathering (Claridge, 1960; O’Byrne, 1967; Gage & Black, 1979; Black, 1980; Pearce \(\text{et al.}\), 1981; McLeod \(\text{et al.}\), 1995). Nevertheless, the high proportion of smectitic clays in rocks and regoliths in the Waipaoa River basin and the incidence of gully erosion indicate the importance of physical weathering (Claridge, 1960; Gage & Black, 1979; DeRose \(\text{et al.}\), 1998; D’Ath, 2002; Marden \(\text{et al.}\), 2005). Shallow landsliding on soil-mantled, hillslopes influences both soil depth and terrain resistance, (Trustrum & DeRose, 1988; Crozier & Preston, 1998), and precludes a positive relationship between regolith thickness and weathering. Clay mineral and petrographic analyses of Waipaoa River sands indicate that contemporary bedload is dominated by sediment derived from the crushed and sheared sedimentary rocks of the East Coast Allochthon (D’Ath, 2002; James, 2003). The bulk sand composition is dominated by sedimentary lithic fragments, samples typically include rounded and subrounded grains, and beach sand in the vicinity of the Waipaoa River mouth is compositionally indistinguishable from fluvial sand. Gravel clasts are derived predominately from mudstone and siltstone lithologies (Rosser, 1997). The softer particles weather and abrade rapidly during transport, and the presence of rounded particles (Cory shape factor 0.45–0.70) throughout the river’s lower reaches is indicative of the repeated mobilization and movement of channel alluvium.

![Fig. 3. Relation between the degree of weathering (indexed by the ratio of Sm to Na normalized to the composition of upper continental crust) of modern fluvial sediment and the long-term suspended sediment concentration of selected world rivers (after Gaillardet \(\text{et al.}\) (1999)). The data point for the Waipaoa River is based on the analysis of 24 suspended sediment samples Gomez \(\text{et al.}\) (2003) obtained at Kanakania, at discharges of between five and 11 times the mean flow of 34.7 m\(^3\) s\(^{-1}\).]

Sediment yield

Gauging and proxy data compiled by Griffiths (1982), DeRose et al. (1998) and Hicks et al. (2000, 2004) show that, as is the case in many other drainage basins (Walling, 1983), there is an inverse relationship between suspended sediment yield and basin area that reflects the greater erosion rates recorded in small headwater basins where gully complexes may each generate the equivalent of 2–3% of the Waipaoa River’s total annual suspended sediment load (Fig. 4a). The organic carbon content of suspended sediment transported by the Waipaoa River also is consistent with the notion that, at the present time, gully erosion is the dominant process responsible for delivering sediment to the channel system during small, frequent events when there is no threshold limitation on sediment supply (Hicks et al., 2000; Gomez et al., 2003).

Except at times when the vegetation cover was perturbed by natural events such as wildfires, storms and volcanic eruptions (Grant, 1985; Wilmshurst et al., 1999), before deforestation the presence of a well-developed canopy and root network would have helped minimize soil loss from the landscape as a whole. Much of the sediment routinely would have been derived from hillslopes that were closely coupled to stream channels and riparian storage areas. Sediment accumulates in these areas during large storms, and subsequently is reworked and flushed from storage by more frequent, low magnitude events (Marutani et al., 1999). Indeed, variations in the organic carbon content of the flood plain alluvium indicate that before European deforestation bedrock outcrops on steep riparian, rather than soil-mantled hillslopes were a primary sediment source (Gomez et al., 2004a).

Variable suspended sediment yields per unit area from higher order tributary basins reflect differences in the rainfall regime and underlying geology. The latter in turn determines the dominance of the different erosion processes or combinations thereof (i.e. gully and sheet erosion, earthflow activity and shallow landsliding) that determine sediment availability (Jessen et al., 1999; Hicks et al., 2000, 2004). Thus, there are differences in the magnitude–frequency relations for tributary basins. For example, the relation for the Mangatu River, where gullies are prevalent, is somewhat flatter and plots higher than that for the Waipora River, which drains predominately landslide-prone terrain (Figs 1 and 4b). The offset between the two relations is indicative of overall sediment availability in the basin, and their gradients reflect relative sediment availability as a function of return period (Hicks et al., 2000, 2004). Consistent with the limitation on sediment supply during frequent (sub-annual return period) events that are below the threshold for landsliding, the relation for the Waipora River has a lower intercept and steeper gradient than that for the Mangatu River. In the latter river there is greater sediment availability during all events, especially during small, frequent events (return period ≤ 1 year). For the Waipaoa River basin as a whole, Hicks et al. (2000) showed that more than half the annual suspended sediment yield is transported by flows with a recurrence interval of <1 year. Gullies remain active and riparian sources contribute sediment during these events (Trustrum et al., 1999; Hicks et al., 2000; Gomez et al., 2003). Large inputs of material from hillslopes are minimized because the threshold for landsliding is not exceeded and the relation between infiltration capacity and runoff is such that overland flow is not a widespread phenomenon during small rainstorms. However, sediment availability increases dramatically during storms that exceed the threshold for shallow landsliding, and large storms affect suspended sediment yields for up to 3

Fig. 4. (a) Relation between suspended sediment yield and drainage basin area for gully complexes and tributary drainages in the Waipaoa River basin ($R^2 = 0.93$) derived using data compiled by Griffiths (1982), DeRose et al. (1998) and Hicks et al. (2000). (b) Magnitude and frequency relationships of event suspended sediment yields for the Mangatu and Waipora Rivers based on continuously recorded data (after Hicks et al. (2004)). See Fig. 1 for the location of the gauging stations and gully complexes.
years after the event (Hicks et al., 2000). Rapid inter-basin transport of fine sediment is facilitated by the short time (10-20 h) it typically takes the flood wave to travel the ~80 river km between gauging stations located in the headwaters and near the river mouth, and sediment sources and sinks are closely linked (Gomez et al., in press).

Sediment storage

For the past ~5.58 ka calendar years BP fine-grained alluvium has been accumulating on the Poverty Bay Flats (Pullar & Penhale, 1970; Brown, 1995; Gomez et al., 2004b). Before European settlement the rate of (vertical) infilling of the Poverty Bay Flats was ≤ 2.3 mm year⁻¹ (Pullar & Penhale, 1970; Gomez et al., 2004b), but once European settlers arrived and headwaters were deforested the rate of vertical accretion increased to 16 mm year⁻¹. Pullar & Penhale (1970) estimated that ~570 x 10⁶ m³ of alluvium were deposited on the Poverty Bay Flats between 3.47 ka calendar years BP and 1950 AD (when the construction of flood-control levees reduced the area susceptible to flooding by ~70%). This equates with a rate of ~0.164 x 10⁶ m³ year⁻¹ or ~0.213 Mt year⁻¹, using a conservative value of 1300 kg m⁻³ for the dry bulk density of the alluvium (Gomez et al., 1999; McLeod et al., 1999). On this basis, alluvial storage is equivalent to ~2% of the current annual suspended sediment load (~10.7 x 10⁶ t/year⁻¹) of the Waipaoa River at the upstream end of the Poverty Bay Flats (Hicks et al., 2000), a figure that is in keeping with Gomez et al.’s (1999) estimate (5%), which was obtained using data covering a much shorter (157-month) period. Phillips et al. (2007) found that the rate of remobilization or removal from storage is non-linear, and suggested that a significant portion (60–70%) of the recent alluvium likely will remain in storage on the floodplain for >100 years. The half-life for older (Holocene) alluvium is >2000 years but, by comparison with other fluvial systems (cf. Goodbred & Kuehl, 1999; Phillips & Slattery, 2006), neither the amount of storage nor the residence time of the alluvium is large.

The Poverty Shelf mud deposit has a predominately terrestrial origin and is derived primarily from the Waipaoa River basin (Gomez et al., in press). Based on an analysis of core MD972122 which was obtained at 38°48.67’S, 178°10.18’E, in 55-m-deep water (Fig. 1), changes in sediment source dynamics that occur at a variety of temporal and spatial scales leave distinctive signals in the shelf-sediment record. The signals record the landscape response to vegetation and land use change, fluctuations in climate and extreme storms and subduction–thrust earthquakes (Gomez et al., 2004b, in press). Moreover, despite a middle Holocene climate-forced change in circulation which led to more effective sediment retention on the continental shelf, after the major phase of Holocene downcutting in the Waipaoa River basin ended (~a. 5.58 ka calendar years BP) there was a marked decline in the terrigenous flux, from 0.48 to 0.21 g cm⁻² year⁻¹ (Carter et al., 2002; Gomez et al., 2004b). Thereafter, in the 5000 years before European deforestation, there was little variation in the rate of terrigenous mass accumulation on the continental shelf (Fig. 5). After the Waipaoa River basin was deforested water discharge and the amount of suspended sediment discharged to the shelf are estimated to have increased by ~4 and ~660%, respectively (Kettner et al., in press), and there was a ~350% increase in the rate of terrigenous mass accumulation on the middle shelf.

Basin mass balance

A stable mass balance implies limited variation in sediment yield over time, as is reflected in the rate of terrigenous mass accumulation on the continental shelf at core site MD972122 (Fig. 5). The stability of the mass balance system is contingent on a positive influence of regolith thickness on weathering rates, and a positive link from alluvium to yield, indicating that sediment is placed into and removed from alluvial storage in response to increases or decreases in sediment supply. This appears to have been
the case in the Waipaoa River basin before the native forests in the headwaters were cleared and the land converted to pasture in the late 19th and early 20th centuries, as slopes presumably maintained sufficient regolith cover to buffer underlying rock from both chemical and physical weathering. For the prehistoric period, a positive alluvium-yield link is also suggested by climatic forcing that has been directly linked to geomorphic change in the Waipaoa River basin (Berryman et al., 2000; Gomez et al., 2004b).

The mass balance model represented in Fig. 6 is applicable to conditions in the contemporary Waipaoa River basin, and the relationships in the drainage basin mass balance system for the post-deforestation (historic) period are summarized in Table 2. On upland hillslopes regolith is directly or indirectly derived from rock weathering in situ, weathering is limited by the availability of transportable regolith. However, the soft, highly erodible bedrock in the headwaters of the Waipaoa River basin may obviate its significance, and for the reasons noted below the regolith-to-weathering feedback is negative. Where transport capacity exceeds the supply of transportable debris (weathering-limited) all regolith produced eventually is removed, and to maintain the system in a transport-limited state, regolith production must keep pace with transport capacity. The sign of the links from regolith-to-alluvium and yield is reversed if soil stripping by shallow landsliding results in the net removal of regolith and, for the case of a declining regolith cover, yield and/or alluvial storage increase. This, however, has no effect on system stability (Phillips, 2003), although stability cannot be maintained in weathering-limited systems. The direct link from weathering to yield accommodates incision into bedrock. The alluvium-to-yield link is negative, and reflects a ‘competitive’ relationship. A positive alluvium-to-yield link, which likely obtains in many systems (Phillips, 2003), exists when yield is partly a function of the amount of alluvium available for transport (i.e. sediment yield is augmented by remobilizable alluvium during periods of low supply, and may decline if such material is unavailable). To the extent valley floors and stream beds are aggrading, storage moderates sediment yield (Gomez et al., 1999; Kettner et al., in press). However, the condition in which a low sediment supply is augmented by the remobilization of alluvium does not apply to the historic period, during which time the river system was overwhelmed by sediment generated in response to land use change (Gomez et al., 2001; Kettner et al., in press).

Feedback calculations

The feedback calculations Phillips (2003) performed for the mass balance system in general apply to the Waipaoa River basin into the 19th century, whereas the mass balance model represented in Fig. 6 is applicable to the post-deforestation (historic) period. Employing the interaction matrix for the post-deforestation (historic) period derived from Fig. 6 (Table 3) in conjunction with Eqn. (2):

\[ F_1 = -a_{11} < 0, \]
\[ F_2 = a_{12}(-a_{21}) + (-a_{34})(-a_{43}) > 0 \]

Table 2. Summary of interactions in the mass balance stability model for the Waipaoa River basin in the post-deforestation (historic) period (cf. Fig. 6 and Table 3).

<table>
<thead>
<tr>
<th>Interaction</th>
<th>Field evidence or reasoning</th>
</tr>
</thead>
<tbody>
<tr>
<td>Negative self-effects on weathering</td>
<td>Weathering rates are inherently limited by factors such as climate, lithology, geochemical kinetics and the depletion of unstable minerals</td>
</tr>
<tr>
<td>Positive effects of weathering on regolith thickness</td>
<td>Regolith is directly or indirectly derived from rock weathering</td>
</tr>
<tr>
<td>No direct link from weathering to alluvium</td>
<td>Relationships are indirect, via erosion and deposition processes</td>
</tr>
<tr>
<td>Positive link from weathering-to-yield</td>
<td>Gully incision into bedrock</td>
</tr>
<tr>
<td>Positive link from regolith-to-weathering</td>
<td>The thin soil cover in some sediment source areas does not inhibit weathering</td>
</tr>
<tr>
<td>Positive effects of regolith on alluvial storage, and sediment yield</td>
<td>A thicker regolith cover implies greater amounts of transportable material are available for delivery to the fluvial system</td>
</tr>
<tr>
<td>Negative effects of alluvial storage on yield; negative yield-to-alluvium link</td>
<td>Competitive relationship obtains during the post-deforestation (historic) period of rapid aggradation</td>
</tr>
<tr>
<td>No direct links from alluvial storage or yield to weathering or regolith</td>
<td>Any relationships are indirect, via influences on valley and hillslope evolution</td>
</tr>
</tbody>
</table>
Weathering (\(W\)) including earthflows and sheet erosion, to one where incipient-dominated by landsliding and other diffusive processes, change in process dominance, from an erosional regime opening hillslopes and precipitating the well-documented transition to pasture. The change in land use profoundly affected geo-morphologic forcings in the prehistoric period, which constituted a hierarchy of temporally sensitive phenomena whose impact is conditioned by events of low frequency and high magnitude (cf. Pain & Hoskin, 1970; Thomas, 2000), had little effect on sediment delivery to the shelf (Fig. 5). The very high contemporary erosion rates are a product of this spatially sensitive change in land use, which had a direct impact on sediment source areas throughout the headwaters and re-defined the sensitivity of the landscape to erosion by precipitating the transition to an erosional regime that impacted sediment production and dispersal across the entire magnitude–frequency spectrum of events that regulate sediment delivery to and transport in stream channels (Hicks et al., 2000). For this reason the historic change in erosion and sediment transport within the basin registers clearly against, rather than being masked by the metabolic rate (despite steep slopes and active up-lift). Deforestation not only increased upland sediment production, but also affected the regolith thickness–weathering relationship by reducing or removing regolith cover. The filling of the alluvial valleys meant that sediment fluxes could not be moderated by alluvial storage. These changes lead to dynamical instability in the mass balance system, and heighten the sensitivity of sediment yield to changes in sediment production and transport within the basin. Accordingly, despite extensive alluvial storage on the Poverty Bay flats, aggradation of river channels throughout the basin in the historic period, and intermittent transfers into and out of storage in the headwaters (Pullar & Penhale, 1970; Gomez et al., 2004b). This is the reverse of what occurred after the end of the major phase of Holocene downcutting, when there was a marked decline in the rate of terrigenous mass accumulation on the continental shelf (Fig. 5). The very high contemporary erosion rates are a product of this spatially sensitive change in land use, which had a direct impact on sediment source areas throughout the headwaters and re-defined the sensitivity of the landscape to erosion by precipitating the transition to an erosional regime that impacted sediment production and dispersal across the entire magnitude–frequency spectrum of events that regulate sediment delivery to and transport in stream channels (Hicks et al., 2000). For this reason the historic change in erosion and sediment transport within the basin registers clearly against, rather than being masked by the metabolic rate (despite steep slopes and active up-lift). Deforestation not only increased upland sediment production, but also affected the regolith thickness–weathering relationship by reducing or removing regolith cover. The filling of the alluvial valleys meant that sediment fluxes could not be moderated by alluvial storage. These changes lead to dynamical instability in the mass balance system, and heighten the sensitivity of sediment yield to changes in sediment production and transport within the basin. Accordingly, despite extensive alluvial storage on the Poverty Bay flats, aggradation of river channels throughout the basin in the historic period, and intermittent transfers into and out of storage in the headwaters (Pullar & Penhale, 1970; Gomez et al., 1999; Trustrum et al., 1999; Phillips et al., 2007), alluvial buffering does not moderate sediment delivery to the continental shelf.

**DISCUSSION AND INTERPRETATIONS**

After the major phase of Holocene downcutting ceased, and the present climatic regime and indigenous vegetation were established the mass balance system stabilized. Neither short-term fluctuations in climate, nor the disturbance of the native forests by airfall ash from volcanic eruptions, fire and cultural activities in the prehistoric period appear to have profoundly impacted sediment output from the Waipaoa River basin, at least in the long term (Kettner et al., in press). This is reflected in the rate of accumulation of terrigenous sediment on the Poverty Shelf during the past 5000 years (Fig. 5). Gunnell (1998) suggested that relief and uplift provide critical first-order controls of mechanical denudation, and give rise to a ‘metabolic’ rate of erosion. Upward deviations from this rate, due to changes in erosion associated with changes in climate, tectonics, land use, etc. likely occur only in relatively short bursts, so that the metabolic rate dominates over long periods. This helps explain why natural and anthropogenic forcings in the prehistoric period, which constitute a hierarchy of temporally sensitive phenomena whose impact is conditioned by events of low frequency and high magnitude (cf. Pain & Hoskin, 1970; Thomas, 2000), had little effect on sediment delivery to the shelf (Fig. 5). The situation changed dramatically in the late 19th and early 20th centuries, as the indigenous forests in the headwaters were cleared by European settlers and the land converted to pasture. The change in land use profoundly affected geomorphic processes and sediment production by over-steepening hillslopes and precipitating the well-documented change in process dominance, from an erosional regime dominated by landsliding and other diffusive processes, including earthflows and sheet erosion, to one where incisive processes generated most of the sediment transported by the Waipaoa River (Hicks et al., 2000; Gomez et al., 2004b). This is the reverse of what occurred after the end of the major phase of Holocene downcutting, when there was a marked decline in the rate of terrigenous mass accumulation on the continental shelf (Fig. 5). The very high contemporary erosion rates are a product of this spatially sensitive change in land use, which had a direct impact on sediment source areas throughout the headwaters and re-defined the sensitivity of the landscape to erosion by precipitating the transition to an erosional regime that impacted sediment production and dispersal across the entire magnitude–frequency spectrum of events that regulate sediment delivery to and transport in stream channels (Hicks et al., 2000). For this reason the historic change in erosion and sediment transport within the basin registers clearly against, rather than being masked by the metabolic rate (despite steep slopes and active up-lift). Deforestation not only increased upland sediment production, but also affected the regolith thickness–weathering relationship by reducing or removing regolith cover. The filling of the alluvial valleys meant that sediment fluxes could not be moderated by alluvial storage. These changes lead to dynamical instability in the mass balance system, and heighten the sensitivity of sediment yield to changes in sediment production and transport within the basin. Accordingly, despite extensive alluvial storage on the Poverty Bay flats, aggradation of river channels throughout the basin in the historic period, and intermittent transfers into and out of storage in the headwaters (Pullar & Penhale, 1970; Gomez et al., 1999; Trustrum et al., 1999; Phillips et al., 2007), alluvial buffering does not moderate sediment delivery to the continental shelf.

Although alluvial storage in the lower reaches of the meso-scale Waipaoa River basin is extensive, it is not particularly large relative to recent sediment transport in the river (Gomez et al., 1999; Gomez & Trustrum, 2005). Nor is the apparent residence time of the alluvium large relative to time scales of Holocene and Quaternary environmental change. This is the opposite of what is observed in fluvial systems where alluvial buffering has been shown to be important (Meißner & Gaudemar, 1999; Phillips, 2003). In the Trinity River, Texas, for example, remobilizable alluvial storage in the lower river represents about 140 000 years of the average annual sediment yield (Phillips et al., 2004), compared with ~50 years for the Waipaoa River. There is also ample evidence to suggest that hinterland to margin transfer is accomplished very rapidly (Gomez et al., in press). Such rapid transfer is equally uncharacteristic of coastal plain rivers with strong alluvial buffering (Phillips et al., 2007). Under these circumstances the first Routh–Hurwitz criterion is violated and the system is dynamically unstable. Owing to the signs of the constituent terms \(F_3\) and \(F_4\) must be positive, and \(F_2 > 0\) if the alluvium-to-yield feedbacks are stronger than the regolith-to-weathering links. This is the case in the Waipaoa River basin, due to the strength of the former and the weakness of the latter.

| Interaction matrix derived from Figure 6. Weak links (dotted lines in Fig. 6) are shown in parentheses. |
|---------------------------------|-----------------|-------|-------|
| \(W\) (Weathering) \(- a_{11}\) | \(R\) (Regolith) \(a_{12}\) | \(A\) (Alluvium) \(a_{33}\) | \(Y\) (Yield) \(- a_{41}\) |
| \(F_3 = -[(- a_{43})(- a_{34})(- a_{11})] > 0\) | \(F_4 = -[a_{12}(- a_{21})(- a_{43})(- a_{34})] > 0\) |
where the small amounts of upper-basin sediment that reach the lower river are overwhelmed by lower-basin sediment sources.

Finally, it should be noted that studies and conceptual models of feedbacks between regolith cover, erosion, and weathering often incorporate the notion that bedrock has very low erodibility, and that a significant amount of time is needed for weathering to produce transportable debris (e.g. Minasny & McBratney, 1999; Heimsath et al., 2000; Phillips, 2003). These concepts are of limited applicability to the Waipaoa River basin, where gully erosion (incision into bedrock) constitutes an important sediment source. Much of the underlying bedrock also weathers so rapidly on exposure that there may be no significant time lag between regolith stripping and the point when this debris becomes available for transport.

**CONCLUSIONS**

The unglaciated, tectonically active Waipaoa River basin has been influenced by Late Quaternary and Holocene climate change, coseismic and volcanic events, and anthropogenic land cover change (Gomez et al., in press). Sediment production is focused on the headwaters, channels throughout the basin are aggrading, and the lower portion of the basin is characterized by extensive alluvial sediment storage on the Poverty Bay Flats. The drainage basin mass balance stability model provides a holistic framework for interpreting sediment production, storage, and transport in this steepland river system in which, despite the strong first-order geologic controls on erosion, and extensive alluvial storage, the flux of fluvial sediment to the ocean is sensitive to hillslope destabilization driven by land use change in the last ~100 year. As indexed by the rate of terrigenous mass accumulation on the continental shelf, the delivery of fluvial sediment to the ocean was not greatly affected by either natural or anthropogenically induced environmental change in the late Holocene (pre-historic period) before deforestation by European settlers, and the mass balance model suggests that the system was stable for much of the past 5000 years. The impact of these temporally sensitive changes probably was felt at the large magnitude, low-frequency end of the event spectrum, and the stability of the system likely was facilitated by periodic changes in alluvial storage. By contrast, an analysis of the mass balance model for conditions typical of the contemporary Waipaoa River basin much of the sediment supplied to stream channels is readily transportable, and the remobilizable alluvium on the flood plain represents only ~50 year of average annual suspended sediment yield (which is orders of magnitude lower than the corresponding figure for passive-margin coastal plains rivers where alluvial buffering is known to have a significant effect on basin sediment yield). The implication is that the strength of the signal of historic erosion in Waipaoa River basin recorded in the marine depocentre may not only be due to the intensity of the erosion processes involved, but also to the fact that the land use change destabilized the drainage basin mass balance system. This observation has important ramifications for models that seek to retrodict or predict the behaviour of fluvial systems over extended time periods, because it suggests that the impact of climate or land use change on sediment yield, for example, should be assessed not only in the context of the processes that lead to the production of transportable debris (and its removal), but also in terms of the interactions that occur between this material, its storage on hillslopes and in channels or floodplains, and sediment yield.

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