Reconstructing high-magnitude/low-frequency landslide events based on soil redistribution modelling and a Late-Holocene sediment record from New Zealand

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Abstract

A sediment record is used, in combination with shallow landslide soil redistribution and sediment-yield modelling, to reconstruct the incidence of high-magnitude/low-frequency landslide events in the upper part of a catchment and the history of a wetland in the lower part. Eleven sediment cores were obtained from a dune-impounded wetland at Te Henga, west Auckland, northern New Zealand. Sediment stratigraphy and chronology were interpreted by radiocarbon dating, foraminiferal analysis, and provisional tephrochronology. Gradual impoundment of the wetland began c. 6000 cal yr BP, coinciding with the start of a gentle sea-level fall, but complete damming and initial sedimentation did not begin until c. 1000 cal yr BP. After damming, four well-defined sediment pulses occurred and these are preserved in the form of distinct clay layers in most of the sediment cores. For interpreting the sediment pulses, a physically based landslide model was used to determine spatially distributed relative landslide hazard, applicable at the catchment scale. An empirical landslide-soil redistribution component was added and proved able to determine the volumes and spatial pattern of eroded and deposited soil material, sediment delivery ratio and the impact on total catchment sediment yield. Sediment volumes were calculated from the wetland cores and corresponding landslide scenarios are defined through back-analysis of modelled sediment yield output. In general, at least four major high-magnitude landslide events, both natural and intensified by forest clearance activities, occurred in the catchment upstream of Te Henga Wetland during the last c. 1000 years. The spatial distribution of modelled critical rainfall values for the catchment can be interpreted as an expression of shallow landslide hazard. The magnitude of the sediment...
1. Introduction

In many geomorphological settings, shallow landsliding is one of the most important components of hillslope denudation and therefore can play an important role in determining catchment sediment yield. Because the actual triggering of shallow landslides by rainfall events, and the consequent amounts of erosion and deposition, are highly dependent on (natural or human-induced) land use and land-cover changes, there is an increased need for methodologies that can assess the effects of these changes on landslide occurrence and catchment sediment yield (Burton and Bathurst, 1998; Glade, 2003). Furthermore, there is a lack of understanding of the possible linkages between climate change and corresponding change of geomorphic activity and resulting sediment yield (Evans and Slaymaker, 2004). The relation between sediment production in upland areas and the sediment yield at a basin outlet has been the subject of research for over half a century (Glymph, 1945). Understanding the connections between cause and response, however, remains far from complete (Trustrum et al., 1999). Despite a relatively good understanding of the mechanics of individual landslides (e.g. Selby, 1993), few studies have analysed the cumulative effects of soil redistribution by landsliding over large spatial and (or) temporal scales (Martin et al., 2002; Claessens et al., in press). As sediment derived from landsliding is generated and transported mainly during extreme rainfall triggering events with a low frequency of occurrence, studying the link between spatial and temporal occurrence of these events and the resulting sediment yield is of great interest.

1.1. Magnitude-frequency of landsliding

A major obstacle when assessing rates of landsliding is the difficulty of obtaining data that are relevant over medium to long time scales. Longer term magnitude-frequency distributions of landslides are usually estimated from rates over decadal time scales derived from large inventories of aerial photographs (Hovius et al., 1997; Martin et al., 2002). Analytic expressions such as power-law models are then fitted through probability density functions for empirically derived landslide properties such as area or volume. Other researchers try to link climate data with landslide inventories and identify e.g. the magnitude of storms that trigger landslides (Crozier, 1996a; Glade, 1996). Relationships between the frequency of landslide-generating storms and mean annual rainfall (Hicks, 1995) or between rainstorm magnitude and landslide frequency (Reid and Page, 2002) have also been established.

In general, the temporal resolution for most magnitude–frequency analyses is rather short for the largest events to be properly represented. Sediment yield resulting from landsliding within a catchment could be translated into process magnitude and frequency if it is transported out of the catchment and trapped in a lake or swamp. Lake or swamp sediments are the product of the environmental processes, physical, biological and chemical, that have been operating within the surrounding catchment. They provide a record of the timing and magnitude of environmental processes, both natural and human-induced (Goff et al., 1996; Page and Trustrum, 1997). As opposed to hillslopes, where contemporary processes destroy the evidence of earlier erosion events, lake sediments have the potential to provide an undisturbed record over a long time frame (Brunsden, 1993). When large rainstorms are the main cause of erosion, as is the case in many New Zealand steeplands (Page et al., 1994a; Glade, 2003), lake sediments can provide more information about the cumulative effects of these episodic events (Page and Trustrum, 1997; Trustrum et al., 1999). Furthermore, recent research on sediment budgets, river discharge and suspended sediment load suggests that most of the sediment transported by streams and deposited in lakes or swamps originates...
from landslides and landslide-gully complexes in the upland catchment (e.g. Page et al., 1994a; Eden and Page, 1998; Hicks et al., 2000). Especially in smaller basins, magnitude–frequency relationships for landslide erosion and sediment deposition seem to be closely related and high-magnitude/low-frequency landsliding events are often responsible for most of the deposition (Hovius et al., 1997; Trustrum et al., 1999). In larger basins, by contrast, landsliding makes a smaller relative contribution to catchment suspended sediment yield than does that arising from other processes (e.g. gullies, sheetwash, and fluvial erosion).

1.2. Previous work on sediment yield

Many researchers have used sediment records preserved in lakes or swamps to assess magnitude and frequency of sediment production in the upland area and the resulting sediment yield (Owens and Slaymaker, 1993; Page et al., 1994a,b, 2004). Sediment records of lakes are studied mostly by extracting and interpreting sediment cores. The stratigraphy and chronology of these cores is often established by a combination of techniques which include radiocarbon dating, $^{210}$Pb and $^{137}$Cs dating (Goff et al., 1996; Newnham et al., 1998; Trustrum et al., 1999), pollen and diatom analysis (Page and Trustrum, 1997; Sandiford et al., 2003), and tephrochronology (Lowe and Green, 1987; Eden and Page, 1998; Evans and Slaymaker, 2004).

Calculation of sediment yields is conventionally undertaken by converting sediment thicknesses in individual cores to volumes by averaging across the lake area (Foster et al., 1990). Other approaches incorporate spatial variability within a lake by combining individual core volume estimates with the area of thiessen polygons constructed around the core location (O’Hara et al., 1993). Evans and Slaymaker (2004) used a regression model of the accumulation surface to predict sediment accumulation with error intervals for each core.

The relation between upland erosion and sediment yield is complex because not all material detached from hillslopes will reach the sediment reservoir (Owens and Slaymaker, 1993). Material deposited in a reservoir may represent only a small fraction of that mobilized within the catchment (Walling, 1983). Several sources of error occur when reservoir sedimentation data are transferred into sediment yield and erosion rates for the upland catchment (see review in Butcher et al., 1993). To make this conversion, it is necessary to know the sediment delivery ratio, the ratio between total erosion on hillslopes within a catchment, and sediment delivery to the stream network. Delivery ratios are scale-dependent and catchments of $<10$ km$^2$ show maximum sediment yields per unit area compared with intermediate yields in the largest systems (Butcher et al., 1993). Increasing rainfall also influences the delivery ratio: during high-magnitude events, ratios are usually very high (up to 0.7, Trustrum et al., 1999). Variations in sediment delivery ratio have also been studied to assess the relation between land use change and erosion (Page et al., 1994a; Page and Trustrum, 1997; Trustrum et al., 1999; Glade, 2003).

Several studies have attempted to link extrapolated erosion rates from measurements at a particular site (sediment loads) to the total catchment sediment yield (Caine and Swanson, 1989; Hattanji and Onda, 2004). Dearing and Foster (1993), however, argued that a monitored record of sediment yield may remain constant or may fluctuate wildly without giving any clues as to what is actually happening in the catchment. Other researchers have analysed landslide characteristics from large field and (or) aerial photograph inventories to estimate landslide volumes, frequencies and sediment yields (e.g. Hovius et al., 1997; Martin et al., 2002). Other methods for estimating total landslide sediment volumes include statistical (sample) techniques applied to field data (Megahan et al., 1991; Page et al., 1994a), or cut-and-fill calculations based on pre-landslide and post-landslide surfaces from a digital elevation model (DEM) (Korup et al., 2004).

Only a few studies have used erosion/sedimentation modelling to calculate landslide sediment volumes and/or sediment yields from a catchment. Spatially distributed erosion and sedimentation models have long been restricted to processes dealing with surface erosion by overland flow (e.g. Wicks and Bathurst, 1996). Istanbulluoglu et al. (2004) modelled the effects of forest vegetation and disturbances on total sediment production by several erosion processes (runoff, creep, gully erosion and landsliding). Recently, models have been developed to assess the spatial patterns and effects of landslide erosion and deposition within a catchment (e.g. Claessens et al., in...
press); some models also deal with the resulting sediment yield and delivery ratio (Burton and Bathurst, 1998). Bathurst et al. (1997) tested several empirical modelling approaches for determining the delivery ratio of landslide sediment to streams and the resulting sediment yield. These models are simple relationships that estimate landslide runout distance from slope geometry. Another statistical model estimating sediment delivery directly has been tested. In general, none of the models was completely accurate.

Burton and Bathurst (1998) built on Vandré’s (1985) model to estimate runout distance from hillslope geometry, and incorporated this formula in a spatially distributed landslide hazard and soil redistribution model. Regarding the spatial pattern of landslide sediment delivery, the approach used in this paper is based on the principles of this model (see further below). A sediment record is analysed and, in combination with shallow landslide soil redistribution and sediment-yield modelling, a reconstruction is made of the incidence of high-magnitude/low-frequency landslide events in the upper part of the Waitakere River catchment and the history of the Te Henga wetland at the basin outlet. We test the ability of the LAPSUS-LS model (Claessens et al., in press) to estimate the delivery ratio and sediment yield as a result of multiple landslides at the catchment scale (magnitude). The stratigraphy and chronology of the wetland sediment record provides control over the frequency of high-magnitude landslide events.

2. Study area

2.1. Waitakere river catchment

The Waitakere River catchment lies in the Waitakere Ranges Regional Park, west of Auckland, North Island, New Zealand (174.8°E, 36.9°S, Fig. 1). Altitude ranges from sea level to 474 m. The catchment is approximately 31 km² and its mean elevation is 175 m asl. The area has a warm and humid climate with a mean annual rainfall ranging from ~1400 mm near the coast to 2030 mm at higher altitudes (Auckland Regional Council, 2002). After forest clearance by early European farmers during the second half of the 19th century (Hayward and Diamond, 1978), much of the study area is now covered with regenerating native vegetation and patches of undisturbed rainforest. The catchment lithologies consist mainly of early Miocene andesitic volcaniclastic submarine deposits of Piha Formation and associated terrestrial andesite flows of Lone Kauri Formation, all within the Manukau Subgroup of the Waitakere Group (Hayward, 1976; Edbrooke, 2001). The landscape is mantled by deep soils which are classified as Haplic Acrisols in the World Reference Base classification (Deckers et al., 2002). The clay fractions of the soils are dominated by kaolinite but varying amounts of smectite, halloysite and vermiculite are also present. Although landslides are usually less common under forest than under e.g. pasture, because of protection by vegetation cover and reinforcement by tree root systems (Phillips and Watson, 1994), they are still a dominant erosion process in the study area (Hayward, 1983; see other examples in Crozier et al., 1992; Eden and Page, 1998; Moon et al., 2003). The Waitakere Reservoir was constructed on the headwaters of the Waitakere River in 1910 to contribute to Auckland’s water supply. This obviously affected the natural river flow and reduced both the mean stem flow and peak flows (Watercare Services Ltd., 2001).

2.2. Te Henga wetland

The Te Henga wetland (also known as Bethell’s swamp) is situated in the northern part of the Waitakere Ranges and forms the outlet of the Waitakere River to the Tasman Sea. The wetland is ~1.7 km² and was impounded by a landward-prograding dune complex in a similar way to that for nearby Lake Wainamu (Fig. 1). The Late Holocene coastal ‘blacksand’ of the dune complex (Mitiwai Sand, Hayward, 1983; Karioitahi Group, Isaac et al., 1994) is present in beach and dune deposits along the west coast of the Waitakere Ranges. The sands are erosion products primarily of Quaternary andesitic volcanic and volcanoclastic rocks of western Taranaki and the central North Island. They have been transported along shore by shallow marine currents and were subsequently concentrated by wave and wind action into beach and dune lag deposits (Edbrooke, 2001). The moving dune fields have probably accumulated within the last c. 1300 cal years based on entrapment of Te Henga by the equivalent of dune belt 4 (c. 1500–300 14C years...
ago), as described by Schofield (1975) and Lowe and Green (1987) for the South Kaipara Barrier that lies to the north. The wetland has a very high ecological value and its extent and quality are of regional significance in the Auckland area (Denyer et al., 1993). Archaeological evidence suggests that forest around the wetland was cleared during early Polynesian (Maori) settlement around the margins of the wetland (Diamond and Hayward, 1979; Hayward and Diamond, 1978). It was filled substantially with sediments from upstream by the time of the arrival of the first European settlers in the early 19th century (Waitakere Ranges Protection Society, 1979). A flax mill operated around the head of the wetland from 1880–1890 (Hayward and Diamond, 1978). In the 1920s, kauri tree logging and milling activities took place and a launch towed logs through the wetland (Diamond and Hayward, 1980). Whereas the Waitakere Dam upstream reduces peak flows in the wetland at present, the flood storage available within the wetland system has a greater impact on hydrological conditions than the influences of the dam (Watercare Services Ltd., 2001). Near the outlet of the wetland, the small Mokoroa and Wainamu streams also drain into the wetland but contribute little to the sediment delivery because of low geomorphic activity and damming by dune-impounded Lake Wainamu upstream, respectively.

3. Methods

3.1. LAPSUS modelling framework

The LAPSUS modelling framework was developed to study the long term effects of geomorphic
processes on the landscape scale (Schoorl et al., 2000). To include the effects of erosion and sedimentation by landsliding, the LAPSUS-LS component was embedded in the model (Claessens et al., in press). Effects of shallow landsliding within a time-frame of years to decades are simulated in a dynamic way by adapting digital elevation data between yearly timesteps, according to the calculated soil redistribution (i.e. implementing landslide erosion and deposition). The overall aim of LAPSUS-LS is to assess the impact of landsliding on landscape evolution and to identify possible feedbacks with other geomorphic processes; it is not intended to simulate detailed changes in hillslope geomorphology caused by individual failures. Bearing in mind its assumptions and limitations, this approach has been demonstrated to retain the essence of the physical control of topography and soil properties on landsliding and remains parametrically simple for ease of calibration, validation and application.

LAPSUS-LS consists of several modelling steps. Relative landslide hazard distribution is calculated from topographical and geotechnical attributes. Historical rainfall-landslide distribution datasets and magnitude-frequency scenarios can then be used to calibrate and run the model for consecutive timesteps. Soil redistribution algorithms are applied to visualise feedbacks between mass movements or interactions with other hillslope processes. Although the model was originally not intended to quantify erosion or sedimentation, we added a spatially explicit sediment delivery algorithm to simulate scenarios of landslide sediment yield at the catchment level.

3.1.1. Relative hazard for shallow landsliding

For the analysis of shallow landslide hazard, a steady state hydrologic model is combined with a deterministic infinite slope stability model. This approach has been described previously by Montgomery and Dietrich (1994), and has performed well in many applications (e.g. Montgomery et al., 2000; Pack et al., 2001; Claessens et al., in press). We calculated the minimum steady state critical rainfall predicted to cause slope failure, $Q_{cr}$ [m day$^{-1}$], which can be written as:

$$Q_{cr} = T \sin \theta \left( \frac{b}{a} \right) \left( \frac{\rho_s}{\rho_w} \right) \left[ 1 - \left( \frac{\sin \theta - C}{\cos \theta \tan \phi} \right) \right]$$

(1)

where $T$ is saturated soil transmissivity [m$^2$ day$^{-1}$], $\theta$ is local slope angle [°], $a$ the upslope contributing drainage area [m$^2$], $b$ the unit contour length (the grid resolution [m] is taken as the effective contour length as in Pack et al. (2001), $\rho_s$ wet soil bulk density [g cm$^{-3}$], $\rho_w$ the density of water [g cm$^{-3}$], and $\phi$ the effective angle of internal friction of the soil [°]. $C$ is the combined cohesion term [-], made dimensionless relative to the perpendicular soil thickness and defined as:

$$C = \frac{C_t + C_s}{h \rho_g g}.$$  

(2)

With $C_t$ root cohesion [N m$^{-2}$], $C_s$ soil cohesion [N m$^{-2}$], $h$ perpendicular soil thickness [m], and $g$ the gravitational acceleration constant (9.81 m s$^{-2}$). The spatial distribution of critical rainfall values calculated according to Eq. (1) can be interpreted as an expression of the potential for shallow landslide initiation.

3.1.2. Failed landslide material redistribution

To determine landslide soil redistribution within a catchment, sophisticated formulae applicable to well-specified individual failures are inappropriate and simpler, empirical formulae were developed (Claessens et al., in press). Following the initial failure, in the erosional phase, an amount of unstable soil material $S$ [m] is eroded following the steepest descent direction and estimated as:

$$S = \frac{\rho_s \cos \theta (\tan \theta - \tan z)a}{C_s}$$

(3)

with $z$ [°] minimum local slope for landslide erosion and $a$ [m$^2$] a dimensional correction factor. The point at which deposition begins is reached once the gradient falls below an area specific slope angle $z$. Although the most critical factor in dictating runout distance is the volume of the initial failure (Crozier, 1996b), in this approach the elevation loss within the erosional phase is used as a measure of momentum at the start of deposition. The number of down slope grid cells involved in the deposition of landslide material, defined ‘cell-distance’ $D$ [-], is calculated as:

$$D = \left( \frac{\Delta v \phi}{b} \right)$$

(4)
where \( b \) is the grid resolution [m], \( \Delta y \) [m] is the elevation difference between the head of the slide and the point at which deposition begins, and \( \varphi \) [-] is an empirically derived ‘runout fraction’ (Vandre, 1985; Burton and Bathurst, 1998). To incorporate hillslope morphology into the spatial deposition pattern, the accumulated soil material is routed with multiple flow principles: for down slope neighbours of the point where deposition starts we expressed the sediment, which is effectively delivered to grid cell \( i \), as:

\[
S_i = \left( \frac{B_{i-1}}{D_{i-1}} \right) f_i. \tag{5}
\]

The term \( B_{i-1}/D_{i-1} \) is the amount of sediment derived from erosion upslope (grid cell \( i-1 \)), divided by the cell-distance (Eq. (4)), and deposited in grid cell \( i \). The remaining sediment budget of grid cell \( i \), which is not deposited but ‘passed through’ to grid cell \( i+1 \), can be expressed as

\[
B_i = B_{i-1} \left( 1 - \frac{1}{D_{i-1}} \right) f_i. \tag{6}
\]

In Eqs. (5) and (6), \( f_i \) is the fraction allocated to each lower neighbour and determined by the multiple flow concept described by Quinn et al. (1991). In each down-slope grid step, the cell-distance is lowered by one and when \( D<1 \) all the remaining sediment is deposited.

### 3.1.3. Delivery ratio and sediment yield

The impact of landsliding on basin sediment yield depends on whether the eroded material is deposited in, and transported by, the stream network. The percentage delivery or delivery ratio is dependent on the interaction between landslide soil redistribution patterns and channels able to route and transport the material further towards the catchment outlet. Instead of estimating or extrapolating delivery ratios and sediment yields from site measurements or large field inventories, we determined the amount of sediment yield from the modelled spatial pattern of soil redistribution and the consecutive interaction with a topographically delineated stream network. Landslide material displacement was modelled using Eqs. (3)–(6). For determining the stream network, different methods are available, ranging from specified contributing area and/or slope thresholds (e.g., O’Callaghan and Mark, 1984; Martin et al., 2002) to the use of upward curved grid cells (Tarboton, 2000) and grid network pruning by order (Peckham, 1998). For our application, the sediment transporting stream network was determined by simply specifying a minimum contributing area threshold. Flow direction was assigned according to the steepest descent, and flow accumulation was calculated as a measure of the drainage area in number of grid cells (this method is typically called ‘D8’ algorithm, Fairfield and Leymarie, 1991). All grid cells draining more than a threshold drainage area are defined as part of the stream network and able to transport landslide material to the catchment outlet. When a grid cell, which is part of the depositional pathway of a landslide, intersects with a grid cell from the transporting stream network, the remaining sediment budget of that grid cell, according to Eq. (6), is added to the catchment sediment yield.

By modelling the spatial pattern of landslide soil redistribution and the interaction with the channel network, buffering of the depositional response by temporary storage of landslide material on foottops is taken into account. If the depositional pathway does not cross a transporting channel, the landslide material is not delivered to the outlet but remains on the slope and hence excluded from the sediment yield. The delivery ratio is also determined by use of this method (and does not have to be estimated): deposition that occurs out of reach of a channel able to transport the material is not added to the sediment yield.

Determining the stream network by assigning a threshold value of contributing drainage area, calculated from the DEM, implies that delineated streams are assumed to be able to transport the sediment in its entirety to the catchment outlet. Field evidence supports this assumption — even shortly after sediment-producing events, the streams in the study catchment contain little suspended sediment and stream beds appear as bare rock.

### 3.1.4. Parameterisation and calibration of LAPSUS-LS

Data requirements necessary for applying the model are good quality topographical information and some geotechnical soil parameters for use in Eqs. (1)–(3). A DEM with a 25-m grid resolution was derived from vector line and point data sourced
from the topographic database from Land Information New Zealand. Possible effects of choice of DEM resolution on the results of the model are discussed in Claessens et al. (2005). Attributes derived from the DEM are the local slope $h$ and the upslope contributing drainage area $a$, computed using the algorithm of multiple downslope flow (Quinn et al., 1991). Values for $T$, $C$, $q_s$ and $/F$ for the two main parent materials of the study area are based on field and laboratory measurements (Table 1); saturated shear strength of the soil has been determined by consolidated-drained direct shear tests on undisturbed samples taken from soils developed in the two parent materials of the catchment. All parameters for Eq. (1), except slope and contributing area, are grouped within areas with soils developed in the same parent material.

Root strength of both tree and understory vegetation provides significant apparent cohesion to the soil. Root cohesion is very hard to quantify, certainly spatially distributed on the catchment scale, and in our approach it is used to calibrate the model regarding the spatial distribution of slope failures. Back analysis of landslides mapped in the field and by aerial photography interpretation made it possible to calibrate the model for our study area by adapting the root cohesion $C_r$ in the combined cohesion term $C$ (Eq. (2)) for different vegetation classes (Claessens et al., in press). Calibrated values of $C$ are shown in Table 2. The default settings of the empirical parameters used in the soil redistribution algorithms (3) to (6) are based on field evidence and literature and further subjected to a sensitivity analysis. The ‘runout fraction’ $\varphi$ was set at 0.4 and the slope angle $z$, at which deposition begins, was set at 10° (Vandre, 1985; Burton and Bathurst, 1998; Claessens et al., in press). The threshold value of contributing area for stream development and the threshold critical rainfall for landslide initiation are both scenario dependent and also further analysed regarding model sensitivity.

Calibration concerning the location of landslide initiation sites has been undertaken on the basis of fieldwork and a series of aerial photographs covering a limited timeframe. It is very likely that high-magnitude/low-frequency events are not all correctly represented and are underestimated in the dataset. Building model scenarios based on the long-term sediment record of the wetland will give an indication about the relative importance of these events over time.

3.2. Sediment record analysis

3.2.1. Core stratigraphy and chronology

A corer with a diameter of 2 cm and extends up to 5 m in total length was used in February and March, 2003, to obtain a detailed record of sediments across the wetland (Fig. 2). Stratigraphy, texture, colour, sorting, degree of weathering and type of organic material and content were described.

The chronology of the cores was established using radiocarbon ages and tentative tephra (volcanic ash) correlations based on microprobe analyses of glass. Two samples for $^{14}C$ dating were taken from core 6 that straddled a well-defined sediment pulse seen in nine of the eleven cores along the transect. It was hypothesised that the dates from this core would indicate the frequency of occurrence of sediment

<table>
<thead>
<tr>
<th>Parent material</th>
<th>$C$ [kPa]</th>
<th>$\phi$ [°]</th>
<th>$\rho_s$ [g cm$^{-3}$]</th>
<th>$T$ [m$^2$ day$^{-1}$]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Piha*</td>
<td>5.976 ± 1.946</td>
<td>38.8 ± 1.6</td>
<td>1.447 ± 0.7</td>
<td>18</td>
</tr>
<tr>
<td>Lone Kauri**</td>
<td>12.223 ± 2.157</td>
<td>39.4 ± 1.8</td>
<td>1.455 ± 0.7</td>
<td>15</td>
</tr>
</tbody>
</table>

** Submarine andesitic volcaniclastic sediments.

* Terrestrial andesitic lava flows.
pulses with magnitudes calculable from the stratigraphy. Two other $^{14}$C samples were taken from cores 18 and 22 at ~4 m depth, where a thin tephra layer was present. Because this part of the core is situated within the sand phase of the record (i.e. before the lake was completely dammed and sedimentation started), and around the transition from coarse to fine sand, these dates would enable a maximum age to be assigned to the initial impounding of the wetland.

Tephra layers when correlated and dated help provide a chronology for sedimentary records (Newnham and Lowe, 1991; Eden and Froggatt, 1996; Lowe et al., 1999; Shane and Hoverd, 2002; Sandiford et al., 2002; Newnham et al., 2004). In New Zealand, the recently active rhyolitic Taupo and Okataina caldera volcanoes, within the central Taupo Volcanic Zone (TVZ), are the two most frequently active rhyolite centres on Earth (Shane, 2000). Positive correlations commonly require multiple criteria (Froggatt and Lowe, 1990). In proximal settings (<50 km from vent), tephra beds can usually be identified from their stratigraphic position, lithology, and ferromagnesian mineral assemblages. In more distal settings, however, these features become less diagnostic and geochemical fingerprinting must be employed (Lowe, 1988; Shane, 2000). Two thin rhyolitic tephra deposits were identified in Te Henga Wetland, probably reworked in the wetland sediment record rather than in their primary form (e.g. see Moore, 1991). Major element compositions of glass shards of the two tephra deposits were analysed by electron microprobe. The efficacy of this technique to fingerprint tephra deposits was established by Froggatt (1983) and has been widely used (e.g., Lowe, 1988; Shane, 2000). A number of shards per sample are analysed and populations of identical composition are expressed as a mean and standard deviation. Glass compositions can then be compared with those from known (and dated) tephra deposits elsewhere.

3.2.2. Foraminiferal analysis

The occurrence of fossil benthic foraminifera has been documented in many marine and brackish environments around the New Zealand coast (Hayward et al., 1999). Interpretation depends on knowledge of their present-day ecological distribution in sheltered harbours and tidal inlets in northern New Zealand. These studies have shown that tidal elevation and salinity are the major environmental factors influen-
cing benthic foraminiferal distribution in these settings (e.g. Hayward et al., 1999, 2004). Sand layers in core 18 contained shells and benthic foraminifera which could indicate changes in marine or tidal influence and possibly give insight to the sedimentation history of the wetland. For the analysis, an approximate volume of 5–10 cm³ sediment per sample was taken. The mud fraction (<63 μm) of the sediment samples was washed out and the foraminifera were concentrated by floating on heavy liquid for searching with a microscope.

4. Results and discussion

4.1. LAPSUS-LS and sediment yield

4.1.1. Sensitivity analysis

After calculation of relative landslide hazard for the catchment (Eq. (1)), four parameters remain essential in constructing model scenarios for the subsequent soil redistribution and sediment yield. A sensitivity analysis is shown in Fig. 3, a plot of the changes in the model caused by varying one parameter but keeping others constant (default value) (Table 3). The slope limit for landslide erosion $\alpha$ determines where the erosional phase halts and the deposition begins. If this slope limit is raised, less total landslide erosion (and deposition) occurs and, as a consequence, the delivery ratio and sediment yield are lowered as well. The runout fraction $\varphi$ determines the total reach of the depositional phase. It has no influence on the total amount of erosion but increases the delivery ratio and sediment yield when raised because more material reaches the stream network. By increasing the threshold contributing area for determining a sediment transporting stream, the stream network becomes less dense and a lower sediment yield and delivery ratio are obtained. The model is very sensitive in the lower range of threshold contributing area values; the stream network becomes so dense that almost all landslide material is intercepted and the delivery ratio tends towards 1.0. The critical rainfall threshold ($Q_{cr}$) represents the landslide scenario and all grid cells with a value equal to or lower than the threshold fail and induce erosion and sedimentation. Much higher amounts of erosion are obtained when the critical rainfall threshold is raised because more grid cells, with a progressively lower landslide hazard, fail and cause soil redistribution.

4.1.2. Landslide scenarios and sediment yield

The sensitivity analysis (Fig. 3) shows that the slope limit for landslide erosion and the runout fraction have relatively small influences on modelling results for delivery ratio and sediment yield. Concerning the threshold contributing area for determining the stream network, it is an issue to decide which is the most appropriate minimal contributing area to represent a stream, capable of transporting sediment, or whether some other attribute such as slope should be part of the threshold (Tarboton et al., 1992; Montgomery and Foufoula-Georgiou, 1993). The choice of the threshold value is important in approximating the actual shape of the stream network and in obtaining accurate stream flow hydrographs as well. An arbitrary threshold value is usually chosen on the basis of visual similarity between the extracted network and topographic maps. However, in many cases this method poorly represents channel networks observed in the field because first-order channels and many second- and third-order channels may not be determined. Tarboton et al. (1992) suggested selecting the appropriate contributing area threshold for determining the channel network from an inflection in the drainage area–slope relation for averaged data. Montgomery and Foufoula-Georgiou (1993), however, discussed conceptual and procedural problems with this approach. Because sediment is transported typically by higher-order streams, a very accurate extraction of all lower-order streams is not required for our application. A threshold contributing area of 400 grid cells (0.25 km²) shows a good visual similarity with streams indicated on the topographic map, which are streams with a minimum length of 500 m (Land Information New Zealand, 2000). Furthermore, the modelling results for delivery ratio and sediment yield are relatively insensitive in this range of contributing area thresholds (300–500 grid cells, see Fig. 3).

The critical rainfall threshold largely defines the landslide scenario and strongly influences the modelling outcomes for delivery ratio and sediment yield. Table 3 illustrates three examples of scenarios in which the critical rainfall threshold and the threshold contributing area are both varied. Slope limit and runout
Fig. 3. Sensitivity analysis for sediment yield related parameters in LAPSUS-LS.
constant were kept fixed at 10° and 0.4°, respectively. Defining a denser channel network in scenario 1 resulted in a higher sediment yield and delivery ratio than for the default scenario. By raising the critical rainfall threshold in scenario 2, landslide erosion and sediment yield increased but there was no change in delivery ratio. The spatial patterns and interactions between landslide processes and channel network delineation for the three scenarios are shown in Fig. 4.

4.2. Sediment record interpretation

4.2.1. Core stratigraphy and chronology

The stratigraphy of the eleven sediment cores is shown in Fig. 5. Depths are given relative to the level of the causeway (which is less than 1 m above the mean wetland water level and 3–4 m above mean sea level). Below ~3 m depth, the cores consist of Holocene sand. A transition from coarse to finer sand occurs from ~3.5 m upwards. An irregular but clear boundary (varying between 300 and 266 cm) marks the start of sedimentation in the wetland, i.e. after it had become completely dammed by a landward-prograding dune system. The wetland acts as a highly efficient sediment trap, especially in terms of episodic or event-based input fluxes. The wetland is a very shallow lake, densely populated with vegetation (mostly reed), except for the main channels, which are the deeper parts, draining the wetland. In the vegetated ‘basins’, water flow is seriously reduced, suspended clay particles can easily settle and consequently the thickest clay layers are evident. These basins receive only water-containing sediments when significant extra water enters the wetland, typically during high-magnitude/low-frequency events triggering landsliding in the upstream catchment. In this way, the system works in a similar fashion as occurs, for example, in the marshes in the Wolga delta (Overeem et al., 2003). Most of the cores exhibited four well-defined grey homogeneous clay layers, interspersed with layers of coarse sand.
with peat and/or organic-rich mud, enabling correlation between cores to be made on the basis of these visible lithological changes. The sediment pulses are interpreted to represent high-magnitude landslide erosion events being preserved as overbank deposits of the main channels draining the wetland. Cores 18 and 21 lacked the sediment pulses, this lack being attributed to the core positions within the main channel draining the wetland where sediment is not preserved. Two fine tephra layers were identified in the sediment cores, both containing rounded pumice gravels/lapilli (3–8 mm in diameter) and so most likely have been reworked. Reworked, stranded tephra deposits are widely reported at coastal sites in New Zealand, especially along the North Island’s east coast, and are generally attributed to sea-rafting processes (e.g. Lowe and de Lange, 2000). A very thin tephra layer was present in eight of the eleven cores around ~3 m, in the upper part of the fine sand phase. A second tephra layer was identified in five cores at a depth of ~4 m, around the transition from coarse to fine sand (Fig. 5).

Four stratigraphic positions in cores 6, 22 and 18 were radiocarbon dated (Table 4). Electron microprobe analyses of the two tephra beds (T4 and T22), and tentative correlatives, are presented in Table 5. The glass shards are rhyolitic, have high FeO and CaO contents and thus are compositionally closely matched with glass from Holocene eruptives of Taupo caldera volcano (e.g. Stokes et al., 1992; Lowe et al., 1999; Shane, 2000). The radiocarbon samples from cores 18 and 22 were taken 15 and 19 cm above the deepest tephra layer (T22), respectively. The radiocarbon ages, stratigraphy and probable reworking in the wetland suggest that tephra T22 probably correlates with Taupo-derived tephras ranging from 6000–10,000 cal
4.2.2. Foraminiferal analysis

The samples from core 18 provided clear evidence that at least a 1-m interval (320–435 cm) accumulated in a sheltered estuarine environment (Tables 6 and 7). The upper two foraminiferal samples in this interval (320, 385 cm) contain rare agglutinated foraminifera typical of high tidal, low salinity salt marsh. The shells from the interval 412–435 cm and the lowest foraminiferal sample (431–435 cm) from the same interval comprise faunas that live preferentially in unvegetated, intertidal (low-mid tide) mud or sand flats in sheltered inlets and harbour edge settings with near normal or slightly reduced salinities. The abundance of *Arthritica bifurca* suggests that a lower tidal elevation was more likely. The specimen of the small, narrow limpet *Notoacmea helmsi scapha* provides evidence for the presence of *Zostera* seagrass because this limpet is adapted to living on its narrow blades. The single specimen of the foraminifer *Zeaflorilus parri* lives only in shallow subtidal exposed environments and must have been washed into the estuary from the open coast. Core 18 contains fossil evidence for the former presence of a sheltered estuary where Te Henga wetland now exists. The interval shallows, presumably with sediment accumulation from low tidal, moderately high salinity, sand flats up to a high tidal lower salinity salt marsh. This transition occurred ~6000 cal yr BP, estimated according to the stratigraphy, the ${}^{14}$C dates and occurrence of (probable) Motutere Tephra (Table 5). Subsequent compaction may account for some

Table 4

<table>
<thead>
<tr>
<th>Core number + depth (cm)</th>
<th>Laboratory number&lt;sup&gt;a&lt;/sup&gt;</th>
<th>Conventional age&lt;sup&gt;b&lt;/sup&gt; (14C yr)</th>
<th>Calibrated age range (yr BP) + probability (%)&lt;sup&gt;c&lt;/sup&gt;</th>
<th>$\delta^{13}$C (%)&lt;sup&gt;d&lt;/sup&gt;</th>
<th>Material</th>
</tr>
</thead>
<tbody>
<tr>
<td>18 (400)</td>
<td>Wk15043 NZA20362</td>
<td>5844 ± 40</td>
<td>6730–6490 (95.4)</td>
<td>–26.8</td>
<td>Charcoal</td>
</tr>
<tr>
<td>22 (420)</td>
<td>Wk15044 NZA20363</td>
<td>5331 ± 44</td>
<td>6200–5940 (95.4)</td>
<td>–24.5</td>
<td>Charcoal/Wood</td>
</tr>
<tr>
<td>6 (231)</td>
<td>Wk15045 NZA20221</td>
<td>196 ± 37</td>
<td>300–60 (79.0)</td>
<td>–28.7</td>
<td>Peat</td>
</tr>
<tr>
<td>6 (242)</td>
<td>Wk15046 NZA20222</td>
<td>474 ± 40</td>
<td>550–430 (91.6)</td>
<td>–29.2</td>
<td>Peat</td>
</tr>
</tbody>
</table>

* All ages determined by AMS (Accelerator Mass Spectrometry).
  * Wk refers to Waikato University Radiocarbon Dating Laboratory; NZA refers to the Rafter Radiocarbon Laboratory (Institute of Geological and Nuclear Sciences, Lower Hutt, New Zealand).
  * Ages in conventional radiocarbon years BP (Before Present where ‘present’ is AD1950 ± 1 Standard Deviation. Ages based on Libby half-life of 5568 for 14C (Stuiver and Polach, 1977), with correction for isotopic fractionation ($\delta^{13}$C) applied.
  * All ages were calibrated using the OxCal calibration program applying the IntCal98 calibration curve (Stuiver et al., 1998; Bronk Ramsey, 2001).
  * Parts per thousand difference (per mille) between the sample carbon 13 content and the content of the international PDB standard carbonate (Aitken, 1990); PDB refers to the Cretaceous belemnite formation at Peedee in South Carolina, USA.
of the difference between the thickness of the sequence and the indicated shallowing of ~2.5–3 m (with respect to the tidal range). The transition also coincides with the start of the ~2 m sea-level fall since ~7000 cal yr BP (Gibb, 1986).

4.2.3. Sediment pulses and corresponding landslide scenarios

Nine out of the eleven sediment cores, on both sides of the main channel (cores 18 and 21), exhibited four well-defined clay sediment pulses (Fig. 5). Sediment...
ment thickness and estimated sediment volumes can be used as a surrogate for the amount (magnitude) of erosion in the catchment. Together with some age control, frequencies of occurrence of the sediment producing landslide events can be established. Several sources of error are often involved in sedimentation surveys and especially in the calculation of sediment volumes (Butcher et al., 1993). Various authors have stressed the importance of the variation of the bulk density of sediments both between and within reservoirs and changes of volumes with compaction over time (Rausch and Heinemann, 1984; Pizzuto and Schwendt, 1997). Other researchers, however, concluded that the bulk densities of deposits from storm sediment pulses in cores showed little variation both between and within cores (Page et al., 1994b). Furthermore, others have argued that the use of volumes is less error-prone than transformations to mass (e.g. Butcher et al., 1993). No corrections were made for bulk density differences between eroded/transported soil and resulting core sediment layers or between sediment layers within the cores. A more accurate measure of sediment yield would also require more cores to be taken because of the variable nature of the wetland and consequently the spatial variability of sedimentation. Volumes were calculated by multiplying the mean sediment thickness (± one standard deviation) for each sediment pulse within the wetland depositional area (1.7 km²). The range of sediment volumes of the four pulses can then be linked to the sediment yield of a corresponding landslide scenario (critical rainfall threshold range (Table 8); sites with a value equal to or lower than this threshold are triggered and enter the soil redistribution and sediment yield algorithms of the model. It should be noted that these magnitudes were probably underestimated because the wetland has not complete trap efficiency. Furthermore, trap efficiency depends on the change in reservoir level and capacity, and high rates of sedimentation will cause it to vary over time, usually decreasing as the reservoir continues to infill (Butcher et al., 1993). Subsequent compaction of sediment, not only by its own weight but also intensified by construction of the overlying causeway, may account for another underestimation of sediment volumes (although most of the compaction probably involved the readily compressable peat or organic mud layers). Holocene compaction ratios of 0.2–0.5 have been recorded for estuarine organic-rich sediments (Pizzuto and Schwendt, 1997). Because of these assumptions, emphasis should be placed on the order of magnitude of the volumetric estimates rather than the precise values.

As also noted by other researchers who used a combination of a steady state hydrologic model and a deterministic infinite slope stability model to calculate landslide hazards (Eq. (1)), the values of $Q_{cr}$ can

<table>
<thead>
<tr>
<th>Table 6</th>
<th>Shell samples from core 18</th>
</tr>
</thead>
<tbody>
<tr>
<td>Depth (cm)</td>
<td>Description</td>
</tr>
<tr>
<td>412</td>
<td>Fragments of unidentifiable bivalve</td>
</tr>
<tr>
<td>427</td>
<td>Double valued (in-situ) cockle, <em>Austrovenus stutchburyi</em></td>
</tr>
<tr>
<td>435</td>
<td>Double valued (in-situ) small cockle, <em>Austrovenus stutchburyi</em></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Table 7</th>
<th>Raw census counts from three foraminifera-bearing samples from core 18</th>
</tr>
</thead>
<tbody>
<tr>
<td>Description</td>
<td>Depth (cm) 320–330 385–400 431–435</td>
</tr>
<tr>
<td>Foraminifera</td>
<td></td>
</tr>
<tr>
<td><em>Haplophragmoides wilberti</em></td>
<td>1 6 0</td>
</tr>
<tr>
<td><em>Jadammina macrescens</em></td>
<td>1 2 0</td>
</tr>
<tr>
<td><em>Miliammina fusca</em></td>
<td>4 0 0</td>
</tr>
<tr>
<td><em>Trochamminita salsa</em></td>
<td>0 2 0</td>
</tr>
<tr>
<td><em>Ammonia aoteana</em></td>
<td>0 0 17</td>
</tr>
<tr>
<td><em>Zeollorilus parri</em></td>
<td>0 0 1</td>
</tr>
<tr>
<td>Diatoms</td>
<td>rare 0</td>
</tr>
<tr>
<td>Ostracods</td>
<td>0 0 7</td>
</tr>
<tr>
<td>Echinoderm spines</td>
<td>0 0 2</td>
</tr>
<tr>
<td>Barnacle plates:</td>
<td></td>
</tr>
<tr>
<td><em>Austrominius australis</em></td>
<td>0 0 1</td>
</tr>
<tr>
<td>Molluscs shells:</td>
<td></td>
</tr>
<tr>
<td><em>Arthritica bifurca</em></td>
<td>0 0 20</td>
</tr>
<tr>
<td><em>Austrovenus stutchburyi</em></td>
<td>0 0 7</td>
</tr>
<tr>
<td><em>Notoacmea helmsi</em></td>
<td>0 0 1</td>
</tr>
<tr>
<td><em>Notoacmea helmsi scapha</em></td>
<td>0 0 1</td>
</tr>
</tbody>
</table>
only be interpreted as a relative measure of the potential for shallow landslide initiation (Montgomery and Dietrich, 1994; Borga et al., 2002). A real translation of critical rainfall values to real rainfall data is very difficult to make because the steady state hydrologic model requires the assumption that the predicted spatial pattern of critical steady state rainfall represents that which occurs during an unsteady, landslide producing rainfall event.

According to the ages determined in core 6, at least two high-magnitude events (sediment pulses 3 and 4, Figs. 5 and 6) occurred in pre-European times. These may have been caused by natural landslide activity or the influence of early Polynesian (Maori) settlement around the margins of the wetland (Diamond and Hayward, 1979; Hayward and Diamond, 1978), or both. Similar mud layers have been found in some Waikato lakes and analyses of associated pollen and δ13C values have shown the layers coincide with catchment deforestation of an unprecedented scale and thus attributable to Polynesian burning at around 700 cal yr BP (e.g., Hogg et al., 1987; Green and Lowe, 1994; see also Hogg et al., 2003). The two younger events occurred over about the last 150 years and may have been caused or at least intensified by logging and quarry operations upstream from the late 1830s to 1940s (Diamond and Hayward, 1980). A more precise timing of the influence of the first European (after 1830) forest clearance operations could be better distinguished from natural impacts by pollen analysis, which can show the introduction of exotic European species (e.g. Wilmshurst et al., 1999). It should be noted that landslide hazards are calculated with parameter settings for the present, forested study

<table>
<thead>
<tr>
<th>Layer thickness (cm) (± 1 SD)</th>
<th>Sediment volume (m³)</th>
<th>Minimum $Q_{cr}$ (m day$^{-1}$)</th>
<th>Mean $Q_{cr}$ (m day$^{-1}$)</th>
<th>Maximum $Q_{cr}$ (m day$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.94 (± 1.81)</td>
<td>66,980 (± 30,770)</td>
<td>0.0100</td>
<td>0.0248</td>
<td>0.0392</td>
</tr>
<tr>
<td>5.72 (± 1.62)</td>
<td>97,240 (± 27,540)</td>
<td>0.0260</td>
<td>0.0390</td>
<td>0.0518</td>
</tr>
<tr>
<td>10.67 (± 4.77)</td>
<td>181,390 (± 81,090)</td>
<td>0.0404</td>
<td>0.0783</td>
<td>0.1189</td>
</tr>
<tr>
<td>12.33 (± 4.00)</td>
<td>209,610 (± 68,000)</td>
<td>0.0596</td>
<td>0.0919</td>
<td>0.1271</td>
</tr>
</tbody>
</table>

Table 8
Volumes of the four sediment pulses, from top downward, and their corresponding range of landslide scenarios

Fig. 6. Back-analysis of calculated wetland sediment volumes to corresponding LAPUS-LS landslide scenarios.
area according to Tables 1 and 2. In the parts of the area where logging took place, lower root reinforcement (lower \( C_r \) values) would result in higher landslide hazards and relatively more sediment yield implying a small overestimation of the critical rainfall thresholds representing the last two landslide scenarios. Moreover, logging would cause different hydrological conditions (changes in transmissivity, reduction in interception of rainfall). However, regarding the relatively simple level of modelling envisaged here, interception is not included in the model and transmissivity is treated as a soil intrinsic parameter (i.e., without influence of vegetation, Table 1).

5. Conclusion

In this paper we have assessed the possibility of combining a wetland sediment record with the LAP-SUS-LS landslide model to reconstruct the sedimentation history of the wetland and the occurrence of high-magnitude/low-frequency landslide events in the catchment upstream. By using radiocarbon dating, tephrochronology and foraminiferal analysis, we established the stratigraphy and chronology for eleven sediment cores. A small drop in sea-level following the Holocene sea-level maximum is represented in the lower part of the sediment record, dated at c. 6000 cal yr BP, and marked by a transition from coarse to finer sand and a change in foraminiferal content. The actual damming of the wetland by a landward prograding dune system inducing the start of freshwater sedimentation was completed by c. 1000 cal yr BP. At least four clay sediment pulses are recognised in the cores and interpreted as representing high-magnitude landslide events. The two oldest events occurred are either natural phenomena or the result of early Maori settlement, or both, whereas the last two events are most likely caused or intensified by forest clearance and logging activities in the upland catchment from the 1830s to the 1940s. Sediment volumes were calculated from the cores and corresponding landslide scenarios were defined through back-analysis of LAPSUS-LS sediment yield model output. Although initially not intended to quantify landslide erosion and sediment yield, the model seems capable of linking a catchment scale calculated sediment yield, resulting from a landslide scenario and expressed by a threshold critical rainfall, with late Holocene sediment pulses preserved in the wetland at the basin outlet.

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