Process-Based Models for Simulation and Prediction of Erosion-Related Organic Carbon Losses on the National Scale: an Example for New Zealand

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Abstract: The system of integrated process-based models was used to predict the rates and interconnection of erosion processes at the national scale as part of an Erosion-Carbon programme established to account for New Zealand's obligations to decrease greenhouse gas emission in terms of the Kyoto Protocol.

1 Introduction

The islands of New Zealand, covering 265,846 km^2 (0.18% of the global landmass), discharge (240—370) 10^6 t • a^{-1} of sediment to the ocean (~1.7%—2.6% of the global amount). The main reasons for this a disproportionately large sediment production are, firstly, New Zealand's position on the active tectonic boundary between the Australian and Pacific plates, which causes relatively high relief and weak lithology. Second, New Zealand's position relative to the prevailing westerly winds and maritime location, which induces high annual precipitation of up to 10,000 mm—12,000 mm in the South Island's Southern Alps. Third, high rates of natural erosion have already been significantly accelerated by human deforestation of about 50% of the landmass and the establishment of pastures on shallower and weaker soils.

As part of the overall effort to address the issue of greenhouse gas emissions in terms of the Kyoto Protocol, an estimate was recently made of the contribution that the mobilisation and redistribution of soil organic carbon by erosion make to New Zealand's annual carbon budget (Tate *et al*., 2001). That estimate was rather broad (3 Mt/a—11 Mt/a), and a two-year project was initiated in 2001 by Landcare Research (see Trustrum *et al*., this volume). The aim was to reduce the large uncertainty in the contribution of erosion to New Zealand's terrestrial C budget (Tate *et al*., 2000) to less than 50%, as part of a project to prepare New Zealand for possible full C accounting beyond the first commitment period of the Kyoto Protocol. This is to be achieved through the application of process based models representing erosion, transport and deposition of both sediments and associated particulate organic carbon). Thus, elements of the carbon cycle are to be linked to a classic "source to sink" terrestrial sediment budget for the whole of the New Zealand mainland. Although the initial focus is on the behaviour of carbon, the methodology might equally be applied to other particulate matter or nutrients that are transported within regional sediment fluxes.

In this paper, the overall approach being taken to the modelling of the erosion-carbon budget at regional/national scales is described, and a brief example is given of the individual process models.

2 The erosion-carbon budget

The rate of change of soil organic carbon volume for a given spatial unit (e.g., pedon) can be written in a simplified form as:

$$\frac{d(SOC*d_s)}{dt}=I_{SOC}-f_{SOC}*SOC-a_0\left(E_{SOC}-D_{POC}\right) \qquad (1)$$

where SOC is the content of organic carbon within a soil pedon; I_{SOC} is the rate of organic carbon input due to soil formation processes; $f_{SOC}*SOC$ is the rate of CO_2 emission to the atmosphere and the rate of SOC solution in ground water, which is generally linear to SOC; E_{SOC} is the rate of removal of organic carbon from the soil due to mechanical erosion; D_{POC} is the rate of addition of organic carbon to the soil due to mechanical deposition of sediments; t (s) is time; and d_s (m) is soil depth.

Similarly, the amount of particulate organic carbon in transported sediment can be written in simplified form as:

$$\frac{\partial(POC * q_s)}{\partial X} = I_{POC} - f_{POC} * POC - a_0 \left(E_{SOC} - D_{POC} \right) \tag{2}$$

where POC is the content of particulate organic carbon in transported sediments; I_{POC} is the rate of organic carbon input to the suspended sediment load; $f_{POC}*POC$ is the rate of CO_2 emission to the atmosphere and the rate of POC solution in the water, which is generally linear to POC; X (m) is the distance along the flow; and q_s is the specific volumetric sediment discharge (m^2/s).

For any point along a regional flow network, a simple sediment budget can be written as:

$$\frac{\partial q_s}{\partial X} = a_0 \left(E - D \right) \tag{3}$$

where E is the rate of erosion and D is the rate of deposition. All terms in equations (1)—(3) are in units of m/s. The coefficient a_0 reflects the spatial heterogeneity of erosion and is calculated as the ratio of the actual area affected by erosion to the whole contributing catchment area.

A *sediment* budget (i.e. Equation (3)) integrates the erosion and deposition attributable to each individual erosion process. Closing the *erosion-carbon* budget requires solution of Equations (1) and (2), and thus the integration of the contribution of each of these processes to the erosion-carbon flux. For each process, therefore, estimates of E_{SOC} and D_{POC} are required. Although E_{SOC} and D_{POC} will be proportional to E and D respectively, the nature of that proportionality will vary between processes. For example, on hill slopes affected by sheet and rill erosion, the organic carbon flux due to erosion (E_{SOC}) is a function of erosion/deposition rates and of organic carbon content in eroded soil and deposited sediments.

$$E_{SOC} = \frac{d_s * SOC + D * POC}{d_s + D} E \tag{4}$$

For other types of erosion processes, these relationships are more complicated. The most difficult for analysis is a large river channel with low SOC in the bed load and several sources of POC (river bank erosion, sediment input from the catchment).

3 Model design and structure

Modelling of the erosion-carbon flux is to be undertaken at the national scale, using the NZ Digital Elevation Model, with 25 m to 25 m pixel. The model run will cover a 25-year period, thus approximating the erosion-carbon flux occurring between 1990 and the Kyoto target date. Meteorological and hydrological data were obtained from ~600 stations with more than 20 years of measurements. The hydraulic characteristics of flow in both ephemeral and permanent streams of different roughness were estimated by field experiment. Soil and rock parameters (or their surrogates) were taken from the New Zealand National Soil Database (McDonald *et al.*, 1988) (NSD) and from the New Zealand Land Resource Inventory (NZLRI). Organic carbon content in topsoil and rocks was obtained from the NSD and from the national Soil Carbon Monitoring System (Scott *et al.*, 2001). Vegetation cover parameters were estimated from the Land Cover Database and from the Land Resource Inventory (Newsome, 1987). The Land Resource Inventory was also used to define the erosion process types at the given area (Eyles, 1985).

A system of integrated process-based and statistical models was used to predict the rates and interconnection of selected erosion processes at the national scale. The national Land Resource Inventory (Eyles, 1985) shows that wind erosion is active over 12% of the country. More than 52% of the land is subjected to flow-induced processes: sheet and rill erosion, gully and tunnel gully erosion and stream bank erosion. The highly concentrated (including viscous) flows, e.g., debris avalanches, earthflows and mudflows, occur on about 15% of mapping units. Mass movement (landslides of different depth) destroys vegetation cover and contributes sediments to the streams and lower parts of the slopes on 40% of mapping units. These were considered to be the most important forms of erosion in terms of magnitude of contribution to the national sediment flux.

There is an important difference in the way in which individual processes contribute to sediment and carbon budgets. This lies in the distinction between flow-driven and mass movement processes. Flow-driven processes are more or less spatially continuous along flow lines and are temporally continuous at least for the duration of a process-forming event. Accordingly, they can be described by the Eulerian approach (Equations (1)—(3)). By contrast, mass movement is spatially discrete and is only triggered by the process-forming event but does not persist for the duration of that event (as will be discussed below, slow, continuous mass movements are not modelled here). The dynamics of these features and associated carbon budget can be better described using a Lagrangian approach. Representation of the mass movement process within the overall erosion-carbon budget is therefore indirect. As New Zealand landslides are quite rapid, they can be assumed to be instantaneous for the daily time step incorporated in this modelling. In this case, the effect of mass movement will be to occasionally change initial conditions (such as soil depth, carbon content, elevation, slope angle, vegetation cover properties, etc.) for the processes that can be represented in terms of Equations (1)—(3).

4 Individual processes

4.1 Mass movement

The most widespread type of mass movement in New Zealand is shallow translational regolith landsliding. The actual area of active or recent landslides changes over time due to temporal variation in the occurrence of rainfall, spatial variability of the soil and vegetation properties, and rates of soil rehabilitation on landslide scars. At regional scales, however, the area that may be affected by landsliding has been estimated within the New Zealand Land Resource Inventory as ca. 34,000 km^2 of the North Island and ca. 20,000 km^2 of the South Island.

The occurrence of shallow regolith landslides is modelled using a limiting equilibrium stability equation:

$$FS = \frac{c + (\gamma - m\gamma_w) z \cos^2 \beta \tan \varphi}{\gamma z \sin \beta \cos \beta} \tag{5}$$

Failure occurs in cases where the balance of strength and stresses – the Factor of safety (*FS*) – in a given regolith unit falls to unity. The principal factors of this static model are slope angle (β), weight density (γ) and depth (z) of regolith, moisture and strength of regolith characterised in terms of cohesion (c) and angle of internal friction (ϕ). Given that these are rainfall-triggered failures, the most dynamic factor determining failure is soil moisture. Variation in this is represented in simple form using m, the ratio of the depth of the perched water table to the total regolith depth, which also determines the value of γ (see e.g., Preston and Crozier, 1999). The reduction in strength resulting from soil water is estimated using this ratio and the density of water (γ_w). Slope angle is available for each 25 by 25 m pixel of the national DEM. Regolith density and depth are characterised using values for the main NZ soil groups from the National Soil Database. The same soil groups were used to stratify field and laboratory measurements of regolith cohesion and friction. A time series of daily values of m was calculated for each pixel, derived from the daily rainfall time series using an hydrological model based on the TOPMODEL approach (Woods et al., in prep.). This enables a simple estimation of volumes of sediment generated by shallow landslides. This sediment is redistributed in space using an algorithm based on the site geometry (principally slope angle) and characterisation of the failed mass, i.e. indices of density and viscosity (see Crozier, 1996 and Dymond *et al.*, 1999). An increment of deposition, in terms of both soil depth and *SOC*, is added to each pixel along the flow line. This model thus provides initial conditions to be used in models for other processes. These include: percentage bare ground, remaining soil depth, and *SOC* content.

Deep-seated rotational landslides occur in deeply weathered rocks and ashes, and are thus relatively limited in their distribution within New Zealand. Nevertheless, they are considered to influence the behaviour of other processes. Although, no attempt is made within the present project to model their occurrence using a process-based approach, information concerning their distribution is available from the NZLRI, and can be used to define initial conditions for use in other process models.

4.2 Viscous and hyperconcentrated flows

There are three main types of viscous flows: earthflows, mudflows, and debris avalanches. The behaviour of these processes is rather difficult to predict, as the parameters of the process-based models are difficult to estimate. Earthflows involve very slow viscous movement of the upper 2 m—3 m of saturated ground, typically with a width of 50 m—100 m. However, although a large volume of material may be in motion, the net sediment transport effect over a period of 20 years is quite low. Nevertheless, earthflows significantly disturb the vegetation cover and increase surface roughness, and thus influence initial conditions for the modelling of other processes. Therefore, information about the distribution of earthflows (i.e. from LRI) will be required. Mudflows are indicated in the Land Resource Inventory as occurring only rarely. Indeed, in the New Zealand context, they are analogous to the saturated mass of a shallow landslide failure, and can be represented as such.

The most active hyperconcentrated flow for New Zealand steeplands is the debris avalanche, indicated in the NZLRI as potentially occurring for 11% of the land area, mainly in the indigenous forest. All other erosion processes are of much lower intensity in these areas, and the sediment delivery ratios for debris avalanches are close to 100%. Accordingly, the rate of erosion attributable to debris avalanches can be estimated using the measured sediment discharges recorded at hydrological stations within indigenous forest catchments affected by this process.

4.3 Hillslope erosion

Sheet and rill erosion from bare surfaces, arable land, pasture and forested areas are modelled through calculation of the detachment and deposition of soil aggregates (Sidorchuk, 2001, see also details in Sidorchuk, this volume). This approach enables the calculation of redistribution of sediments and associated soil carbon on the hill slopes, and also the estimation of sediment yields to gullies and to the river network. The model uses a net of flowlines derived from a DEM using ArcInfo procedures. The model is three-dimensional, in that the multi-layered soil texture (including the top layer with the vegetation cover) can be used to evaluate *SOC* contribution from different soil horizons.

The model used to represent gully erosion is based on empirical observations of the evolution of gullies from key sites and on mathematical descriptions of the aggradation/degradation of gully beds and sidewalls (Sidorchuk, 1999), and shows the contribution of sediments and (partly) soil organic carbon to the river network. The overall rate of gully erosion is controlled by the erosion and deposition of sediments in the gully bed (E_{bed}), the erosion of the banks during gully flow (E_{bank}), and landsliding from the gully sidewalls. Analysis of experimental results (Sidorchuk, 1999) shows that, for areas with steep slopes, which are common for gullies, the rate of soil erosion in the gully bed (E_{bed}) is linearly correlated with the product of bed shear stress ($g\rho_w dS$) and mean flow velocity (U):

$$E_{bed} = k_e U g \rho_w dS \tag{6}$$

where g is acceleration due to gravity, ρ_w is density of water, d is depth of flow, and S is gully bed inclination. Gully bank erosion (E_{bank}), i.e. increase in the width of the active gully bottom (W_b), is proportional to the rate of bed erosion (E_{bed}) transformed with the ratio of lateral (U_2) to longitudinal velocities (U):

$$\frac{dW_b}{dt} = k_b \frac{U_2}{U} E_{bed} \tag{7}$$

where k_b is a constant that varies with the ratio between the erosivity of bed and bank material. If $W_b < 10W$, $\frac{U_2}{U} = 0.22d/W$, if $W_b > 10W$, $\frac{U_2}{U} = 2.2d/W_b$, and if $W_b > 20W$, $\frac{U_2}{U} = 0$ depth (d) and velocity (*U*) can be estimated using the Chezy formula. The width (*W*) of the flow in gullies can be calculated using discharge (*Q*) with the empirical formula:

$$W = 0.3 * Q^{0.4} \tag{8}$$

The rate of deposition in gullies is generally low and, in the case of fine particles and high turbulence (which is usual for gully flow), can be non-existent.

Gully erosion is a typical example of the combination of flow-driven and mass movement processes. In the model, these processes are linked using a procedure of sequential calculation. First, the depth of gully incision D_v is calculated using Equation (6). A critical depth of incision for bank stability (D_{crit}) is calculated as:

$$D_{crit} = \frac{2c}{\gamma} \cos\varphi \Big/ \sin^2 \frac{1}{2}\left(\varphi + \frac{\pi}{2}\right) \tag{9}$$

If the depth of incision is greater than the critical depth, a landsliding algorithm must be applied. A straight stable slope model is used to predict gully sidewall slope (ϕ):

$$\frac{c}{\gamma D_v} = \frac{\gamma - w\gamma}{\gamma} \tan\varphi * \cos^2\phi - \frac{\sin(2\phi)}{2} \tag{10}$$

where *w* is volumetric water content in the soil. With known bottom width, sidewall slope and volume of incision V_0, the shape of the gully cross-section can be transformed into a trapezium with bottom width W_b, depth $D_t = \left(\sqrt{W_b^2 + \frac{4V_0}{\tan\phi}} - W_b\right)\frac{\tan\phi}{2}$ and top width $W_t = W_b + 2D_t / [\tan\phi]$

4.4 Sediment erosion and deposition in the river network

The model used to represent river channel bed aggradation/degradation is based on fluvial dynamics equations (Sidorchuk, 1996). This model represents the storage and re-entrainment of sediments and (partly) soil organic carbon in and from the river channel. The equation of mass conservation can be solved for the river network with the following assumptions:

(1) The lateral specific discharge q_w is constant for a river network segment with length *L*, and water discharge *Q* in the channel increasing linearly from an initial value Q_o with distance *X*:

$$Q = Q_0 + q_w X \tag{11}$$

(2) The upward sediment flux can be described with Equation (6).

(3) The rate of the bank erosion can be calculated from the empirical formula:

$$\frac{dW_b}{dt} = k_{bank} Q^m S^n \tag{12}$$

using the coefficients calibrated for New Zealand conditions through empirical observation of key river sites.

(4) The rate of deposition is expressed by the formula:

$$D = \delta V_f \tag{13}$$

where V_f is the fall velocity of sediment particles in turbulent flow. When the flow depth *d* is greater than floodplain height z_{bf}, i.e. when the floodplain is itself submerged, $\delta = \frac{d - z_{bf}}{d}$. When flow is confined within the channel banks, $\delta = 1$.

(5) Channel width and depth, bank height, sediment particle size and sediment concentration in the lateral input are constant for a channel reach with length *L*.

4.5 Wind erosion

Process-based modelling of wind erosion has not yet been attempted in New Zealand. Direct and indirect quantitative measurements of contemporary soil loss from wind erosion for the Canterbury plains and downlands, the Manawatu plains, and the Canterbury and Otago mountains are available and can be used for the first order estimation of wind erosion rates (Basher and Painter, 1997).

Background wind transport rates have been directly measured on the Canterbury plains and in the mountains using mast-mounted traps. Painter (1978) suggests that continual soil movement throughout the year on the plains at background rates results in soil losses of the order of 0.1 t/(ha • yr), while Butterfield's (1971) data for the mountains suggest background rates of 1.7 t/(ha • yr). Storm losses are typically of the order of 20 t/ha—125 t/ha (Hunter and Lynn, 1988). Medium-term wind erosion rate estimates, derived from Caesium—137 distribution, have been made on both cropland and rangeland. Rates range from 5 t/(ha • yr)—15 t/(ha • yr) from cropland on the Manawatu and Canterbury plains (Basher and Painter, 1997), to 10 t/(ha • yr) in severely degraded rangeland under a semi-arid climate (Hewitt, 1996), to 40 t/(ha • yr) in severely degraded rangeland under a humid climate (Basher and Webb, 1997). On cropland in the South Canterbury downlands there was a high rate of soil redistribution but no net loss from wind erosion (Basher *et al.*, 1995).

4.6 Model calibration

Calibration is performed on the basis of ~100 points with available suspended sediment yield estimates (Hicks *et al.*, 1996). This calibration optimises the mean square difference between calculated and measured sediment yields. The individual erosion/deposition models are verified and calibrated for key sites at which different erosion processes and their controlling factors have been measured.

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