The PESERA coarse scale erosion model for Europe: I – Model rationale and implementation

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Abstract

The principles and theoretical background are presented for a new model (PESERA) that is designed to estimate long term average erosion rates at 1 km resolution for most of Europe. The model is built around a partition of precipitation into components for overland flow (infiltration excess, saturation excess and snowmelt), evapotranspiration and changes in soil moisture storage. Transpiration is used to drive a generic plant growth model for biomass, constrained as necessary by land use decisions, primarily on a monthly time step. Leaf fall, with corrections for cropping, grazing etc, also drives a simple model for soil organic matter. The runoff threshold for infiltration excess overland flow depends on vegetation cover, organic matter and soil properties, varying dynamically over the year, and drives overland flow using the distribution of daily rain amounts. Total erosion is driven by erodibility, derived from soil properties, powered overland flow discharge and gradient; and is assessed at the slope base to estimate total

loss from the land. The model is run, using monthly averages and distributions of daily precipitation, to equilibrium in order to estimate long term averages, and is being validated against the limited erosion data available. Data sources, uniformly available across Europe, include the European Soils Data base, CORINE land use, MARS 50 km interpolated climate data and 90m DEM (SRTM).

Introduction

Soil erosion has long been identified as an important global issue, with implications for the maintenance of fertile soil and crop yields (e.g. Bennett, 1939 for North America, Seely and Wohl, 2004 for semi-arid and arid areas worldwide). Early models, particularly the USLE and its derivatives (Wischmeier and Smith, 1958, 1978; Renard et al, 1991) focussed on a broad-scale approach which could be readily applied in a wide range of conditions to give advice of conservation practice. Much of the more recent work has, however, concentrated on detailed process models (eg. WEPP, Nearing et al 1989; EUROSEM, Morgan et al, 1994; KINEROS, Smith et al, 1995; LISEM, de Roo, 1996) which have progressively incorporated improving knowledge of processes for runoff generation and sediment transport, but which lack the ease of application to new sites poor in data.

Here we present a theoretical framework for developing a simplified process based model, in particular providing explicit dependence on climate and vegetation, and implement this model as a 1-km resolution model across most of Europe. This paper describes the innovative aspects and theoretical background to the model, and a companion paper will describe its performance, calibration scenario application at the European scale.

The PESERA model (Pan–European Soil Erosion Risk Assessment) is a physically based and spatial distributed model developed to quantify soil erosion of environmentally sensitive areas relevant to a regional or European scale and define soil conservation strategies. The current version of model was developed within the structure of the PESERA project (contract No QLK5-CT-1999-01323) funded by the European Commission, Research Directorates General, DG VI (Quality of Life and Management of Living Resources), and was also based on previous funded and un-funded research (Kirkby and Neale, 1987; de Ploey et al, 1991; Kirkby and Cox, 1995; Kirkby et al, 2000). The PESERA model combines the effect of topography, climate, vegetation cover and soil into a single integrated forecast of runoff and soil erosion

The importance of soil erosion at a regional scale

Erosion by running water has been identified as the most severe hazard threatening the protection of soil in Europe (EC, 2006). By removing the most fertile topsoil, erosion reduces soil productivity leading, where soils are shallow, to a progressive and ultimately irreversible loss of natural farmland, and in vulnerable areas, is one major process of desertification. Severe erosion is commonly associated with the development of temporary or permanently eroded channels or gullies which can fragment farmland. The soil and runoff removed from the land during a large storm accumulates below the eroded areas, spilling offsite and in severe cases blocking roadways or channels and inundating buildings. Erosion rate is very sensitive to both climate and land use, as well as to detailed conservation practice at farm level. In a period of rapid changes in both climate and land use, due to global change, revised agricultural policies and international markets, it is valuable to be able to assess the state of soil erosion at a European level, using an objective methodology which allows the assessment to be repeated as conditions; pressures and drivers, change, or to explore the broad scale implications of prospective global or Europe-wide changes. This provides an estimate of the overall costs attributable to erosion under present and changed conditions, and objectively suggests areas for more detailed study and possible remedial action.

The PESERA model provides such an objective estimate of current rates of soil erosion, averaged over a series of years with current climate and land use. European estimates have been made at a resolution of 1 kilometre, and indicate the rate of loss of material from hillsides. Sediment delivery through the river system is explicitly not taken into account, and most of the eroded material generally remains close to its source, with significant off-site effects generally confined to a local area.

Process model approach

There are a number of possible methodologies for creating a coarse scale erosion map (Gobin et al, 2004). Some of these are based on the collection of distributed field observations, others on an assessment of factors, and combinations of factors, which influence erosion rates, and others primarily on a modelling approach. All of these methods require calibration and validation, although the type of validation needed is different for each category. There are also differences in the extent to which the assessment methods identify past erosion of an already degraded soil resource, as opposed to risks of future erosion, under either present climate and land use, or under scenarios of global change. Here a physically-based process model is presented, within the limitations of resolution and available data.

Process models have the potential to respond explicitly and rationally to changes in climate or land use, and so have great promise for developing scenarios of change, and what-if analyses of policy or economic options. Set against this advantage, process models generally make no assessment of degradation up to the present time, and can only incorporate the impact of past erosion where this is recorded in other data, such as soils data bases. Models also generally simplify the set of processes operating, so that they may not be appropriate under particular local circumstances. Although the USLE and RUSLE have been the most widely applied models in Europe (e.g. van der Knijf *et al*, 2000) , this approach is now widely considered to be conceptually flawed, and other models are now emerging, based on runoff thresholds (e.g. Kirkby *at al*, 2000) or the MIR(Minimum Information Requirement) approach (Brazier *et al*, 2001) applied to the more complex USDA WEPP model (Nearing *et al*, 1989).

The application of a process model has been preferred here for three main reasons.

1. It applies the same objective criteria to all areas, and so can be applied throughout a region, subject to the availability of suitable generic data.
2. It provides a quantitative estimate of erosion rate which can be compared with long term averages for tolerable erosion.
3. The methodology can be re-applied with equal consistency as available data sources are improved, and for past and present scenarios of changed climate and land use.

Point hydrology and land cover

The model presented represents a fundamental advance on previous models of comparable simplicity, most notably the USLE and its derivatives, by explicitly separating hydrology from sediment transport. That is to say that it first estimates storm overland flow runoff, and then uses the runoff to estimate sediment transport. Soil properties therefore enter separately into these two stages, replacing the separation in USLE between erosivity as a purely climatic property and erodibility as a pure soil property.

At the same time, the PESERA model has been designed to provide an estimate of long term erosion and must therefore scale up from our knowledge of instantaneous sediment transport as a function of shear stress or flow power to firstly an aggregate relationship between event discharge and event sediment discharge, and secondly from single events to the aggregate of storm events across the relevant distribution of storms. This temporal scaling up provides the essential link between climate, defined by the distribution of rainfall events and long term sediment transport. Although this scaling up has been discussed and partially implemented in previous models (Kirkby Kirkby et al, 1996; Kirkby, 1998), it has not previously been applied within a soil erosion model.

Runoff in a single storm

Figure 1 outlines the hydrological balance within the PESERA model. Precipitation is divided into daily storm events, expressed as a frequency distribution, that drive infiltration overland flow and soil erosion, and monthly precipitation, some of which may be as snow, driving saturation levels in the soil. Infiltration excess overland flow runoff is estimated from storm rainfall and soil moisture. Sediment transport is then estimated from overland flow and routed, in principle, downslope. Alternative methods for making these estimated are discussed below. To obtain long term estimates of soil erosion these estimates must then be scaled up by integrating over time. This process of scaling up has two stages, first from momentary to event-integrated dependence, and secondly from events to long term averages via the frequency distribution. For the first stage, if instantaneous sediment discharge can be expressed as a power law dependence on instantaneous water discharge, the relationship between event total sediment discharge and event total discharge will, in general, also be a power law, but the exponent will differ according to how hydrograph form changes with flood volume. Table 1 indicates how different generalisations of storm profile influence the relationship between instantaneous and time-averaged exponents. Other possibilities exist if there are thresholds for movement and/or hysteretic sediment stores, but in general it is reasonable to assume a similar power law relationship between sediment transport and discharge for event totals as for instantaneous values, but with some modification to allow for systematic changes in hydrograph shape.

In the second stage of scaling up, individual storm totals are integrated over the frequency distribution of storms. Two assumptions are normally made, first that the distribution of storms can be replaced by the distribution of daily rainfalls, and second that overland flow can be estimated on the basis of monthly average soil moisture conditions. The first of these assumptions avoids the discussion of how rainfall is divided, more or less arbitrarily, into storm events. The use of a daily unit is both convenient, in that daily rainfall data is relatively widely available, and appropriate in the sense that bursts of rainfall within a single day are significantly influenced by raised soil moisture levels from previous bursts, whereas for longer periods there may be significant drying between bursts. Similarly monthly updating of soil moisture is sufficient to reflect important seasonal differences in weather, to respond to seasonal differences in land cover and to make use of widely available meteorological data. These assumptions are however a compromise, attempting to simplify the estimation of storm runoff while retaining the frequency signature of storms (daily) and soil moisture (monthly).

This approach can be applied using either a historic (or simulated historic) sequence of daily rainfalls, or by summing over a frequency distribution of daily rainfall events for each month. The former approach is preferable for comparison with observed data, whereas the latter is more suitable for estimating long term average rates, but has the disadvantage that it does not respond to inter-annual differences or to the timing of consecutive storms within a month. These methods thus provide an explicit link to available climate data, providing an improved physical basis for comparisons across large regions, and with climate scenarios or historic data.

There are a number of simple methods for estimating storm runoff from storm rainfall. Implicitly these are all based on an understanding of the infiltration process, and an understanding that erosive overland flow can generally be represented as an infiltration excess, or Hortonian, process. The effect of subsurface flow, where and when it is important, may then be used to modify potential rates of infiltration, with lower infiltration under wet conditions. Similarly the role of vegetation and soil organic matter can modify the infiltration rates through changes in soil structure and/or the development over time of surface or near-surface crusting. Three models are coupled to provide the dynamics of these responses; first an at-a-point hydrological balance which partitions precipitation between evapotranspiration, overland flow, subsurface flow and changes in soil moisture; second a vegetation growth model which budgets living biomass and organic matter subject to the constraints of land use and cultivation choices; and third a soil model which estimates the required hydrological parameters from moisture, vegetation and seasonal rainfall history.

At-a-point soil hydrology can be described through the Richards' equation, although with reservations where both matrix and macropore flow are active. Solutions may be approximated through the use of infiltration equations, such as the Green-Ampt (1911) or Philip (1957) formulations. However these approaches are not compatible with the use of daily time steps, within which the detail of storm profiles is lost, and it is impracticable to provide better estimates of runoff than those from the SCS curve number (Yuan et al, 2001) or a simple bucket model. Here the bucket model is prefered, which offers a simple conceptual insight into the volume of infiltration before runoff occurs, that can be linked directly to concepts of soil moisture storage, as it varies within and between sites. In the bucket model, runoff $\boldsymbol{r}$ is given by:

$$r = p(R - R_0) \qquad (1),$$

In which $\boldsymbol{R}$ is total storm rainfall, $\boldsymbol{R_0}$ is the runoff threshold, or bucket storage capacity, and $\boldsymbol{p}$ is the proportion of subsequent rainfall that runs off.

Figure 2 illustrates the typical large scatter in relationships between observed total rainfall and runoff, and none of these models can provide a satisfactory fit to the spread of data for daily time steps, and the bucket model (equation 1) has been adopted in the PESERA model, in which storms are treated as independent random events. Figure 3 shows the application of the Green-Ampt equation, with assumed parameters, to a set of storm events taken from a continuous record for a semi-arid area in SE Spain. The results of this analysis have been plotted as cumulative runoff against cumulative storm rainfall, showing a similar scatter to that seen in figure 2. Equation (1) has been freely fitted to the data, and it can be seen that, without a more detailed knowledge of storm profiles than can be derived from the daily record, it is both impracticable to apply a more sophisticated model, and unwise to make runoff forecasts for any individual storm.

Soil Water

Water infiltrating into the soil is limited by the runoff threshold, which is conceptualised as an available near-surface water store. The upper limit for this store is constrained by soil properties, and currently estimated from mapped soil classes in the European Soils Data Base (Gobin et al, 2004). The store may be decreased where the soil is crusted, and/or if subsurface flow brings saturated conditions close to the surface. Additional considerations apply where the soil is frozen or snow covered. Both sub-surface flow and the near-surface water store are available for evaporation and for evapotranspiration linked to plant growth.

After allowing for interception, evapo-transpiration is partitioned between the vegetated and unvegetated fractions of the surface according to the proportional vegetative crown cover. Interception is calculated as a fraction of rainfall rather than a fixed capacity, and this fraction increases with vegetation biomass (Llorens et al, 1997). Each evapotranspiration component is associated with a rooting depth, according to the land cover type for the vegetated area and normally set at 10mm for the bare soil. For each component, potential evaporation (PE), after subtraction of interception, is then reduced exponentially to an actual rate (AE) of:

$$AE = WUE.PE.\exp(-D / h_R) \qquad (2)$$

Where WUE = water use efficiency for stage of plant growth (or 1.0 for bare soil)

D is saturated subsurface deficit

and h_R is the rooting depth for each partiton.

Contributions to evaporation are weighted for the fractional plant cover to give a combined estimate.

Subsurface flow is estimated using TopModel (Beven and Kirkby, 1979), with topographic properties estimated from local relief (from DEM) and soil parameters (saturated hydraulic conductivity and TopModel soil parameter, m) from the soil type. The average saturated deficit is estimated in monthly steps, to provide the background hydrological conditions and, in particular, the saturation constraint on the runoff threshold which controls overland flow runoff in each storm. Deficit is updated monthly from the TopModel expression:

$$\begin{aligned} D &= D_0 + m \ln\left\{ \frac{j_*}{i} \exp\left(-\frac{D_0}{m}\right) + \left[1 - \frac{j_*}{i} \exp\left(-\frac{D_0}{m}\right)\right] \exp\left(-\frac{it}{m}\right) \right\} \mathit{for}\ i \neq 0 \\ D &= D_0 + m \ln\left[1 + \frac{j_* t}{m} \exp\left(-\frac{D_0}{m}\right)\right] \ \mathit{for}\ i = 0 \end{aligned} \qquad (3)$$

where D is the deficit after time t (as in equation 2)

D_0 is the initial deficit,

i is the net rainfall intensity

m is the TopModel soil parameter

and j_* is the average saturated runoff rate.

This expression also estimates the net subsurface runoff over the month as

$$D - D_0 + it = m \ln\left[1 - \frac{j_*}{i} \exp\left(-\frac{D_0}{m}\right) + \frac{j_*}{i} \exp\left(\frac{it - D_0}{m}\right)\right] \qquad (4)$$

In these calculations the total net rainfall is used, corrected for the overland flow runoff where this is a significant fraction.

This combination of an infiltration excess mechanism, represented by the bucket model, with a saturation excess mechanism, represented by TopModel, provides a robust hydrological sub-model which provides an adequate response across the humid to semi-arid continuum. As will be seen below, the evapo-transpiration stream is also used to drive a simple plant growth model which is also responsive to this range of conditions.

Cold Climate modifications

Where temperatures fall below freezing, the hydrological model needs to respond to snow and frozen soil conditions. For the monthly model, the range of temperatures is used to estimate the proportional time below freezing, and the day-degrees above and below freezing. Rainfall is assumed to fall as snow for the fraction of each month freezing, and lying snow is accumulated and melted according to a linear degree-day model. Next a depth of soil freezing is calculated using a simple conductivity model, and assuming that the snow pack has a conductivity 20% that of the soil. This is equivalent to an accumulated day-degree model, with the calculated freezing depth proportional to the square root of the day-degree sum. The effective soil storage capacity is then allowed to fall exponentially with the estimated freezing depth, increasing the estimated overland flow runoff. However, practical experience suggests that both saturation excess overland flow and snowmelt runoff are less flashy, and therefore less erosive than infiltration excess overland flow, so that the corresponding erosion estimates are reduced heuristically.

Soil properties

The runoff threshold for infiltration excess overland flow is estimated as an area-weighted average of the thresholds under vegetation and in the bare gaps between.

Under vegetation, rainfall is lost to interception, and the runoff threshold is calculated as the lesser of two values:

(1) available near-surface water storage capacity (depending on soil textural properties), or

(2) the sub-surface saturation deficit (from the TopModel estimate described above)

In arable areas, surface roughness represents the full storage capacity of furrows immediately after ploughing, and this decays exponentially with time in the subsequent period, eventually falling to a minimum value representing the textural roughness of the surface. Naturally vegetated areas are also assumed to present this minimum roughness.

Bare areas are also considered to be subject to crusting, with a tendency to crusting referred to mapped soil classes, largely interpreted in textural terms as a minimum runoff threshold for a fully crusted surface (Le Bissonnais et al, 2002). For arable areas, the runoff threshold for a bare area is re-calculated as beneath vegetation immediately after tillage, this decays exponentially towards the minimum for each soil type with accumulated monthly rainfalls.

This formulation provides a seasonal response in runoff thresholds, and therefore in infiltration excess overland flow. For a conventionally ploughed annual crop, for example, thresholds are high on first planting, but fall very rapidly immediately afterwards, particularly if there is rain, as crusting develops while the crop provides little cover. As the crop grows, the runoff threshold recovers, recovering to high values as the crop matures. After harvest these high values fall again, depending on how or whether the surface is protected. Under natural vegetation there is much less annual variation, with runoff thresholds responding to the seasonality of cover.

The distribution of infiltration excess overland flow in storms

Storm rainfalls are considered as independent random events, defined by a frequency distribution for each month of the year. The autocorrelation between successive events is weakly represented by the seasonal variations in soil moisture, but there is some loss of information by using this approach. This represents a trade-off between simplicity and accuracy, with the least impact on estimates for the semi-arid areas where soil erosion is generally considered to be most severe, because soils normally dry out between major events.

As noted above, daily rainfalls have been used as the basis for analysis because, while recognising the limitations of this approach, it allows the use of the widespread daily precipitation data. On a month by month basis, daily rainfalls are analysed to give monthly total, mean rain per rain-day and the standard deviation of rainfalls on rain-days. These statistical moments allow fitting most observed data for daily rainfalls to the probability density function for a Gamma distribution as follows:

$$pd(R) = \frac{\alpha}{\overline{R}} \frac{\left(\alpha R/\overline{R}\right)^{\alpha-1}}{\Gamma(\alpha)} \exp\left(-\alpha R/\overline{R}\right) \tag{5}$$

where $\overline{\mathrm{R}}$ is the mean rain per rainday

and $\alpha = (1/\mathrm{CV})^2$

where CV is the coefficient of variation $= \sigma / \overline{R}$

Figure 4 shows an example of the cumulative frequency distribution for data from SE Spain. The gamma distribution has been found to provide a robust fit, giving a good balance between small and large events. The CV is generally between zero and unity, so that the probability density distributions peak at zero rainfall.

Infiltration excess overland flow for a storm of rainfall *R* is then given by equation (1) above, and the total overland flow runoff for the month integrated numerically as:

$$\sum r = \int_{R_0}^{\infty} (R - R_0) \frac{\alpha}{\overline{R}} \frac{\left(\alpha R/\overline{R}\right)^{\alpha-1}}{\Gamma(\alpha)} \exp\left(-\alpha R/\overline{R}\right).dR \qquad (6)$$

This is used directly as a component of the water balance, but it will be seen below that a power of event runoff is used to estimate sediment transport. For a power law of 2.0, the corresponding summation of $(\text{Runoff})^2$ then takes the form:

$$\sum r^2 = \int_{R_0}^{\infty} (R - R_0)^2 \frac{\alpha}{\overline{R}} \frac{\left(\alpha R/\overline{R}\right)^{\alpha-1}}{\Gamma(\alpha)} \exp\left(-\alpha R/\overline{R}\right).dR \qquad (7),$$

And similarly if other powers are used. This then gives the correct strong weighting to the largest events in the accumulated total.

Land use and vegetation cover

The hydrological components of the model, as described above, are strongly dependent on vegetation cover, which is understood to be a major control on both runoff and erosion. Figure 5 illustrates the effect of changed land cover in a loess area with 1500-2000 mm annual precipitation. It can be seen that runoff on bare soil exceeds 80%, and falls to 2% under a dense vegetation cover, and that this 40-fold difference in runoff gives a 2000-fold difference in sediment loss. Other experiments (e.g. Hudson and Jackson, 1959) have shown that fine netting stretched above the surface of an agricultural field has almost as strong an effect as dense vegetation in reducing runoff and erosion. Thus the importance of crown cover for both runoff and erosion is extremely strong, although it is recognised that root and soil organic matter effects are also important for uncultivated areas.

Land cover has been approached in the model through two alternative strategies, each of which has its advantages; first through direct remote sensing of land cover and second through modelling vegetation growth. Geomatic data has the advantage that it provides a direct measure of real vegetation abundance, which is now available monthly for a period of over twenty years, through AVHRR and LANDSAT images. This integrates the effects of all impacts on the cover in an unambiguous historical record. It therefore includes the impacts of factors which may not all be fully incorporated in a model. However, the analysis is based on the best of three monthly satellite passes, and suffers from the persistence of cloud cover in Northern Europe and other humid areas. It also lacks any direct forecasting potential, and therefore has limited applicability for analyses of scenarios for land use and/or climate change.

Vegetation growth models are well established, with both generic and crop-specific models. The models applied here have been based on a biomass carbon balance for both living vegetation and soil organic matter. Such models may be insufficiently parameterised to cover the full range of functional types, and are commonly limited by absence or inadequate representation of some processes. Fire and grazing are, for example, not directly represented in the models that have been used to date with PESERA. As a result, the vegetation cover is more a ‘potential’ than actual cover, with only indirect parameterisation of some relevant influences. However, growth models respond directly to changes in land use or climate drivers, and so have greater scenario potential.

Analysis of RS images can be based directly on NDVI, but improved results have been obtained using the satellite-derived surface temperature to correct for water content, linearly unmixing in a phase-space triangle between water, vegetation and soil. This gives a measure of vegetation abundance which can be empirically related to cover and/or above ground biomass, and from which some land use classes can be interpreted from the seasonal cover cycle. (Haboudane et al, 2002).

The generic vegetation model estimates gross primary productivity (GPP) as proportional to the plant actual transpiration. This is offset by respiration, at a rate increasing exponentially with temperature and proportional to biomass. Leaf fall fraction is a decreasing function of biomass, to allow for a larger structural component in large plants. Where respiration is greater than gpp, a 'deciduous' response increases an additional leaf fall at a rate that increases with temperature. Finally vegetation biomass may lose a fraction to grazing or plant gathering activities.

Soil organic matter is increased by leaf fall, except where crops are harvested, and decomposes as a single linear store at a rate that increases with temperature.
Cover is calculated independently, with reference to an equilibrium cover defined as the ratio of plant transpiration to potential evapotranspiration rate. Cover converges on this (changing) equilibrium value at a rate which is larger where biomass is small, and is the variable which drives the seasonal partition of runoff threshold between

vegetated and bare areas. This generic model has been calibrated against global distributions of biomass (Kirkby and Neale, 1987). Crop models are variants of this generic model, with additional controls through data on regional patterns of planting and harvest dates, and with an evolution of water use efficiency through the life cycle of the crop (Gobin and Govers, 2003).

Accumulation of runoff discharge downslope

Runoff generated locally may not reach the base of the slope to deliver sediment to a channel, and the runoff coefficient for infiltration excess overland flow has therefore generally been observed to decrease with distance or area downslope. The two dominant reasons for this reduction are thought to be (Kirkby et al, 2005) the patchiness of local runoff generation and the short duration of bursts of intense rainfall within storms. Patchiness occurs at several scales: for uncultivated areas the alternation of shrubs or tussocks of grass with bare areas provides contrasts at the scale of a few metres; while the patchwork of fields with different land use and/or tillage directions provides a coarser mosaic in cultivated areas. If there is good connectivity between areas of above average runoff, then there may be substantial runoff even in storms which do not reach the average runoff threshold. More commonly, however, patches of runoff re-infiltrate within more absorbent areas. Close to a channel or other collector, some patches of enhanced runoff connect directly with the channel, but little reaches the channel from farther away because of intervening re-infiltration (Cammeraat, 2002). The result is that discharge increases with distance downstream only over a distance scaled to the patch size, and then levels off to a near-constant value. In larger or more intense storms, where runoff is generated over an increasing proportion of the area, the region of increasing discharge also increases with the size of individual connected patches.

The second important mechanism for limiting discharge accumulation is that storms, even substantial storms, commonly consist of short (<30 minutes) periods of intense rain (>10mm hr^{-1}) with longer periods at low intensity (Kirkby et al, 2005). During these intense bursts, runoff is generated, and begins to flow downslope at average velocities which are generally of the order of only 1-3 cm s^{-1}. When the intensity falls, this flowing water re-infiltrates, and only reaches the channel from a zone 18 - 55 m wide (in 30 minutes). As for spatial patchiness, this gives a band of increasing discharge and a band of constant discharge; and the width of the band again tends to be greater in larger storms.

Summed over the distribution of storm sizes described above, these factors lead to a less then linear increase of discharge with distance downslope, and this has generally been represented as a logarithmic or power law (with exponent ~ 2/3) relationship. Similar arguments can be applied to saturation excess overland flow to suggest power law exponents >1, but this is not pursued here since the saturation overland flow is generally less flashy and therefore less effective in erosion.

Sediment transport and sediment yield

Estimates of sediment transport are based on infiltration excess overland flow discharge, which has been discussed above. Most sediment transport equations are based on considerations of tractive stress or flow power, and commonly generalised into a power law in discharge and gradient, thus avoiding a more detailed analysis of flow thread geometry. The commonest formulations (e.g. Musgrave 1947) assume that there is an ample sediment supply, and that sediment is everywhere transported by soil erosion at its transporting capacity C, expressed in the form:

$$C = kq^m \Lambda^n \tag{8}$$

where k is the soil erodibility,
Q is the overland flow discharge per unit flow width
Λ is the local slope gradient,
and m, n are empirical exponents, generally in the ranges m = 1.5-3; n = 1-2

In such expressions, discharge is generally associated with distance from the divide, possibly with a change in the exponent m. It has generally been found that the performance of erosion models is remarkably insensitive to the choice of exponents, largely because slope and distance tend to change together., and exponent values of m = 2, n = 1 have therefore been adopted, with computational advantages that are evident below.

Evaluation of appropriate exponents may be made at a range of time and space scales (e.g. Kirkby et al, 2002). The most direct approach is through soil erosion plots, but these are often not corrected for the frequency distribution of storms to provide meaningful long term averages. A second approach is by looking at the critical areas required to support an ephemeral gully formed in a particular storm. This approach requires an analysis of the stability of small depressions, as a balance is reached between infilling by diffusive processes, primarily rainsplash in relevant contexts and their enlargement by soil erosion (rillwash) processes. A third approach is by back analysis of hillslope profile form, which is formed over a period in response to the full distribution of events. The difficulty with this approach lies in uncertainty about whether the observed landscape form has developed under process conditions that are still current, or are inherited from conditions of different climate and/or land cover.

The values that have been adopted here lie within the empirical range, and will be seen to have additional advantages in creating a consistent coarse scale model. Here it is proposed to use:

$$C = k(rx)^2 \Lambda \tag{9}$$

where r is the local runoff in each event, from equation (1) above,

and x is the distance from the divide.

Summing over the frequency distribution of events in any month, the mean total sediment transport takes the form:

$$\sum C = kx^2 \Lambda . \sum r^2 \tag{10}$$

In which the final term may be taken from equation (7) above.

Alternatives to this composite power law approach can simulate selective transportation of different grain sizes, for example by defining transport capacity as the product of detachment rate and travel distance. This approach has the advantage of allowing a spectrum of responses, from a strictly transport limited approach for the coarser soil fractions, to a detachment or supply limited approach for the finest material. Although this approach has merit, there is not sufficient data to properly parameterise it for the proposed coarse scale model. In practice this means that the soil erodibility for fine soils must implicitly be reduced to allow for the limited rate of supply, whether through hydraulic erosion or through removal of previously detached material, and that, for rangeland, selective transportation creates an armour layer over time that reduces erosion rates.

In the PESERA model, sediment transport is interpreted as the mean sediment yield delivered to stream channels, and includes no allowance for downstream routing within the channel network. Sediment Yield *Y* is the sediment transported to the slope base, averaged over the slope length, that is:

$$Y = \frac{\sum C_B}{L} = k\frac{L^2\Lambda_B}{L}\sum r^2 = kL\Lambda_B\sum r^2 \qquad (11)$$

where the suffix *B* indicates evaluation at the slope base, and $L = x_B$ is the total slope length.

The term $L\Lambda_B$ can be expressed, in terms of the total slope relief, $H = L\overline{\Lambda}$, where $\overline{\Lambda}$ is the average slope gradient from crest to base, giving:

$$Y = \varsigma\, kH\sum r^2 \qquad (12)$$

Where $\varsigma = \Lambda_B / \overline{\Lambda}$ is the ratio of slope base to average gradient, a number which is generally less between 0.5 and 1.0 for typical convexo-concave slopes. This correction term can be included where available, but generally defaults to a slight correction in the empirical value for erodibility, *k*.

Equations (11) and (12) are taken as the final form of the expression used in the PESERA model. It may be seen to include three terms:

1. Soil erodibility, which is derived from soil classification data, primarily interpreted as texture (Le Bissonnais et al, 2002).

2. Local relief, which is derived from DEM data as the standard deviation of elevation around each point.
3. An estimate of accumulated (runoff)^2, which is derived from a biophysical model that combines the frequency of daily storm sizes with an assessment of runoff thresholds based on seasonal water deficit and vegetation growth.

Implementation

Currently, the PESERA model can be implemented in two modes. Firstly, to provide an estimate of sediment yield at a point, this is carried out in Excel, supported by Visual Basic Macros, and secondly to produce a distributed estimate of erosion risk, this is achieved in FORTRAN, operating on data extracted from ARC-GIS grids (PESERA-GRID). The same algorithm is applied to each cell in the grid. Although a reduced information system (e.g Brazier et al, 2001) was considered, the number of possible combinations was considered too great to provide significant computational savings without severely restricting the number of possible values for the 128 variables.

Actual erosion is very strongly impacted by the incidence of particular large storms, and the approach adopted makes no attempt to provide a forecast, but estimates the long term average erosion rate over a long series of years. This is considered to be appropriate for assessing the spatial distribution of erosion risk at a regional scale.

On executing the model in either of the two modes, the annual cycle of monthly values are applied repeatedly until the outputs stabilise in an annual cycle. This reduces the dependence on initial conditions. The hydrological components are generally found to stabilise rapidly, within 3-5 years, Figure 6. Vegetation components stabilised more slowly, with a response time increasing with the lifespan and biomass of the plants, but these elements generally stabilise sufficiently within 50 years, and much more quickly (~10 years) for annual crops, Figure 7. Outputs are then reported after stabilisation.

Within the point code model, soil erosion is estimated separately for each month and for each segment of the slope profile. This facility offers the potential to explore the sensitivity of slope form in sediment yield. PESERA_GRID operates on local relief, estimated as the standard deviation of elevation within a defined radius. At the 1-km scale only the immediate cells are considered. At finer resolutions a radius is adopted which reflects the hillslope scale. The model has been used in preparing the Pan European estimates of soil erosion risk under current climate and land-use conditions, Figure 8.

The PESERA-GRID model has been developed primarily in Fortran90 with Arc Micro Language (AML) modules to extract data and convert back to GRID. The Fortran90 executables are compiled and distributed in PC format, requiring at least .

512 RAM and 60GB Hard Drive Space for the European 1 km Grid.

Data Input and Output (PESERA_GRID103)

Monthly climate data contributes the majority of data layers required to execute that drive the PESERA model. Other data layers are derived from a number of sources that primarily describe: land-use , crops and planting dates; soil storage and erodibility and relief. A set of 128 input data layers are required. Where local data is available at higher resolution this local data can be utilised at the users discretion. However, as data resolution is refined (< 100m grid resolution) assumptions applied in the development of the PESERA model may not hold, particularly with respect to assuming that all cells drain directly to the channel network, and therefore do not accumulate from cell to cell. The standard input and output variables for the PESERA model are listed in Tables 2 to 6.

Conclusion

The PESERA model may be seen to have a secure theoretical base, although the accuracy of forecasts is limited by the restriction, based on data availability, to daily rainfall data, and to a greatly simplified analysis of topography. Within these constraints, the model responds both rationally and in accordance with established principles to variations in climate, land use and topography. An important component of the confidence placed in the model lies in this internal validation, in which the model is an explicit up-scaling and simplification of principles that have been widely accepted and validated at finer scales.

In the second part of this paper, the application of the model will be tested against the limited erosion plot data available, which has been used to provide an overall calibration, particularly of the erodibility values and range. We also discuss the use of the model to provide erosion scenarios, in response to climate and land use scenarios drawn from Global Climate Models and literature on trends in land use.

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References cited

Bennett, H.H., 1939. *Soil Conservation*, New York: McGraw-Hill Book Company,

Beven, K.J. and M.J. Kirkby, 1979. Towards a simple, physically based, variable contributing area model of catchment hydrology. *Int Ass Hydrol Sci, Bull* **24**, 43-69.

Brazier, R.E., Rowan, J.S., Quinn, P. and Anthony, S., 2001. Towards an MIR (Minimum Information Requirement) approach to modelling on-site soil loss at the National scale. *Catena*, **42**, 59-79

Cammeraat, L.H., 2002. A review of two strongly contrasting geomorphological systems within the context of scale. *Earth Surface Processes and Landforms* **27**, 1201-1222.

Cerdan, O, 2003. Long-term Soil Erosion Plot Data to evaluate the PESERA (Pan-European Soil Erosion Risk Assessment) Approach**.** *Geophysical Research Abstracts*, **5**, 00439, European Geophysical Society.

de Ploey, J., M.J. Kirkby & F. Ahnert, 1991. Hillslope erosion by rainstorms - a magnitude-frequency analysis. *Earth Surface Processes & Landforms* **16**, 399-409

DeRoo APJ, 1996. LISEM: A Single-Event Physically Based Hydrological And Soil Erosion Model For Drainage Basins .1. Theory, Input And Output. *Hydrological Processes* **10** : 1107

European Commission 2006. Proposal for a directive of the European parliament and of the council establishing a framework for the protection of soil and amending. *Directive 2004/35/EC*, 30pp.

Gobin, A. & Govers, G. (eds) 2003. *PESERA-Third annual report*. EC Contract: QLK5-CT-1999_01323, Chapter 4, 60-81.

Gobin, A., Jones, R., Kirkby, M., Campling, P., Govers, G., Kosmas, C., Gentile. A.R., 2004. Indicators for pan-European assessment and monitoring of soil erosion by water. *Environmental Science & Policy*, **7** (1): 25-38

Green, W.H. and Ampt, G.A., 1911. Studies in Soil Physics. 1. The flow of air and water through soils. *Journal of Agricultural Soils* **4**, 1-24.

Haboudane, D., F Bonn, A. Royer, S. Sommer and W. Mehl, 2002. Land Degradation and Erosion Risk Mapping by Fusion of Spectrally Based Information and Digital Geomorphometric Attributes. *Int. J. Remote Sensing*, **23**, 3795–3820

Hudson, N.W. and Jackson, D.C., 1959. Erosion research. *Henderson Research Station, Report of progress 1958-9*. Federation of Rhodesia and Nyasaland; Ministry of Agriculture.

Kirkby, M.J., 1998. Landscape modelling at regional to continental scales. *Process modelling and landform evolution* (ed. St. Hergarten & H.J. Neugebauer), LNES 78, Springer Verlag, 189-208.

Kirkby, M.J., L.J. Bracken, J. Poesen, J. Nachtergaele and L. Vandekerckhove, 2003. Observed and modelled distributions of channel and gully heads - with examples from S.E. Spain and Belgium. *Catena* **50**, 415-434.

Kirkby, M.J., Bracken, L.J. and Reaney, S.J., 2002. The influence of landuse, soils and topography on the delivery of hillslope runoff to channels in SE Spain. *Earth Surface processes and Landforms* **27** (13), 1459-1473.

Kirkby, M.J., Bull, L.J. and Shannon,, J., 2005. The influence of rainfall distribution and morphological factors on runoff delivery from dryland catchments in SE Spain. *Catena* **62**, 136-156.

Kirkby, M.J. and N.J. Cox, 1995. A climatic index for soil erosion potential (CSEP) including seasonal and vegetation factors. *Catena* **25**, 333-52.

Kirkby, M.J., A.C. Imeson & G. Bergkamp, 1996. Scaling up processes and models from the field plot to the watershed and regional areas. *J. Soil & Water Conservation* **51**(5), 391-396

Kirkby, M.J., Le Bissonais, Y., Coulthard, T.J., Daroussin, J. and McMahon, M.D., 2000. The development of Land Quality Indicators for Soil Degradation by Water Erosion. *Agriculture, Ecosystems and Environment*, **81**: 125 - 136

Kirkby, M.J. and R.H. Neale, 1987. A soil erosion model incorporating seasonal factors. *International Geomorphology*, **II**, John Wiley, 289-210.

Le Bissonnais Y., C. Montier, M. Jamagne, J. Daroussin, D. King (2002). Mapping erosion risk for cultivated soil in France. *Catena*, **46**, 207-220.

Llorens, P., Ramon Poch, R.,Latron, J., Gallart, F., 1997. Rainfall interception by a *Pinus sylvestris* forest patch overgrown in a Mediterranean mountainous abandoned area, I. Monitoring design and results down to the event scale *Journal of Hydrology* **199**, 331-345

Meginnis, H.G., 1935. Effect of cover on surface runoff and erosion in the Loessial uplands of Mississippi. *U.S. Department of Agriculture, Circular* **347**, 15pp.

Morgan RPC, Quinton JN, Rickson RJ, 1994. Modeling Methodology For Soil-Erosion Assessment And Soil Conservation Design - The Eurosem Approach. *Outlook On Agriculture* **23** (1): 5-9

Musgrave, G.W., 1947. Quantitative evaluation of factors in water erosion. *Journal of Soil and Water Conservation* **2**, 133-138.

Nearing, M.A., G.R. Foster, L.J. Lane and S.C. Finkner, 1989. A process-based soil-erosion model for USDA-water erosion prediction project technology. *ASAE Trans*, **32**, 1587-1593.

Philip, J.R., 1957. The theory of infiltration: IV. Sorptivity and algebraic infiltration equations. *Soil Science* **84**, 257-264.

Renard, KG; Foster, GR; Weesies, GA; Porter, JP, 1991. RUSLE. Revised universal soil loss equation. *Journal of Soil and Water Conservation*. **46** (1), 30-33.

Seely M, Wohl H, 2004. Connecting research to combating desertification. *Environmental Monitoring And Assessment* **99** (1-3): 23-32

Smith RE, Goodrich DC, Quinton JN. Dynamic, Distributed Simulation Of Watershed Erosion - The Kineros2 And Eurosem Models. *Journal Of Soil And Water Conservation* **50** (5): 517-520

Van Der Knijff, J.M., Jones, R.J.A. and Montanarella, L., 2000. *Soil Erosion Risk Assessment in Europe*, EUR 19044 EN, 34pp

Wischmeier, W.H. and Smith, D.D., 1958. Rainfall energy and its relationship to soil loss. *American Geophysical Union, Trans*, **39**, 285-291.

Wischmeier, W.H. & Smith, D.D. ,1978. Predicting rainfall erosion losses –a guide for conservation planning. *U.S. Department of Agriculture, Agriculture Handbook* **537**

Yuan Y, Mitchell JK, Hirschi MC, et al, 2001. Modified SCS curve number method for predicting subsurface drainage flow. *Trans. Am. Soc. Civil Engineers* **44** (6): 1673-1682

Figure Captions

1. Schematic hydrological model within the PESERA model
2. Measured rainfall runoff data for storms in a small US catchment. Straight lines indicate application of a linear bucket model with R_0 = 25 mm and p = 0.67.
3. Storm runoff profiles generated for 76 storms over a 3-year period for the Torrealvilla catchment, Murcia, SE Spain. Black dots are generated using a Green Ampt equation (A = 4 mm hr^{-1}; B = 10 $mm^2 hr^{-1}$). Grey line generated from a bucket model with R_0=10 mm; p=0.40.
4. Cumulative frequency distribution for November and December daily rainfalls 1997-2002 at site Nogalte, North 2 (Murcia, SE Spain) fitted to Gamma distribution.
5. Relationship between annual runoff and sediment loss as vegetation cover is altered. Loess soils, Holly Springs, MI. data from Meginnis, 1935.
6. Stabilisation of hydrology in the PESERA model from arbitrary initial conditions.
7. Stabilisation of natural vegetation cover in PESERA model from zero initial conditions
8. Final Pan-European estimates of soil erosion risk for current land use and climate.

Table 1: Comparing exponents for sediment transport v discharge between instantaneous and event-integrated values

Change of hydrograph form with flood volume(..& time)	*Relationship between event total exponent (ETE) and instantaneous exponent (IE)*
Fixed duration	ETE=IE
Fixed peak flow	ETE =1
Fixed shape (peak: duration) ratio	ETE<IE
Larger floods flashier than smaller	ETE>IE

Table2: Monthly Climate Input Data (96 data layers = 8 layers for each month)

Parameter name	**Range of values**	**Units**	**Current Source at European scale**	**Description/Source**
meanrf130_	0-300	mm/mo	BADC/MARS	Mean monthly rainfall
meanrf2_	0-50	mm/d	BADC/MARS	Mean rainfall per rain day (by month)
cvrf2_	1-10	-	BADC/MARS	Coefficient of variation of rain per rain day (by month: computed for rain days only)
mtmean_	-32.4 – 37.3	^{o}C	BADC/MARS	Mean monthly temperature Corrected for altitude
mtrange_	2.4 – 18.4	^{o}C	BADC/MARS	Temperature range (monthly) (Mean daily max – Mean daily min)
meanpet30_	0-300	mm/mo	BADC/MARS	Mean monthly PET Corrected for altitude
newtemp_	-	^{o}C	HADLEY[3]	Predicted future temperature (scenario by month)
newrf130_	-	mm/mo	HADLEY	Predicted future rainfall (scenario by month)

Table 3: Land-use, Crops and Planting date Input Data (25 data layers)

Model Parameter	Range of values	Units	Source	Description/Source
use	-	-	CORINE ABM Survey	Land cover type/management option
eu12crop1	-	-	CORINE/FSS ABM Survey (if applicable)	Dominant Arable Crop
maize_210c	-	-		Maize Crop (default)
eu12crop2	-	-		2nd Dominant Arable Crop
itill_crop1	1-12	-	FSS/PDD ABM Survey (if applicable)	Planting month: Dominant Arable Crop
itill_maize	1-12	-		Planting month: maize
itill_crop2	1-12	-		Planting month: 2nd Dominant Arable Crop
mitill_1	0/1	-	FSS/PDD ABM Survey (if applicable	Planting marker: Dominant Arable Crop
mitill_m	0/1	-		Planting marker: maize
mitill_2	0/1	-		Planting marker: 2nd Dominant Arable Crop
cov_	0-100	%	CORINE/FSS ABM survey, model or data	Ground cover (12 monthly values) – input as management or output from growth model.
rough0	0,5,10	mm	CORINE	Initial surface storage

rough_red	0,50	%	Literature	Surface roughness reduction per month
rootdepth	10-1000	mm		Rootdepth
effective ditch density	0-100	m/km^2		
fire				Frequency and timing of deliberate burns
grazing intensity				Grazing density or fraction of available biomass removed

FSS: Farm Structure Survey (EuroStat): PDD: Planting dates database (Van Orshoven et al., 1999)

Table 4:. Soil Parameters Input Data (6 data layers)

Model Parameter	**Range of values**	**Units**	**Source**	**Description/Source**
crusting	1-5	mm	SOIL DB	Crust storage
erodibility	1-5	mm	SOIL DB	Sensitivity to erosion
swsc_eff_2	0-205	mm	SOIL DB	Effective soil water storage capacity
p1xswap1	0-90	mm	SOIL DB	Soil water available to plants in top 300mm
p2xswap2	0-154	mm	SOIL DB	Soil water available to plants : (300mm and 1000mm depth)

Grid ZM

Description: Scale depth (TOPMODEL) derived from soil texture

Soil Texture		**zm (mm)**
Coarse	**C**	30
Fine	**F**	10
Medium	**M**	20
Medium Fine	**MF**	15
Organic Soils	**O**	10
Very Fine	**VF**	5

Grid TEXT:

Description: Soil Texture

Source: Soil Geographical Database of Eurasia at scale 1:1,000,000 version 4.0 beta, European Soil Bureau, SAI/JRC Ispra.

0 No information

9 No mineral texture (Peat soils)

1 Coarse (18% < clay and > 65% sand)

2 Medium (18% < clay < 35% and >= 15% sand, or 18% <clay and 15% < sand < 65%)

3 Medium fine (< 35% clay and < 15% sand)

4 Fine (35% < clay < 60%)

5 Very fine (clay > 60 %)

Table 5: Topographic Input data (1 data layer)

Model Parameter	**Range of values**	**Units**	**Source**	**Description/Source**
std_eudem2	-	m	GTOPO30/ SRTM90/ digimap	Standard deviation of elevation.

Table 6: Output variables for each cell in the PESERA model *(6 variables for each of 12 months)*

Output Parameter name	**Units**	**Sub-Routine**	**Description**
sedi_out	tonnes/ha	***erosion***	Erosion (monthly)
runoff	mm	***veggrowth***	Overland flow runoff (monthly)
deficit	mm	***veggrowth***	soil water deficit (monthly)
xint	%	***veggrowth***	percentage interception (monthly)
veg	(kg/m^2)	***veggrowth***	Vegetation biomass (monthly)
Cover	%	***veggrowth***	Cover monthly (if not pre-set by land use)
hum	(kg/m^2)	***veggrowth***	Soil organic matter biomass (monthly)