

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

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12

13 **Abstract**

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

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

31

32

33 **Introduction**

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

45

46 Here we present a theoretical framework for developing a simplified process based
47 model, in particular providing explicit dependence on climate and vegetation, and
48 implement this model as a 1-km resolution model across most of Europe. This paper
49 describes the innovative aspects and theoretical background to the model, and a

50 companion paper will describe its performance, calibration scenario application at the
51 European scale.

52

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

63

64 **The importance of soil erosion at a regional scale**

65 Erosion by running water has been identified as the most severe hazard threatening the
66 protection of soil in Europe (EC, 2006). By removing the most fertile topsoil, erosion
67 reduces soil productivity leading, where soils are shallow, to a progressive and ultimately
68 irreversible loss of natural farmland, and in vulnerable areas, is one major process of
69 desertification. Severe erosion is commonly associated with the development of temporary
70 or permanently eroded channels or gullies which can fragment farmland. The soil and
71 runoff removed from the land during a large storm accumulates below the eroded areas,
72 spilling offsite and in severe cases blocking roadways or channels and inundating buildings.
73 Erosion rate is very sensitive to both climate and land use, as well as to detailed
74 conservation practice at farm level. In a period of rapid changes in both climate and land

75 use, due to global change, revised agricultural policies and international markets, it is
76 valuable to be able to assess the state of soil erosion at a European level, using an objective
77 methodology which allows the assessment to be repeated as conditions, pressures and
78 drivers, change, or to explore the broad scale implications of prospective global or Europe-
79 wide changes. This provides an estimate of the overall costs attributable to erosion under
80 present and changed conditions, and objectively suggests areas for more detailed study and
81 possible remedial action.

82

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

89

90 **Process model approach**

91 There are a number of possible methodologies for creating a coarse scale erosion map
92 (Gobin et al, 2004). Some of these are based on the collection of distributed field
93 observations, others on an assessment of factors, and combinations of factors, which
94 influence erosion rates, and others primarily on a modelling approach. All of these
95 methods require calibration and validation, although the type of validation needed is
96 different for each category. There are also differences in the extent to which the
97 assessment methods identify past erosion of an already degraded soil resource, as
98 opposed to risks of future erosion, under either present climate and land use, or under

99 scenarios of global change. Here a physically-based process model is presented,
100 within the limitations of resolution and available data.

101

102

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

115

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

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

123

124

125 **Point hydrology and land cover**

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

132

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

142

143 *Runoff in a single storm*

144 Figure 1 outlines the hydrological balance within the PESERA model. Precipitation is
145 divided into daily storm events, expressed as a frequency distribution, that drive infiltration
146 overland flow and soil erosion, and monthly precipitation, some of which may be as snow,
147 driving saturation levels in the soil. Infiltration excess overland flow runoff is estimated
148 from storm rainfall and soil moisture. Sediment transport is then estimated from overland

149 flow and routed, in principle, downslope. Alternative methods for making these estimated
150 are discussed below. To obtain long term estimates of soil erosion these estimates must then
151 be scaled up by integrating over time. This process of scaling up has two stages, first from
152 momentary to event-integrated dependence, and secondly from events to long term averages
153 via the frequency distribution. For the first stage, if instantaneous sediment discharge can be
154 expressed as a power law dependence on instantaneous water discharge, the relationship
155 between event total sediment discharge and event total discharge will, in general, also be a
156 power law, but the exponent will differ according to how hydrograph form changes with
157 flood volume. Table 1 indicates how different generalisations of storm profile influence the
158 relationship between instantaneous and time-averaged exponents. Other possibilities exist if
159 there are thresholds for movement and/or hysteretic sediment stores, but in general it is
160 reasonable to assume a similar power law relationship between sediment transport and
161 discharge for event totals as for instantaneous values, but with some modification to allow
162 for systematic changes in hydrograph shape.

163

164 In the second stage of scaling up, individual storm totals are integrated over the frequency
165 distribution of storms. Two assumptions are normally made, first that the distribution of
166 storms can be replaced by the distribution of daily rainfalls, and second that overland flow
167 can be estimated on the basis of monthly average soil moisture conditions. The first of these
168 assumptions avoids the discussion of how rainfall is divided, more or less arbitrarily, into
169 storm events. The use of a daily unit is both convenient, in that daily rainfall data is
170 relatively widely available, and appropriate in the sense that bursts of rainfall within a single
171 day are significantly influenced by raised soil moisture levels from previous bursts, whereas
172 for longer periods there may be significant drying between bursts. Similarly monthly
173 updating of soil moisture is sufficient to reflect important seasonal differences in weather, to

174 respond to seasonal differences in land cover and to make use of widely available
175 meteorological data. These assumptions are however a compromise, attempting to simplify
176 the estimation of storm runoff while retaining the frequency signature of storms (daily) and
177 soil moisture (monthly).

178

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

187

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

199 constraints of land use and cultivation choices; and third a soil model which estimates the
200 required hydrological parameters from moisture, vegetation and seasonal rainfall history.

201
202 At-a-point soil hydrology can be described through the Richards' equation, although with
203 reservations where both matrix and macropore flow are active. Solutions may be
204 approximated through the use of infiltration equations, such as the Green-Ampt (1911) or
205 Philip (1957) formulations. However these approaches are not compatible with the use of
206 daily time steps, within which the detail of storm profiles is lost, and it is impracticable to
207 provide better estimates of runoff than those from the SCS curve number (Yuan et al, 2001)
208 or a simple bucket model. Here the bucket model is preferred, which offers a simple
209 conceptual insight into the volume of infiltration before runoff occurs, that can be linked
210 directly to concepts of soil moisture storage, as it varies within and between sites. In the
211 bucket model, runoff r is given by:

$$212 \quad r = p(R - R_0) \quad (1),$$

213 In which R is total storm rainfall, R_0 is the runoff threshold, or bucket storage capacity, and
214 p is the proportion of subsequent rainfall that runs off.

215
216 Figure 2 illustrates the typical large scatter in relationships between observed total rainfall
217 and runoff, and none of these models can provide a satisfactory fit to the spread of data for
218 daily time steps, and the bucket model (equation 1) has been adopted in the PESERA model,
219 in which storms are treated as independent random events. Figure 3 shows the application of
220 the Green-Ampt equation, with assumed parameters, to a set of storm events taken from a
221 continuous record for a semi-arid area in SE Spain. The results of this analysis have been
222 plotted as cumulative runoff against cumulative storm rainfall, showing a similar scatter to
223 that seen in figure 2. Equation (1) has been freely fitted to the data, and it can be seen that,

224 without a more detailed knowledge of storm profiles than can be derived from the daily
225 record, it is both impracticable to apply a more sophisticated model, and unwise to make
226 runoff forecasts for any individual storm.

227

228

229 *Soil Water*

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

237

238 After allowing for interception, evapo-transpiration is partitioned between the vegetated and
239 unvegetated fractions of the surface according to the proportional vegetative crown cover.

240 Interception is calculated as a fraction of rainfall rather than a fixed capacity, and this
241 fraction increases with vegetation biomass (Llorens et al, 1997). Each evapotranspiration
242 component is associated with a rooting depth, according to the land cover type for the
243 vegetated area and normally set at 10mm for the bare soil. For each component, potential
244 evaporation (PE), after subtraction of interception, is then reduced exponentially to an actual
245 rate (AE) of:

$$246 \quad AE = WUE.PE.exp(-D / h_R) \quad (2)$$

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

248 D is saturated subsurface deficit

249 and h_R is the rooting depth for each partiton.
 250 Contributions to evaporation are weighted for the fractional plant cover to give a combined
 251 estimate.

252
 253 Subsurface flow is estimated using TopModel (Beven and Kirkby, 1979), with topographic
 254 properties estimated from local relief (from DEM) and soil parameters (saturated hydraulic
 255 conductivity and TopModel soil parameter, m) from the soil type. The average saturated
 256 deficit is estimated in monthly steps, to provide the background hydrological conditions and,
 257 in particular, the saturation constraint on the runoff threshold which controls overland flow
 258 runoff in each storm. Deficit is updated monthly from the TopModel expression:

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

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

261 D_0 is the initial deficit,

262 i is the net rainfall intensity

263 m is the TopModel soil parameter

264 and j_* is the average saturated runoff rate.

265

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

$$267 \quad D - D_0 + it = m \ln \left[1 - \frac{j_*}{i} \exp\left(-\frac{D_0}{m}\right) + \frac{j_*}{i} \exp\left(\frac{it - D_0}{m}\right) \right] \tag{4}$$

268 In these calculations the total net rainfall is used, corrected for the overland flow runoff

269 where this is a significant fraction.

270

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

276

277 *Cold Climate modifications*

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

291

292

293 *Soil properties*

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

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

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

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

302

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

308

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

315

316 This formulation provides a seasonal response in runoff thresholds, and therefore in
317 infiltration excess overland flow. For a conventionally ploughed annual crop, for
318 example, thresholds are high on first planting, but fall very rapidly immediately
319 afterwards, particularly if there is rain, as crusting develops while the crop provides
320 little cover. As the crop grows, the runoff threshold recovers, recovering to high

321 values as the crop matures. After harvest these high values fall again, depending on
322 how or whether the surface is protected. Under natural vegetation there is much less
323 annual variation, with runoff thresholds responding to the seasonality of cover.

324

325

326 *The distribution of infiltration excess overland flow in storms*

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

334

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

$$pd(R) = \frac{\alpha}{\bar{R}} \frac{(\alpha \bar{R} / R)^{\alpha-1}}{\Gamma(\alpha)} \exp(-\alpha \bar{R} / R)$$

341 where \bar{R} is the mean rain per rainday (5)

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

where CV is the coefficient of variation = σ / \bar{R}

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

346

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

$$349 \quad \sum r = \int_{R_0}^{\infty} (R - R_0) \frac{\alpha}{R} \frac{(\alpha R / \bar{R})^{\alpha-1}}{\Gamma(\alpha)} \exp(-\alpha R / \bar{R}) dR \quad (6)$$

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

353

$$354 \quad \sum r^2 = \int_{R_0}^{\infty} (R - R_0)^2 \frac{\alpha}{R} \frac{(\alpha R / \bar{R})^{\alpha-1}}{\Gamma(\alpha)} \exp(-\alpha R / \bar{R}) dR \quad (7),$$

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

357

358

359 *Land use and vegetation cover*

360 The hydrological components of the model, as described above, are strongly dependent
 361 on vegetation cover, which is understood to be a major control on both runoff and
 362 erosion. Figure 5 illustrates the effect of changed land cover in a loess area with 1500-
 363 2000 mm annual precipitation. It can be seen that runoff on bare soil exceeds 80%,
 364 and falls to 2% under a dense vegetation cover, and that this 40-fold difference in
 365 runoff gives a 2000-fold difference in sediment loss. Other experiments (e.g. Hudson

366 and Jackson, 1959) have shown that fine netting stretched above the surface of an
367 agricultural field has almost as strong an effect as dense vegetation in reducing runoff
368 and erosion. Thus the importance of crown cover for both runoff and erosion is
369 extremely strong, although it is recognised that root and soil organic matter effects are
370 also important for uncultivated areas.

371

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

383

384 Vegetation growth models are well established, with both generic and crop-specific
385 models. The models applied here have been based on a biomass carbon balance for
386 both living vegetation and soil organic matter. Such models may be insufficiently
387 parameterised to cover the full range of functional types, and are commonly limited by
388 absence or inadequate representation of some processes. Fire and grazing are, for
389 example, not directly represented in the models that have been used to date with
390 PESERA. As a result, the vegetation cover is more a 'potential' than actual cover, with

391 only indirect parameterisation of some relevant influences. However, growth models
392 respond directly to changes in land use or climate drivers, and so have greater scenario
393 potential.

394

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

401

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

409

410 Soil organic matter is increased by leaf fall, except where crops are harvested, and
411 decomposes as a single linear store at a rate that increases with temperature.

412 Cover is calculated independently, with reference to an equilibrium cover defined as
413 the ratio of plant transpiration to potential evapotranspiration rate. Cover converges on
414 this (changing) equilibrium value at a rate which is larger where biomass is small, and
415 is the variable which drives the seasonal partition of runoff threshold between

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

421

422 *Accumulation of runoff discharge downslope*

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

440 proportion of the area, the region of increasing discharge also increases with the size of
441 individual connected patches.

442

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

452

453 Summed over the distribution of storm sizes described above, these factors lead to a
454 less than linear increase of discharge with distance downslope, and this has generally
455 been represented as a logarithmic or power law (with exponent ~ 2/3) relationship.

456 Similar arguments can be applied to saturation excess overland flow to suggest power
457 law exponents >1, but this is not pursued here since the saturation overland flow is
458 generally less flashy and therefore less effective in erosion.

459

460

461 **Sediment transport and sediment yield**

462 Estimates of sediment transport are based on infiltration excess overland flow
463 discharge, which has been discussed above. Most sediment transport equations are
464 based on considerations of tractive stress or flow power, and commonly generalised

465 into a power law in discharge and gradient, thus avoiding a more detailed analysis of
466 flow thread geometry. The commonest formulations (e.g. Musgrave 1947) assume that
467 there is an ample sediment supply, and that sediment is everywhere transported by soil
468 erosion at its transporting capacity C , expressed in the form:

$$469 \quad C = kq^m \Lambda^n \quad (8)$$

470 where k is the soil erodibility,

471 Q is the overland flow discharge per unit flow width

472 A is the local slope gradient,

473 and m, n are empirical exponents, generally in the ranges $m = 1.5-3$; $n = 1-2$

474 In such expressions, discharge is generally associated with distance from the divide,
475 possibly with a change in the exponent m . It has generally been found that the
476 performance of erosion models is remarkably insensitive to the choice of exponents,
477 largely because slope and distance tend to change together., and exponent values of m
478 $= 2$, $n = 1$ have therefore been adopted, with computational advantages that are evident
479 below.

480

481 Evaluation of appropriate exponents may be made at a range of time and space scales
482 (e.g. Kirkby et al, 2002). The most direct approach is through soil erosion plots, but
483 these are often not corrected for the frequency distribution of storms to provide
484 meaningful long term averages. A second approach is by looking at the critical areas
485 required to support an ephemeral gully formed in a particular storm. This approach
486 requires an analysis of the stability of small depressions, as a balance is reached
487 between infilling by diffusive processes, primarily rainsplash in relevant contexts and
488 their enlargement by soil erosion (rillwash) processes. A third approach is by back
489 analysis of hillslope profile form, which is formed over a period in response to the full

490 distribution of events. The difficulty with this approach lies in uncertainty about
491 whether the observed landscape form has developed under process conditions that are
492 still current, or are inherited from conditions of different climate and/or land cover.

493

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

$$497 \quad C = k(rx)^2 \Lambda \quad (9)$$

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

499 and x is the distance from the divide.

500

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

$$503 \quad \sum C = kx^2 \Lambda \cdot \sum r^2 \quad (10)$$

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

505

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

515 material, and that, for rangeland, selective transportation creates an armour layer over
516 time that reduces erosion rates.

517

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

$$522 \quad Y = \frac{\sum C_B}{L} = k \frac{L^2 \Lambda_B}{L} \sum r^2 = kL\Lambda_B \sum r^2 \quad (11)$$

523 where the suffix B indicates evaluation at the slope base,

524 and $L = x_B$ is the total slope length.

525

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

$$528 \quad Y = \zeta kH \sum r^2 \quad (12)$$

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

533

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

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

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

543

544 **Implementation**

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

554

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

559

560 On executing the model in either of the two modes, the annual cycle of monthly values
561 are applied repeatedly until the outputs stabilise in an annual cycle. This reduces the
562 dependence on initial conditions. The hydrological components are generally found to

563 stabilise rapidly, within 3-5 years, Figure 6. Vegetation components stabilised more
564 slowly, with a response time increasing with the lifespan and biomass of the plants, but
565 these elements generally stabilise sufficiently within 50 years, and much more quickly
566 (~10 years) for annual crops, Figure 7. Outputs are then reported after stabilisation.

567

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

575

576 The PESERA-GRID model has been developed primarily in Fortran90 with Arc Micro
577 Language (AML) modules to extract data and convert back to GRID. The Fortran90
578 executables are compiled and distributed in PC format, requiring at least .
579 512 RAM and 60GB Hard Drive Space for the European 1 km Grid.

580

581 **Data Input and Output (PESERA_GRID103)**

582 Monthly climate data contributes the majority of data layers required to execute that drive
583 the PESERA model. Other data layers are derived from a number of sources that primarily
584 describe: land-use , crops and planting dates; soil storage and erodibility and relief. A set of
585 128 input data layers are required. Where local data is available at higher resolution this
586 local data can be utilised at the users discretion. However, as data resolution is refined (<
587 100m grid resolution) assumptions applied in the development of the PESERA model may

588 not hold, particularly with respect to assuming that all cells drain directly to the channel
589 network, and therefore do not accumulate from cell to cell. The standard input and output
590 variables for the PESERA model are listed in Tables 2 to 6.

591

592 **Conclusion**

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

600

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

606

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611

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703

703 **Figure Captions**

- 704 1. Schematic hydrological model within the PESERA model
- 705 2. Measured rainfall runoff data for storms in a small US catchment. Straight lines indicate
706 application of a linear bucket model with $R_0 = 25$ mm and $p = 0.67$.
- 707 3. Storm runoff profiles generated for 76 storms over a 3-year period for the Torrealvilla
708 catchment, Murcia, SE Spain. Black dots are generated using a Green Ampt equation (A
709 $= 4$ mm hr^{-1} ; $B = 10$ mm² hr^{-1}). Grey line generated from a bucket model with $R_0=10$
710 mm; $p=0.40$.
- 711 4. Cumulative frequency distribution for November and December daily rainfalls 1997-
712 2002 at site Nogalte, North 2 (Murcia, SE Spain) fitted to Gamma distribution.
- 713 5. Relationship between annual runoff and sediment loss as vegetation cover is altered.
714 Loess soils, Holly Springs, MI. data from Meginnis, 1935.
- 715 6. Stabilisation of hydrology in the PESERA model from arbitrary initial conditions.
- 716 7. Stabilisation of natural vegetation cover in PESERA model from zero initial conditions
- 717 8. Final Pan-European estimates of soil erosion risk for current land use and climate.

718

719

720

721

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

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

723

724

725

726

727

727 **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	°C	BADC/MARS	Mean monthly temperature Corrected for altitude
mtrange_	2.4 – 18.4	°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_	-	°C	HADLEY ³	Predicted future temperature (scenario by month)
newrf130_	-	mm/mo	HADLEY	Predicted future rainfall (scenario by month)

728

729

730 *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	Dominant Arable Crop
maize_210c	-	-	ABM Survey	Maize Crop (default)
eu12crop2	-	-	(if applicable)	2 nd Dominant Arable Crop
itill_crop1	1-12	-	FSS/PDD	Planting month: Dominant Arable Crop
itill_maize	1-12	-	ABM Survey	Planting month: maize
itill_crop2	1-12	-	(if applicable)	Planting month: 2 nd Dominant Arable Crop
mitill_1	0/1	-	FSS/PDD	Planting marker: Dominant Arable Crop
mitill_m	0/1	-	ABM Survey	Planting marker: maize
mitill_2	0/1	-	(if applicable)	Planting marker: 2 nd 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

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

732

733

733 **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)

734

735 Grid ZM

736 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

737

738 Grid TEXT:

739 Description: Soil Texture

740 Source: Soil Geographical Database of Eurasia at scale 1:1,000,000 version
741 4.0 beta, European Soil Bureau, SAI/JRC Ispra.
742
743 0 No information
744 9 No mineral texture (Peat soils)
745 1 Coarse (18% < clay and > 65% sand)
746 2 Medium (18% < clay < 35% and >= 15% sand,
747 or 18% <clay and 15% < sand < 65%)
748 3 Medium fine (< 35% clay and < 15% sand)
749 4 Fine (35% < clay < 60%)
750 5 Very fine (clay > 60 %)
751

751

752 **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.

753

754

755

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

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

757

757

758

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

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