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1 **A conceptual model for physical and chemical soil profile evolution**

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4

5 **Abstract**

6 A simplified soil model is presented for evolution of the mineral soil profile. The model provides
7 a compromise between highly empirical models and highly mechanistic/ geochemical models,
8 and represents some of the main features of observed profiles, with features that can be identified
9 with 'A', 'B' and 'C' horizons. The model is responsive to a range of global environments,
10 which can be represented through climate and parent material parameters. In many cases there is
11 an almost single-valued relationship between surface weathering and soil depth, allowing further
12 simplification of the model, and allowing it to be included within a parsimonious landform
13 evolution models. A key parameter and assumption of the model is the degree of weathering
14 below which no further solution occurs, which limits the maximum extent of soil development.
15 This is speculatively linked to CO₂ turnover rates and the degree of aridity.

16 **Keywords**

17 Soil profile; modelling; bioturbation; weathering

18

19 **Introduction: proposed model framework**

20 Hydrological, biological, chemical and mechanical processes take place throughout the critical
21 zone (Figure 1), interacting at a wide range of temporal scales, and together driving evolution of
22 the soil / regolith profile. Over time, weathering introduces material into the base of the profile,
23 while geomorphic processes transport material downslope, progressively eroding the surface of
24 the land. These processes interact, slower weathering, for example, producing coarser material
25 that is eroded more slowly. Rates of production and removal may be out of balance for millions
26 of years, leading to either indefinite accumulation of soil or stripping to parent material, but in
27 many cases there is an approach to a quasi-equilibrium with a finite depth of soil.

28 In this paper, processes of soil formation have been simplified to provide a tentative theoretical
29 framework, providing a conceptual model of soil profile evolution. This builds on, and expands,
30 prior work (Carson and Kirkby, 1972; Kirkby 1977, 1985a), and has been informed by the many
31 and diverse soil evolution models in the literature (e.g. Minasny & McBratney 1999, 2001; Finke
32 & Hutson 2008; Gabet & Mudd, 2009; Opolot & Finke, 2015). A number of simplifying
33 assumptions have been made to keep the model relatively simple, particularly with respect to the
34 geochemical evolution of the soil, although some can, in principle, be relaxed. A high priority
35 has been to include as many significant interactions between the processes acting within the
36 profile as possible, and particularly those between the evolving profile and soil hydrology. The

37 necessary resulting simplifications have, hopefully, provided some gains in understanding but
38 also consequential sacrifices in rigour.

39

40 Early concepts of soil development express the evolving balance of physical and chemical
41 denudation (Carson & Kirkby, 1972, p 265), leading either to an equilibrium in which the degree
42 of weathering at the surface is a function of the balance between the two forms of denudation, or
43 to indefinite deepening of the soil profile. The increase in soil depth is given by the difference
44 between the rate of weathering and the rate of erosion, corrected for the degree of weathering of
45 the surface material. The rudimentary model proposed by Minasny and McBratney (1996,
46 2001) takes account of a depth-dependent weathering rate and a surface-controlled erosion rate
47 to develop the profile model, additionally taking account of changes in bulk density as
48 weathering proceeds. Brimhall & Dietrich (1987) combine this approach with a much more
49 sophisticated geochemical analysis to show how recalcitrant soil components (Fe and Al) may
50 accumulate as hardpans or laterites as silica is preferentially removed due to its greater solubility.

51 Hoosbeek & Bryant (1994), Lebedeva et al(2010) and Li et al (2014) make use of similar
52 advection-dispersion equations for the movement of solutes to those used by Kirkby (1985a) and
53 van Genuchten & Wierenga (1986), and estimate solute concentrations by appealing to chemical
54 equilibrium between water and solid mineral phases.

55 Willgoose and his group (Cohen et al, 2010; Welivitya et al, 2016) have focussed on the physical
56 breakdown of material into progressively finer fractions, and the accumulation of an armour
57 layer at the surface as water erosion winnows fines to progressively concentrate the coarser
58 material. In the mARM3D model alternative depth-dependent weathering functions are
59 combined with the comminution model to provide a valuable model for critical zones for which
60 physical breakdown is the dominant process. These models have also been coupled with
61 landscape evolution models to generate downslope catenas as the landform as a whole evolves.

62 Other soil profile models has made a much more detailed analysis of chemical processes,
63 generally using kinetic equations to describe rates of solution and solution products for different
64 constituent minerals. Examples of such models are the WITCH model (Godd ris et al, 2006) and
65 the SoilGen model (Opolot & Finke, 2015). These models also explicitly incorporate carbon
66 cycling to provide relevant levels of CO₂ partial pressure in the soil, linking to RothC or
67 ASPECTS to provide carbon cycling and outgassing. They also include some hydrology,
68 allowing them to respond to the external environment. Due to their relative complexity,
69 however, they are more difficult to couple with landscape evolution.

70

71 Vanwallegem et al (2013) have created a model explicitly for coupling with landscape
72 evolution (MILESD), dividing the soil profile and parent material into four layers to model
73 movement of solutes and clay between layers, bounded by depth-dependent weathering of parent
74 material and diffusive sediment transport at the surface. Carbon cycling and bioturbation are
75 also significantly incorporated in exchanges between the upper layers. This approach represents

76 an advance on previous work through integration of hydrology and weathering into a simple and
77 unified model structure that is compatible with landscape evolution models.

78

79 Because of the complexity of the processes operating in soils, and their interaction with surface
80 sediment processes, all models represent compromises, emphasising some aspects and over-
81 simplifying others. The present proposal provides no resolution from this dilemma, but offers an
82 alternative approach which has some strengths, and allows further exploration of some internal
83 linkages. Nevertheless it is recognised that carbon cycling is currently not included, and that
84 changes in bulk have been assumed to be negligible, with loss of substance balanced by a
85 corresponding reduction in density, an assumption that is, in some contexts, demonstrably false.

86 Here we make a number of major assumptions in order to simplify and generalise some of the
87 processes involved in soil profile development. The first, and most important of these, is to
88 combine all chemical constituents into a single term which expresses the degree of weathering,
89 and is defined below. It is implicit in this assumption that weathering is a largely congruent
90 process, so that, for a given parent material and physical environment there is an almost one-to-
91 one relationship between degree of weathering and chemical composition, so that, for example,
92 there is a single-valued relationship between degree of weathering and solute concentration.
93 This relationship is analogous to the commonly used approximation that, at a site, there is a
94 single-valued relationship between soil moisture storage and flow rate.

95 The second major assumption is that modifications in the rates of other processes can also be
96 simply related to degree of weathering. This dependence has been applied implicitly to grain size
97 distribution, and explicitly to the rate of surface sediment erosion (or as one control on the rate of
98 sediment transport in a landscape context). The functional form of these relationships has usually
99 been expressed here as a power law, with exponents reflecting qualitative rather than quantitative
100 forms. The third assumption made is that solute concentrations in soil water are, on the time
101 scales of profile development, in chemical equilibrium with the solid phase, so that
102 concentrations are the product of a solubility that is the mass-weighted concentration of mineral
103 solubilities in the weathered solid phase at a given degree of weathering, again expressible as a
104 single valued function of the current local degree of weathering. This assumption reflects
105 experimental work on silicates (Garrels & Christ, 1965; Bricker et al, 1968; Robinson & Stokes
106 1959), suggesting that equilibration occurs over periods of about 100 hours for silica, and
107 somewhat longer for alumina: all periods that are short compared to the time scale for profile
108 evolution. The fourth major assumption is that there is no change in volume as the soil weathers.
109 This assumption has been used in a number of previous models (e.g. Lebedeva 2010) and is
110 visibly supported by the presence of intact structures within saprolite, for at least some parent
111 materials, but is not universally applicable, for example in limestones and podsoles. The fifth
112 major assumption is that, for a given environment and parent material, there is a maximum
113 degree of weathering beyond which no further soil development occurs. This assumption is
114 speculative, but is essential to the model. It is discussed briefly in the conclusions below.

115 Ignoring, initially, the organic horizons, a simple way to characterize the changing properties of
 116 the profile is by referring to the total loss of substance at any level, based on an analysis of all
 117 constituent minerals. If a hypothetical constituent that is assumed to be totally insoluble is
 118 increased in abundance by k times at some level in the profile relative to its abundance in the
 119 parent material, then the proportion, p , of rock substance at that level $1/k$. In many cases,
 120 simple soil profiles can then be generalised as a sigmoid curve, with the proportion remaining,
 121 p , declining from 1.0 at depth in parent material, and falling to progressively lower values
 122 towards the surface as in the definition sketch of figure 2. In the absence of a non-arbitrary
 123 lower boundary, a generalised 'soil deficit', ζ , can also usefully be defined as the total depth of
 124 material that would have to be replaced to restore the soil to its original, parent material,
 125 condition. This can be written:

$$126 \quad \zeta = \int (1-p).dz \quad (1)$$

127 where the integral is evaluated from the surface downwards, and ζ has the dimension of depth..
 128 This generalisation is necessary since there is not a sharp boundary between soil and parent
 129 material. Where there is such a sharp boundary, with weathering at p^* above, then soil depth, z
 130 $= \zeta / (1-p^*)$.

131
 132 A key variable is the value of p at the surface, p_s . Since, to a first approximation, mechanical loss
 133 takes place at the soil surface, a mass balance for the soil profile is:

$$134 \quad \frac{d\zeta}{dt} = C - M \cdot \frac{p_s}{1-p_s} \quad (2)$$

135 If there is an equilibrium soil profile developed under constant rates of mechanical (M) and
 136 chemical (C) denudation,

$$137 \quad p_s = M / (M + C) \quad (3)$$

138 the state of weathering is also commonly summarised in the 'Chemical Depletion Ratio',

$$139 \quad \text{CDR} = 1 - p_s = C / (M + C) \text{ at soil equilibrium.} \quad (4)$$

140 These relationships are fundamental to almost all soil profile models.

141
 142 In this paper the processes of soil evolution are modelled in a simplified way, primarily in terms
 143 of the one dimensional profile of substance remaining, as sketched in figure 2. It is recognised,
 144 and here assumed, that, for different parent materials, the composition of the soil at different
 145 stages of weathering reflects both its mineral assemblage and its grain size composition, and that
 146 the concentration of solutes in soil water shows a different, but consistent, dependence on
 147 substance remaining at various stages.

148
 149
 150 A first, strongly simplifying, approximation to the course of weathering can be obtained by
 151 analysing rock minerals as a mixture of oxides, and assigning a constant solubility to each,
 152 without allowing for interactions as mineral composition changes. Over time, the most soluble
 153 components of the rock dissolve most rapidly, so that the composition of the weathered material
 154 concentrates the less soluble components, such as Iron and Aluminium oxides, and the average
 155 solubility of the remainder is reduced to reflect this change in composition. Table 1 and figure 3
 156 illustrate this dependence for an idealised igneous rock. This approach, although ignoring the
 157 successive mineral assemblages as weathering proceeds, mirrors the overall course of weathering
 158 sufficiently for the present model, but is included here only as illustrating the overall direction
 159 of the weathering sequence

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For the present model, figure 3(c) illustrates the decrease in solubility as weathering proceeds. This will be represented in the model as a decrease with proportion remaining, falling to zero at a minimum proportion, p_0 , representing the ultimate final achievable state of weathering. The proposed model is intended as one of reduced complexity, allowing readier linkage to landscape evolution models and, inevitably, greatly simplifying the geochemical relationships, which, for the purposes of the model, are summarised entirely through the assumed relationship between solute concentration and proportion remaining, encapsulated in equation (6) below or an alternative

171

172 **Hydrological and Chemical Process representation**

173 Figure 1 indicates the range of processes that have been incorporated into the model, which is
174 driven by these processes within a mass balance framework that accounts for both water and soil
175 substance, generally working implicitly with annual rather than seasonal or storm-based totals,
176 although the time step varies to maintain computational stability. The hydrology is driven
177 externally by precipitation and actual evapotranspiration ($ET = E_A$), the latter derived from
178 potential ET and precipitation using a Budyko transformation (Arora, 2002). ET is assumed to
179 decay exponentially with depth, based on an assumed average root depth z_E , giving a total ET
180 demand below depth z of $E_A \exp(-z/z_E)$. In the present form of the model, ET is assumed to
181 consist of water free of solutes, but may be modified to allow for the solute demands of the
182 vegetation cover. Progressive weathering is currently assumed to take place without change of
183 volume, so that soil porosity increases as $(1-p)$, although with the facility to allow some
184 percolation into intact parent material. At any depth z , the potential for lateral subsurface flow
185 $q(z)$ is related to the porosity through the relationship $q(z) \sim (1-p)^2$ (Terzaghi & Peck 1967).
186 Actual downward percolation is then estimated as the lesser of the two quantities, each with the
187 dimension of mm.a^{-1} [LT^{-1}].

188 Rainfall R depleted by accumulated ET above this depth: $=R - E_A(1 - \exp(-z/z_E))$ (5a)

189 Potential for lateral flow together with ET demand below level z : $= [q(z) + E_A \exp(-z/z_E)]$ (5b)

190 The contribution to lateral sub-surface flow from within the soil profile is then zero if the first
191 expression (equation 5a) is the lesser (commonly near the surface), and is $q(z)$ if the latter
192 expression is less. This condition occurs below the saturated level for a humid climate, but not
193 normally under arid conditions, and increases with hillslope profile convexity. This contribution
194 from the individual soil profile to lateral subsurface flow (down the hillslope) is assumed to
195 carry away solutes at their equilibrium concentration within the soil water. With increasing depth
196 and associated reduction in weathering, losses by advection will also become progressively
197 smaller. The choice of this hydrological sub-model allows the soil hydrology to change
198 dynamically in response to climate and the progressive evolution of the soil profile, and it is
199 representation of this interaction that provides an important driver of the proposed model. It is
200 implicit in this assumption that the soil profile considered is identical to those above and below

201 it, an assumption that is most clooely met for a uniform convex hilltop. For other contexts, the
202 present one-dimensional analysis should be nested within an analysis that represents the entire
203 hillslope catena. These issues have been discussed, though not fully resolved, by Willgoose
204 (2018).

205 Soil water is assumed, in the model, to come to chemical equilibrium with the constituents of the
206 solid phase of weathered material, and to do so over a time span that is short relative to the rate
207 of evolution of the soil profile. The local concentration in the soil water is then the product of the
208 solubility of the weathered material and the proportion of substance remaining, both estimated at
209 the level of interest within the profile. In the model, the assumed relationship between
210 concentration and proportion present, $c(p)$, has been given the form:

$$211 \quad c(p) = c_* \left(\frac{p-p_0}{1-p_0} \right)^{0.5} \quad (6)$$

212 Where c_* is the solute concentration in parent material, and p_0 is the ultimate state beyond which
213 weathering cannot proceed. This relationship is taken to incorporate dependence on pH and
214 $p\text{CO}_2$ in the context of the weathering sequence for a particular parent material. It is noted that
215 this relationship differs appreciable from the curve shown in figure 3c, with a much more abrupt
216 cut-off as p_0 is approached. This difference reflects the observation that observed deep soil
217 profiles appear to show the persistence of a lower limit for weathering, which may take values of
218 0.4 for laterites (e.g. Hamdan & Burnham, 1996), but much higher values of around 0.8 in Sierra
219 Nevada (CA) granites (Riebe et al, 2003). This lower limit appears to be a strong control in
220 practice, although one that is unexplained by the simple linear theory shown in figure 3 and
221 Table 1. This issue is addressed further below. The form of equation 6 with an exponent of 0.5
222 has been chosen to provide a sharp decrease in solubility as p_0 is approached, but is, to some
223 extent, arbitrary.

224

225 Two key processes are represented in the model to allow exchange and removal of solutes,
226 namely diffusion and advection. Diffusion occurs in two important ways. First ionic diffusion
227 dominates the region close to parent material, where there is little exchange of water, so that
228 differences in solute concentration drive diffusion of ions, leading to a net migration away from
229 parent material into the body of the soil. Diffusion coefficients for different ions range over
230 about one order of magnitude. In free water, values are typically in the range $200\text{-}2,000 \text{ cm}^2.\text{a}^{-1}$
231 (Robinson and Stokes, 1959), but allowing for tortuous flow in soil pores, effective values may
232 be two orders of magnitude less. Here the value of $100 \text{ cm}^2.\text{a}^{-1}$ is adopted for all ions. Near a
233 tight parent material, where there is no significant downward percolation (and solute advection)
234 through the rock, ionic diffusion is the dominant solute transfer process acting. In isolation this
235 will be seen to give a soil profile of the form:

$$236 \quad p=1-\exp(-z/z_0) \quad (7)$$

237 for a suitable scaling constant z_0 . The rates on soil development produced by ionic diffusion are
238 rather slow, since movement is only of the ions, which are always in relatively low (1-1000
239 mg.l⁻¹) concentrations within the soil water.

240 Diffusion is also active near the soil surface, where bioturbation, acting through frost-heave,
241 earthworm burrows, tree-throw etc, drives random exchanges that turn over the entire mineral
242 soil and organic matter (Gabet et al, 2003). This bio-diffusion is responsible for creating a bulk
243 density and porosity profile near the soil surface, and for driving the seasonal soil creep that is
244 responsible for many hilltop convexities (Gilbert 1909). In both cases, net outward diffusion
245 towards the surface is balanced by settlement under gravity. Measured diffusion rates (Kirkby,
246 1967; Johnson et al, 2014) of 1-100 cm².a⁻¹ are comparable to those for ionic diffusion, but are
247 responsible for moving much greater (10³-10⁶ times) volumes of material, since the entire soil
248 contents are being exchanged. Since this process is, by far, the most rapid in the context of
249 evolving mineral soil, its effect is to produce a zone of thorough mixing and homogenisation of
250 the mineral soil. However these rates of exchange are comparable to rates of organic matter
251 production and decomposition, so that the near-surface zone retains strong horizonation of
252 organic soil constituents.

253 Advection is driven by flowing water that is carrying solutes. Solutes are dissolved in the water,
254 and are assumed to come to chemical equilibrium with the solid phase composition in times short
255 compared to those required for significant soil evolution. Experiments suggest that the relevant
256 period for substantial equilibration is a few days (Bricker et al, 1968), although some reactions
257 are much slower than this. Only for overland flow runoff may reaction times be significantly
258 shorter, perhaps reducing weathering rates at the surface (Gabet & Mudd, 2009): this effect is
259 neglected here. Waters percolates downwards at a decreasing rate as porosity and hydraulic
260 conductivity decline, is diverted laterally as a contribution to downslope subsurface flow, and is
261 lost to evapotranspiration. Downward and lateral flow contributions are assumed to advect water
262 at the local solute concentration, while ET is assumed to respond to plant requirements, and is
263 here assumed to be free of solutes. The diversion of clean water for evapotranspiration can have
264 the effect of increasing concentration of the remaining dissolved material in the remaining
265 water, but this process is limited as concentrations reach saturation, beyond which further solutes
266 are precipitated back into the solid phase.

267 A Steady State Solution

268 As an introduction to the dynamics of the model system, a straightforward analytical solution can
269 be obtained if the following assumptions are made: no bioturbation; constant solute
270 concentration (replacing equation 5 above); a net downward water flux proportional to $(1-p)$
271 representing the increase in permeability as weathering proceeds; and a constant rate of material
272 loss at the surface through erosion.

273 With these assumptions, the mass balance equation for the soil profile is:

$$274 \quad \frac{\partial p}{\partial t} = \frac{\partial}{\partial z} \left[Qpc - \frac{\partial}{\partial z} (pcD_I) \right] \quad (8)$$

275 where both sides of the equation have units of $[T^{-1}]$

276 z is depth below the surface (mm),

277 Q is the downward water flux (mm.a⁻¹),

278 and D_I is the ionic diffusion rate (mm².a⁻¹)

279 Assuming that the downward water flux is $R(1-p)$, where R is net rainfall, and that, in the steady
280 state, the frame of reference is moving at rate v , then, setting

281 $p_0 = v/(Rc)$, $z_0 = D_I/[R(1-p_0)]$ and $\alpha = (1-p_s)/(p_s-p_0)$, the steady state solution is:

$$282 \quad p = \frac{1+p_0\alpha\exp(-\frac{z}{z_0})}{1+\alpha\exp(-\frac{z}{z_0})} \quad (9)$$

283 which is a sigmoid curve for the dimensionless proportion p , with the general form shown in
284 figure 2, for which $p \rightarrow p_0$ as $z \rightarrow -\infty$; $p = p_s$ at $z = 0$ and $p \rightarrow 1$ as $z \rightarrow +\infty$. An example of the
285 associated fluxes and gains/losses is shown in figure 4. If bioturbation is introduced into this
286 solution, there is a region of almost constant mineral composition that extends throughout the
287 depth of biological mixing, but very little flux within this region because any slight perturbation
288 is quickly removed by the rapidity of diffusion.

289

290 Computer simulation model

291 In order to address non-equilibrium conditions, Equation (8) should be modified to include
292 bioturbation effects, and an additional equation is required for the hydrological mass balance.

293 Equation (8) then becomes:

$$294 \quad \frac{\partial p}{\partial t} = \frac{\partial}{\partial z} \left[Qpc - \frac{\partial}{\partial z} (pcD_I + pD_B) \right] \quad (10)$$

295

297 where D_B is the rate of diffusion by bioturbation..

298 Water balance down the profile is given by

$$299 \quad -\frac{\partial Q_v}{\partial z} = E(z) + q(z) \quad (11)$$

300 where Q_v is the vertical downward water flux, carrying solutes at concentration $c(z)$

301 $E(z)$ is the lateral ET flux at depth z , carrying only solutes to meet plant requirements

302 $q(z)$ is the contribution, across the soil profile to lateral subsurface flow, carrying solutes
303 at $c(z)$. This is determined by equation (5) above. It is generally zero near the surface, and increases with
304 hillslope profile convexity and with increased weathering.

305

306

307

308 The concentration c at any level now follows equations (10) and (11) rather than remaining
309 constant, except that the ET stream is here assumed to be of pure water. A second diffusion term
310 has been added to equation (10), representing bioturbation of the entire soil mass near the
311 surface. The Biological diffusion and ET are each assumed to decay exponentially with depth,
312 each with its own scale depth that is treated as a model input. Finally the rate of mechanical
313 removal for weathered soil, M_0 , is taken as an input variable that is considered to reflect the
314 landscape morphology in which the modelled soil profile is set. However, it is recognised that
315 removal rate also depends on the state of soil weathering, tending to zero for intact unweathered
316 parent material, with a coupling observed by Riebe et al (2003) and modelled by Gabet & Mudd
317 (2009). Here it is recognised that mechanical removal rates decrease sharply for almost
318 unweathered soils, and this qualitative observation is expressed quantitatively in the present
319 model in the assumed form:

$$320 \quad M = M_0(1 - p_s)^{0.1} \quad (12)$$

321 where M_0 is the mechanical erosion on fully weathered material ($p_s=0$)

322 This expression suppresses removal to 80% for $p_s=0.9$, and to 60% for $p_s=0.99$, so that it has little
323 effect except in the initial evolution of parent material surfaces. This approach implicitly takes
324 some account of the physical breakdown of material, the dominant component of the Cohen et al
325 (2009) mARM model applied to weathering on mine waste rock. The observed coupling also
326 requires that chemical weathering rates are reduced as the soil thickness increases, as
327 documented by Heimsath et al (1997) through cosmogenic dating. This relationship has not been
328 included in the model as an explicit assumption, but is seen as an output of the model, reflecting
329 the lesser circulation of water as soil deepens. Table 2 summarises the parameters required to
330 define and run the model, indicating the ranges of values that have been applied.

331 Figure 5 shows an example model output for a humid region. It shows the development of a
 332 near-surface region of uniform concentration, dominated by bioturbation, and broadly
 333 corresponding with the ‘A’ horizon of the soil. Below, a weathering front develops, in which the
 334 proportion remaining increases rapidly with depth, leading to a region where there may be
 335 accumulation of material, as a result of concentration by loss of ET. This region, broadly
 336 corresponding to a ‘B’ horizon, becomes pronounced, as in the example shown, when the scale
 337 depth for ET is greater than the scale depth for bioturbation. Since biological activity influences
 338 both of these distributions, it is expected that they will be broadly similar. Below this zone of
 339 potential enrichment, ionic diffusion dominates to produce the ‘C’ horizon where weathering is
 340 gradually penetrating into parent material. For the same model run, figure 6 shows the
 341 relationship between solution rate and soil deficit (broadly equivalent to depth), an exponential
 342 decline similar to that measured by Heimsath et al (1997), in this case with a scale depth of
 343 approximately 80mm.

344 For a range of climates, defined by their precipitation and potential ET, figure 7 shows how the
 345 modelled soil depth evolves over time. The values on the right show the ratio of precipitation to
 346 Pot ET, which seems to control the overall course of evolution. In all cases the soil deficit
 347 develops at an accelerating rate initially, and finally settles down to an equilibrium value as the
 348 weathering rate falls to the imposed mechanical rate of denudation. These model runs suggest
 349 that, for the most arid climates, the evolution will be towards steady accumulation,
 350 corresponding, for example, to a soil with strong calcrete development.

351 Figure 8 illustrates how the modelled evolution converges towards its equilibrium state for a
 352 range of initial conditions that differ only in the initial rate of mechanical denudation, M_0 . This
 353 corresponds to a set of slopes of different gradients, or to different points along the same slope
 354 profile. For high initial mechanical denudation rates, there is little chemical evolution, whereas,
 355 where mechanical denudation is low, there is substantial weathering, and profiles evolve
 356 asymptotically to a state where the state of surface weathering, p_s , is close to the final state of p_0 .
 357 This curve is compatible with field data such as that presented by Riebe et al (2003), if it is
 358 assumed that weathering rate declines exponentially with soil depth.

359

360 For a set of solutions within the same weathering environment, but with different mechanical
 361 settings, such as those shown in figure 8, there appears to be a more or less single valued
 362 relationship between soil deficit, ζ and the proportion remaining at the surface, p_s . This is shown
 363 for this set of runs in figure 9. An empirical fit to the curve shown is

$$364 \quad x = \exp\left\{A \left[-3 + 3 \exp\left(-\frac{\zeta}{4.3}\right) - \frac{\zeta}{4.3} \right]\right\} \quad (13)$$

365 for constant A and where $x = (p_s - p_0)/(1 - p_0)$. This differs from the equivalent relationship found
 366 for the steady state solution of equation 9, which contains only the final term of equation 13.

367 The model has been set up here as one-dimensional, representing a single soil profile. The
 368 mass balance equations can be written in the forms

369 Total denudation: $\frac{dy}{dt} = -(C + M)$ (14)

370 Solute denudation: $\frac{d\zeta}{dt} = C - M \frac{1-p_s}{p_s}$. (15)

371 Where yz is the elevation of the soil surface.

372 With suitable expressions for the chemical and mechanical denudation rates, these equation can
 373 be solved, either analytically or computationally, but to do so requires an additional relationship,
 374 which may be considered as a ‘rating curve’ that connects soil deficit to the degree of surface
 375 weathering. Figure 9 and equation 11 suggest that this is a workable approach to provide a much
 376 simpler computational model, an example of which is shown in figure 10.

377 The potential value of this approach may be seen to be much greater if the soil model is
 378 embedded in a two- or three- dimensional landscape evolution model, with feedback between
 379 soil and landform evolution as sketched in figure 11. In this case the corresponding mass balance
 380 equations are:

381 Total material transport: $\frac{\partial y}{\partial t} = -\nabla \cdot (\mathbf{V} + \mathbf{S})$ (16)

382 Solute transport: $\frac{\partial \zeta}{\partial t} = -\nabla \cdot \left(\mathbf{V} - \mathbf{S} \frac{1-p_s}{p_s} \right)$ (17)

383 where \mathbf{V} and \mathbf{S} are respectively the vector solute and sediment transport at each point in the
 384 landscape and ∇ is the vector divergence operator. Similar equations must also be applied for
 385 the water balances. As for the 1-D case, a relationship of the form $\zeta = f(p_s)$ can be used to close
 386 the set of equations, subject to boundary & initial conditions. The viability of this approach,
 387 with a simpler soil model, has previously been demonstrated (Kirkby, 1985b).

388
 389 Detailed behaviour in the bioturbation layers is not pursued here, but the homogenization of the
 390 mineral soil is accompanied by much greater changes in the soil organic matter, creating profiles
 391 of organic matter concentration and porosity/ bulk density. Outward diffusion by bioturbation is
 392 balanced by downward settlement under gravity. Organic matter, generated near the surface
 393 through leaf fall, root decay etc, decomposes and is mixed, predominantly downwards. The
 394 surface may also be subject to erosional stripping or the deposition of wind-blown material,
 395 providing near-surface interactions that are critical to ‘A’ horizon development but that may
 396 initially be ignored in the present context of mineral profile evolution.

397

398 **p_0 , the final product of weathering**

399

400 It has been assumed above that weathering cannot, in a particular environment, proceed past a
 401 certain state, and this state has been represented in the model as an input parameter, p_0 . It is
 402 widely observed that there is a global pattern of soil types, and there is a corresponding
 403 distribution for, p_0 , the end point of mineral weathering. Any tentative suggestions here also
 404 assume constant climatic environments, so that they can best reflect conditions in areas where
 405 these have been stable for long enough to approach equilibrium soil conditions. It is tentatively
 406 suggested that the end point is associated with both aridity and soil CO₂ production. Aridity is
 407 closely linked to the ratio of potential ET to precipitation. High aridity is associated with
 408 accumulation of duricrusts, particularly calcrete, within the soil horizon, and so with high values
 409 of p_0 . CO₂ production is associated with high rates of actual evapo-transpiration, and is
 410 associated with high rates of organic complexing with Fe and Al (Hernández-Soriano, 2012),
 411 which may limit their further dissolution, setting the value of p_0 in humid tropical

412 environments. Figure 12 sketches these regions on a diagram of annual precipitation and
413 temperature, relating them to the index

$$414 \quad \log(AE) + \log(r/R) \quad (18)$$

415 where AE is estimated annual actual evapotranspiration, and r/R is the estimated ratio of runoff
416 to precipitation. This diagram conforms with observed patterns, with laterites ($p_0 \sim 0.4$) in
417 humid tropical areas, and the least weathered soils in semi-arid regions. It is acknowledged that
418 these suggestions are highly speculative, with the mechanisms for such a relationship still
419 unclear, and that there is scope for much further work on this topic.

420 **Conclusions**

421 It has been shown that a relatively simple model is able to capture many of the key essentials of
422 soil profile development over time, and in response to a wide range of climatic environments. It is
423 acknowledged there is scope for much greater detail that involves specific mineralogy and
424 weathering paths. It is also recognised that even the generalised path of weathering shows strong
425 dependence on acidity (Ph), partial pressure of CO_2 ($p\text{CO}_2$) and reduction potential (Eh).
426 Nevertheless the simple model is able to show realistic dependence on broad lithological and
427 climatic controls in a way that has the potential to provide a broad-brush picture of global soil
428 patterns in time and space. The relative simplicity also allows incorporation into landscape
429 evolution models for 2-D or 3-D landscapes that retain soil profile evolution and dynamics in a
430 meaningful, if simplified way.

431 It has been noted that the modelling requires defining a lower limit of weathering for an
432 environment, and that there appears to be no satisfactory explanation for this phenomenon. There
433 is weak evidence that this lower limit is connected with the interaction of mineral soils with
434 organic soil components, but no clear mechanism.

435

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Component oxide	Al	Si	Ca	Mg	Na	K	Fe	Ti	Total (%)	Solubility (mg/l)
Initial p	14.0	49.0	7.0	13.0	1.7	0.3	13.0	2.0	100	20.3
solubility (mg/l)	6	15	80	40	60	0.2	2	0		
Relative time										
0	14.0	49.0	7.0	13.0	1.7	0.3	13.0	2.0	100	20.3
10	13.2	42.2	3.1	8.7	0.9	0.3	12.7	2.0	83	13.9
20	12.4	36.3	1.4	5.8	0.5	0.3	12.5	2.0	71	10.2
30	11.7	31.2	0.6	3.9	0.3	0.3	12.2	2.0	62	7.9
40	11.0	26.9	0.3	2.6	0.2	0.3	12.0	2.0	55	6.3
50	10.4	23.1	0.1	1.8	0.1	0.3	11.8	2.0	50	5.2
60	9.8	19.9	0.1	1.2	0.0	0.3	11.5	2.0	45	4.4
70	9.2	17.1	0.0	0.8	0.0	0.3	11.3	2.0	41	3.7
80	8.7	14.8	0.0	0.5	0.0	0.3	11.1	2.0	37	3.2
90	8.2	12.7	0.0	0.4	0.0	0.3	10.9	2.0	34	2.8
100	7.7	10.9	0.0	0.2	0.0	0.3	10.6	2.0	32	2.4

520

521 **Table 1: Illustration of progressive linear solution of rock material. See also figure 3.**

522

Parameter description	Range of values	Units	Symbol	Comment
Solute concentration at p=1	1-100	mg. l ⁻¹	c*	c* in equation (5)
Solute exponent	0.5			Exponent in equation (5)
Precipitation	100-2000	mm. a ⁻¹	P	
Potential evapotranspiration	200-2000	mm. a ⁻¹	E _P	Actual ET calculated as $1/\sqrt{(1/P^2+1/E_P^2)}$
Depth of ET penetration	20-200	mm		Depth in soil
Mechanical denudation at p=0	0.1-100	mm/ Ka ⁻¹	M ₀	In equation (10)
Exponent for mechanical denudation	0.1			Exponent in equation (10)
Capacity for lateral subsurface flow increment at p=0	50-1000	mm. a ⁻¹		Should be proportional to slope profile convexity
Lower limit of weathering	0.4-0.8		p ₀	Proportion remaining can never be less than this limit value
Reduction in solute removal at surface	1			Can allow for reduced solute loss in overland flow
Ionic diffusion rate	1-100	cm ² . a ⁻¹	K _I	Dominant near parent material
Bioturbation diffusion rate	1-100	cm ² . a ⁻¹	K _B	Dominant near surface
Depth of bioturbation penetration	20-200	mm		Depth in soil
Additional percolation capacity for parent material	0	mm. a ⁻¹		Not operational

523 **Table 2: Example Model parameters**

525 **Figure captions**

- 526 1. Hydrological, chemical, biological and mechanical processes within the critical zone
527 2. Definition sketch: a generalised soil profile.
528 3. Linear solution example.
529 a. Initial assumed composition
530 b. Composition after weathering to $p=0.4$
531 c. Relationship between average solubility and proportion remaining
532 d. Proportion remaining for three components over time
533 4. Components of soil development for steady state case of equation 9, with no bioturbation
534 a. Shows fluxes due to ionic diffusion and advection
535 b. Shows net gains or losses. The region of greatest change corresponds to the migrating
536 weathering front in which p is changing most rapidly.
537 5. Example of soil profile evolution in the model.
538
539 Modelled decline in solution rate with soil deficit (\sim depth) over 300Ka. Values taken from
540 model run shown in figure 5.
541 6. Evolution of soil deficit over time for a range of climates from humid to arid. The legend
542 shows the precipitation and potential ET (in mm per year) for each model run. Values in
543 the legend are the ratios of precipitation to Pot ET.
544 7. Model runs that differ only in the initial rate of potential mechanical denudation. The upper
545 curves show the modelled evolution of mechanical and chemical denudation. The lower
546 curve shows the final equilibrium states (not reached in some cases). In which
547 $p_s=M/(M+C)$.
548 8. The almost single-valued relationship between soil deficit (or generalised soil depth) and
549 the degree of weathering at the surface, $\zeta =f(p_s)$, for the set of model runs shown in figure
550 8.
551 9. Simplified soil profile evolution model, using equations 6, 10-12. Note that total denudation
552 greatly exceeds soil deficit.
553 10. Embedding soil profile model within a landscape evolution model
554 11. Tentative identification of climatic regions with the composite variable $\log(E_A)+ \log(r/R)$ as
555 an indicator of regional final state of weathering, p_0 . E_A is actual evapotranspiration, r/R is
556 the ratio of runoff to precipitation, the reciprocal of aridity. The axes of the plot are for
557 annual precipitation and mean annual temperature. Values on the graph indicate suggested
558 values for p_0 .

559

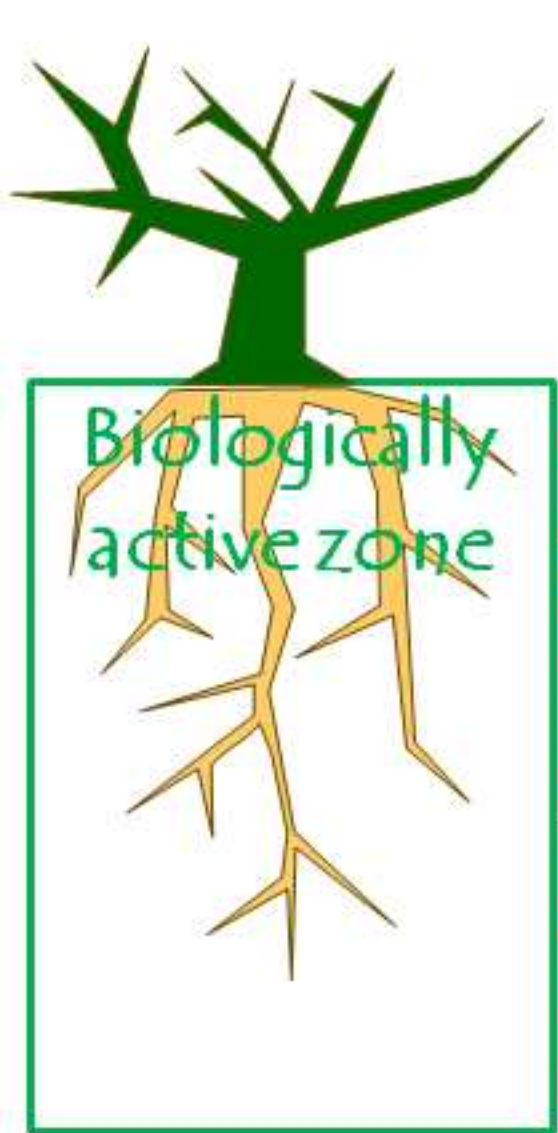
560

561

562

563

564



Evapotranspiration

Hydrology
Precipitation

Infiltration
Lateral flow

Root uptake

Minimal
flow

+/- Static
Saturated

Advection of water & solutes.....throughout

Chemistry
(Rock
constituents)
Leaf-fall

Selective root
uptake

Declining
activity

Ionic diffusion

Biology
(SOM and pCO₂
profiles)

Leaf-fall

Root-fall

Bioturbation

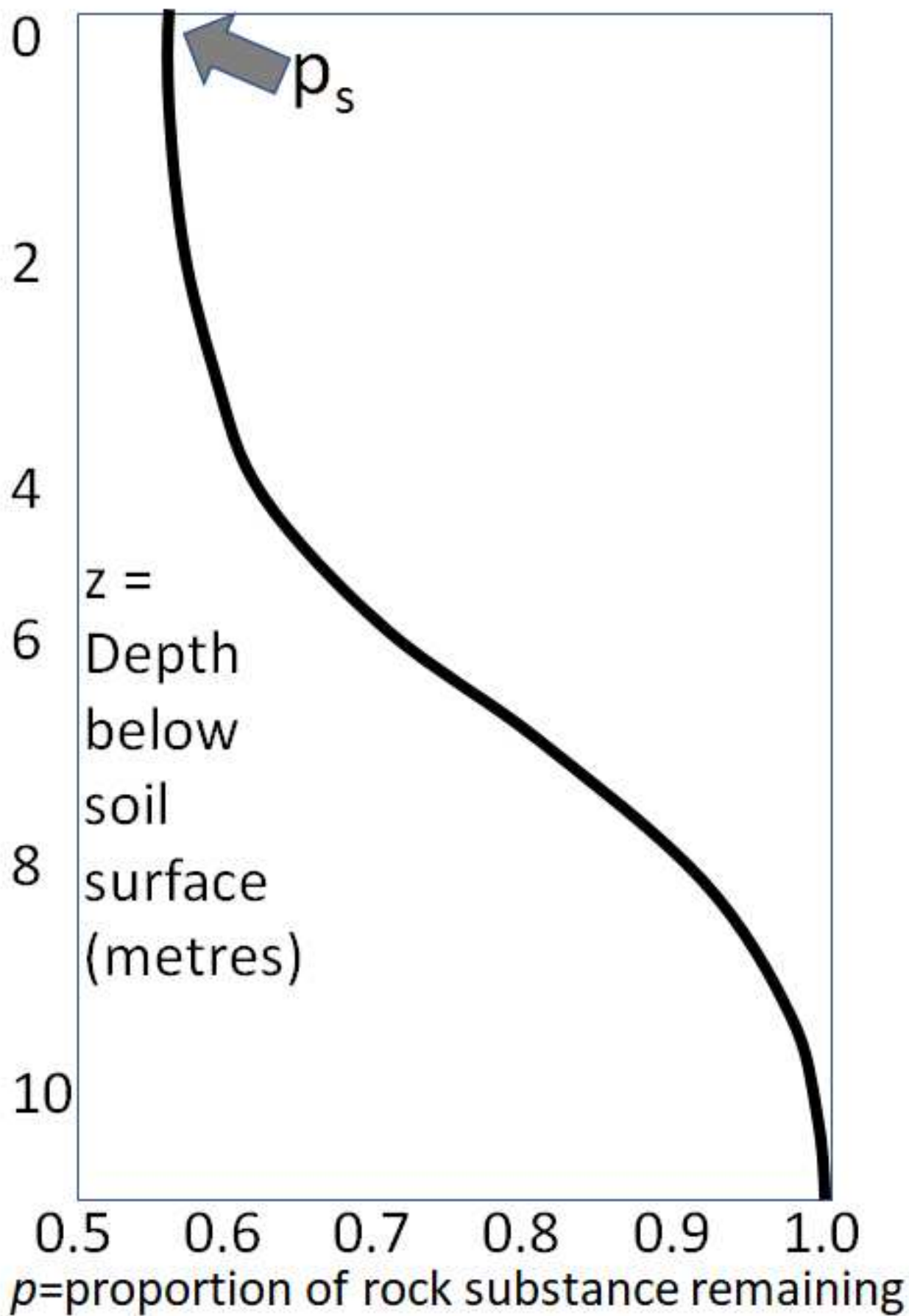
Mechanical

Frost heave/
wetting drying

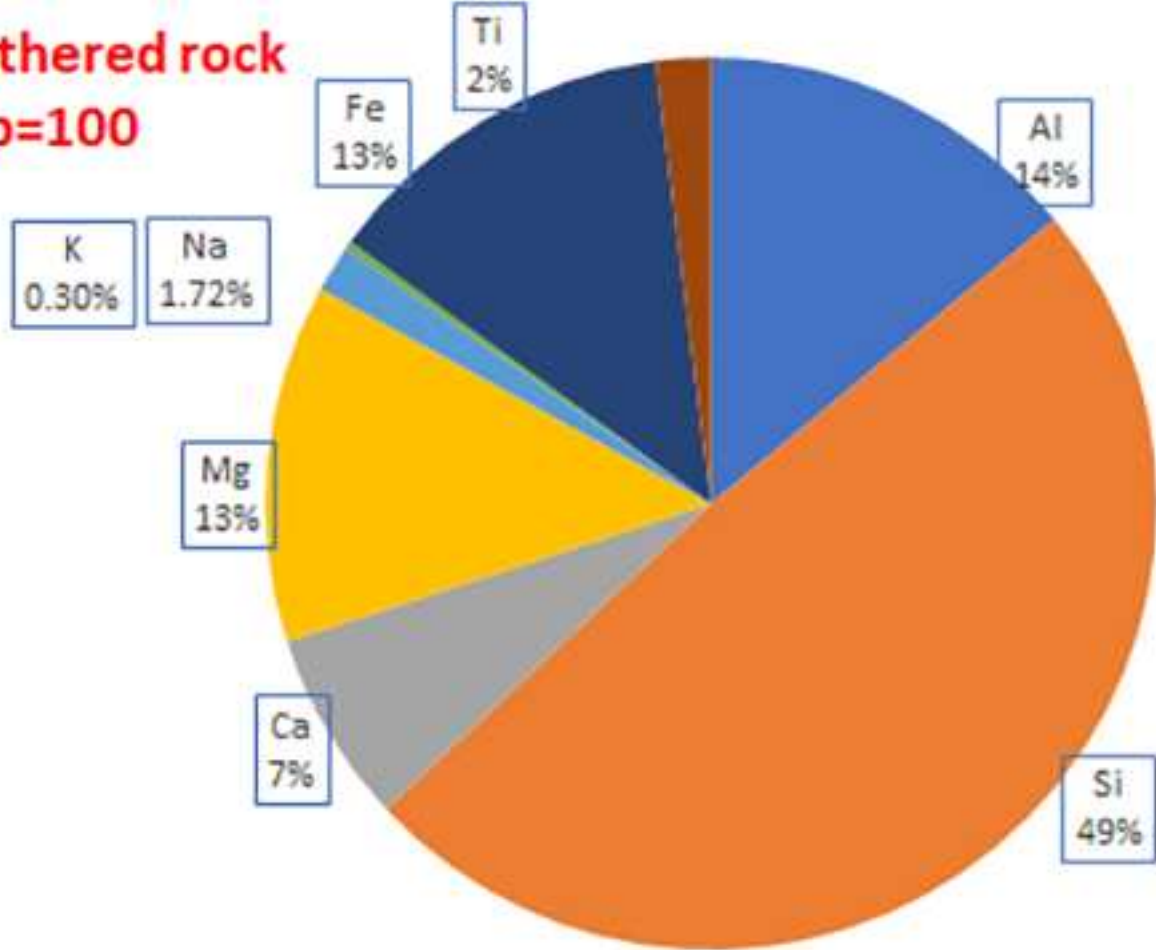
Lateral Sediment
transport at rates
that generally
decrease with
depth below
surface

Dilation with
reduction in
overburden

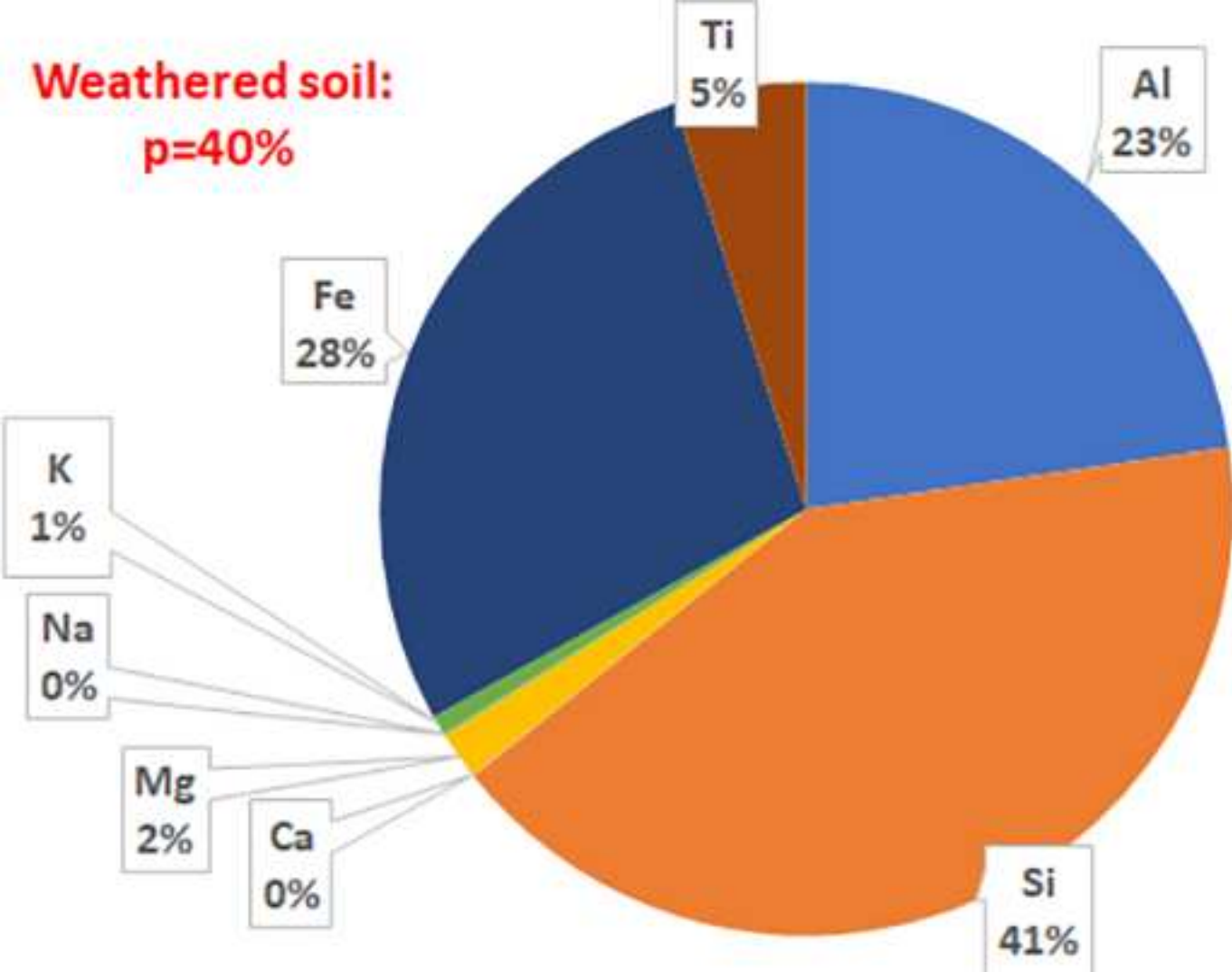


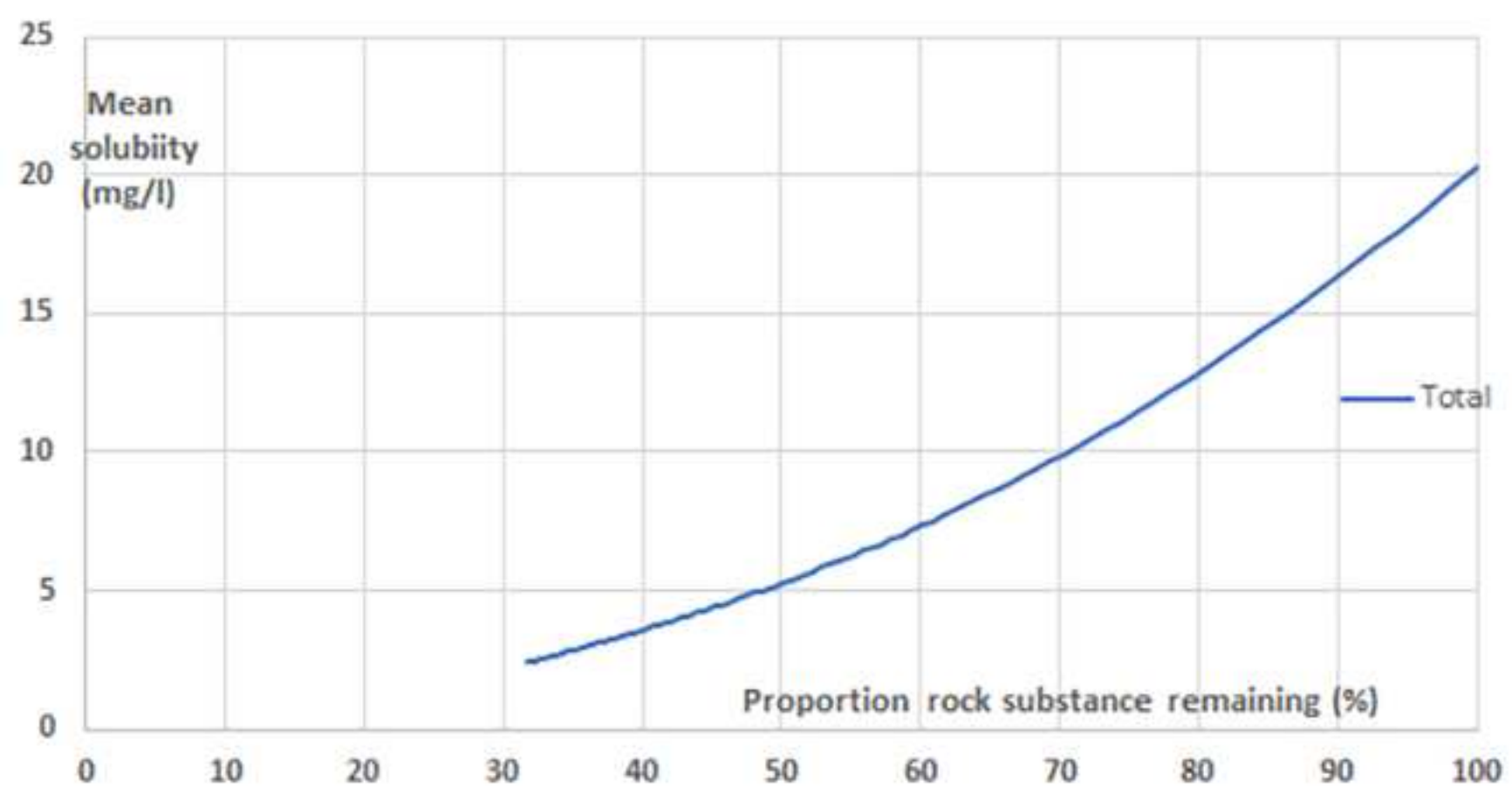


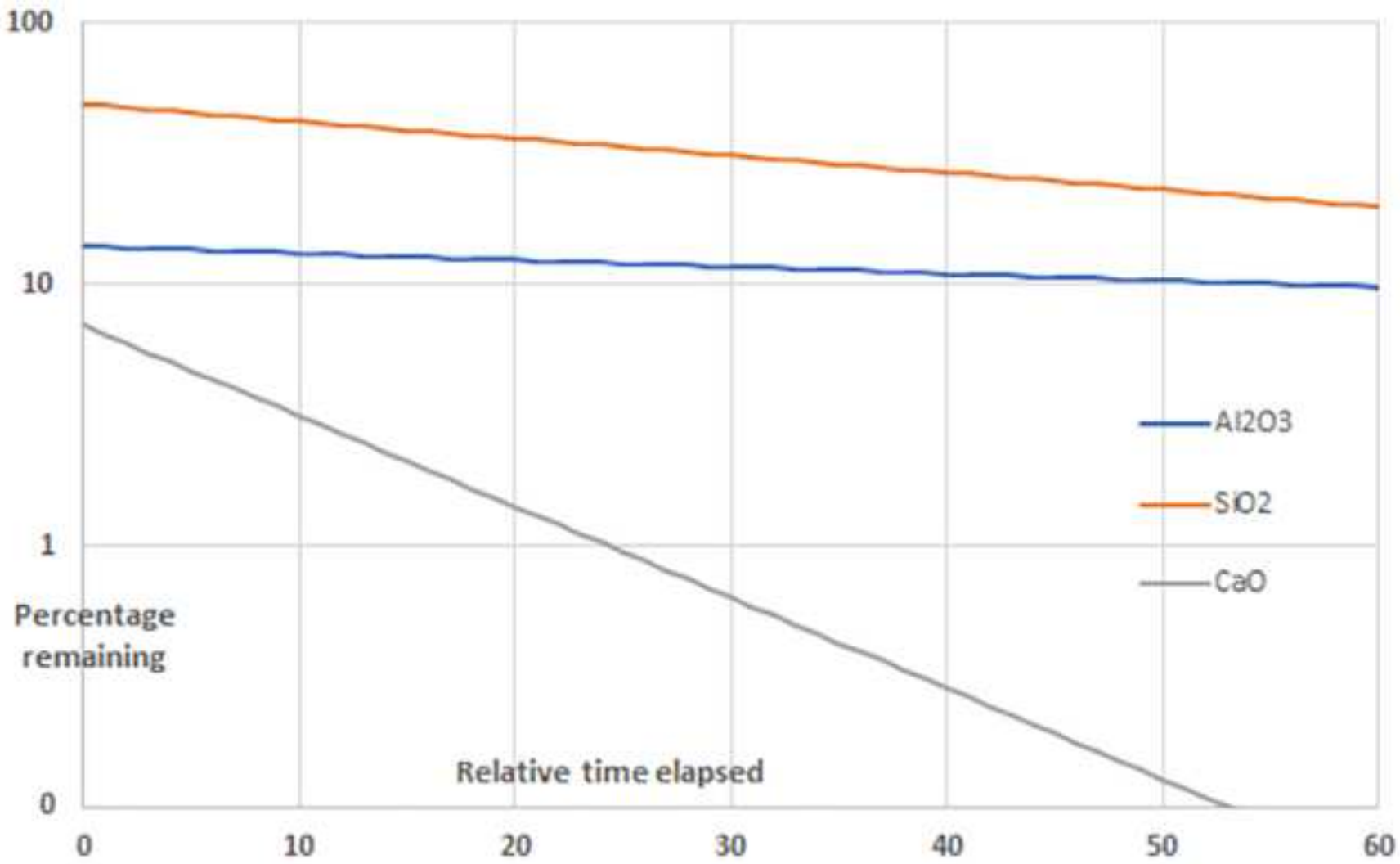
Unweathered rock
p=100



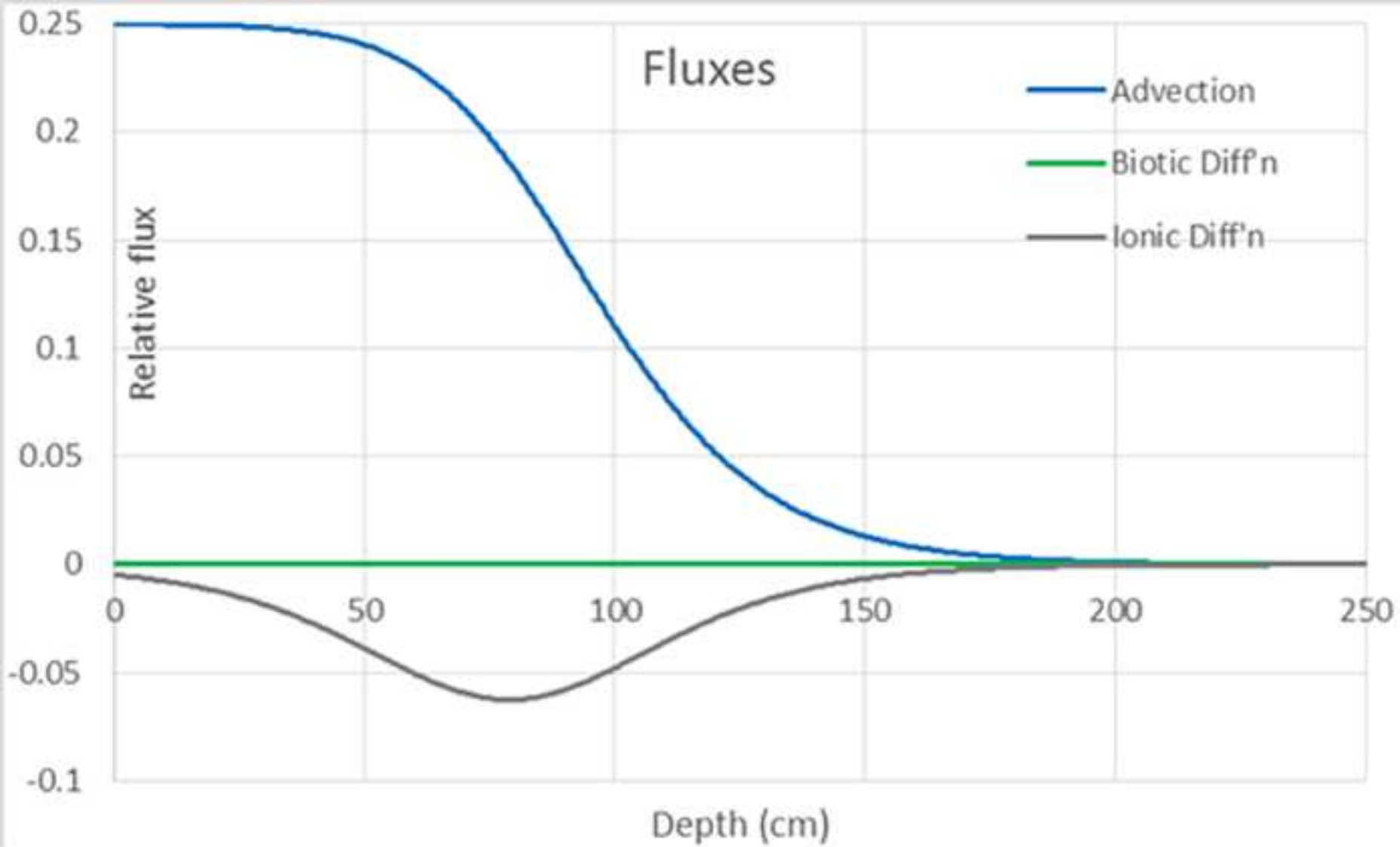
Weathered soil:
p=40%



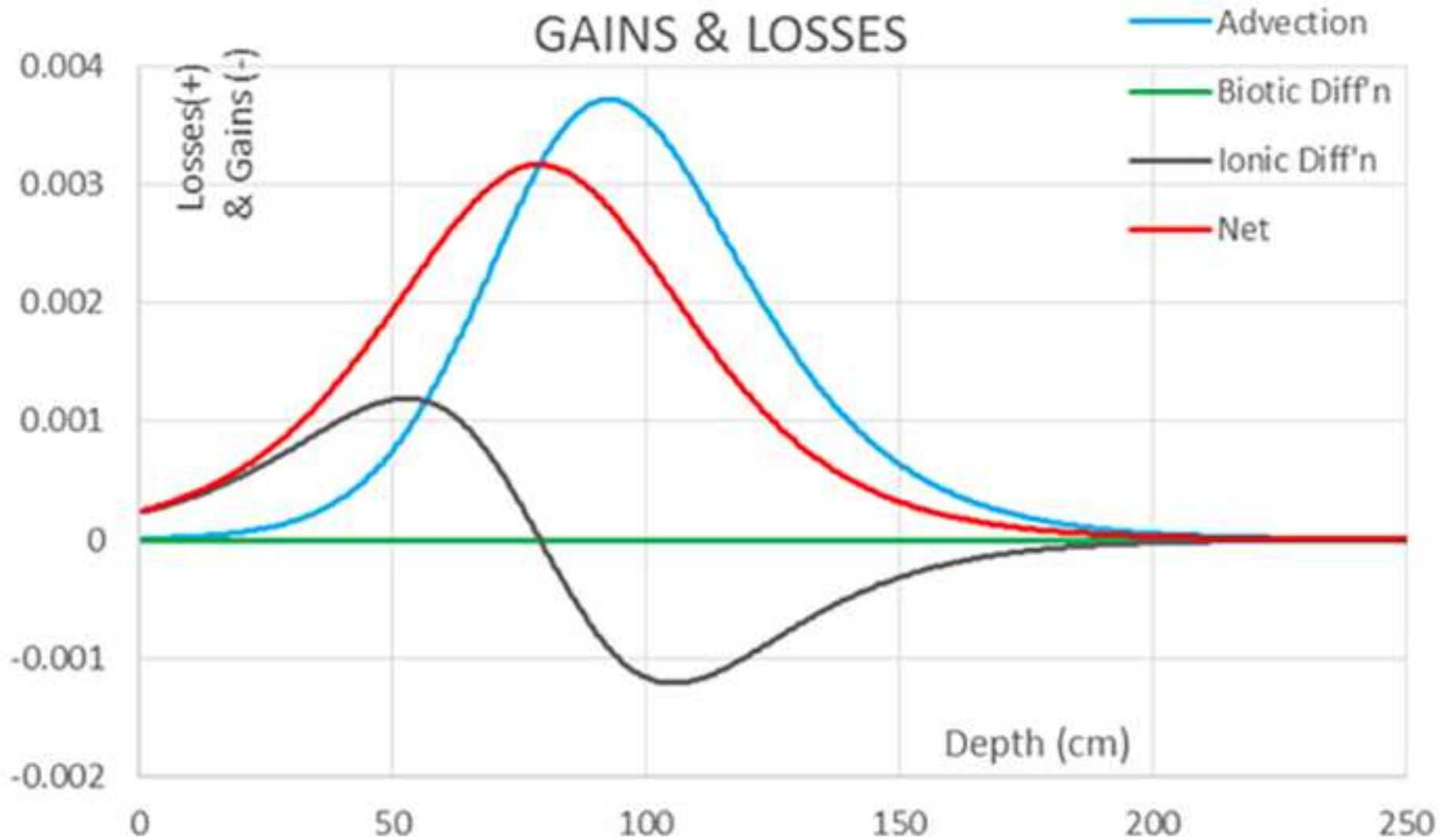


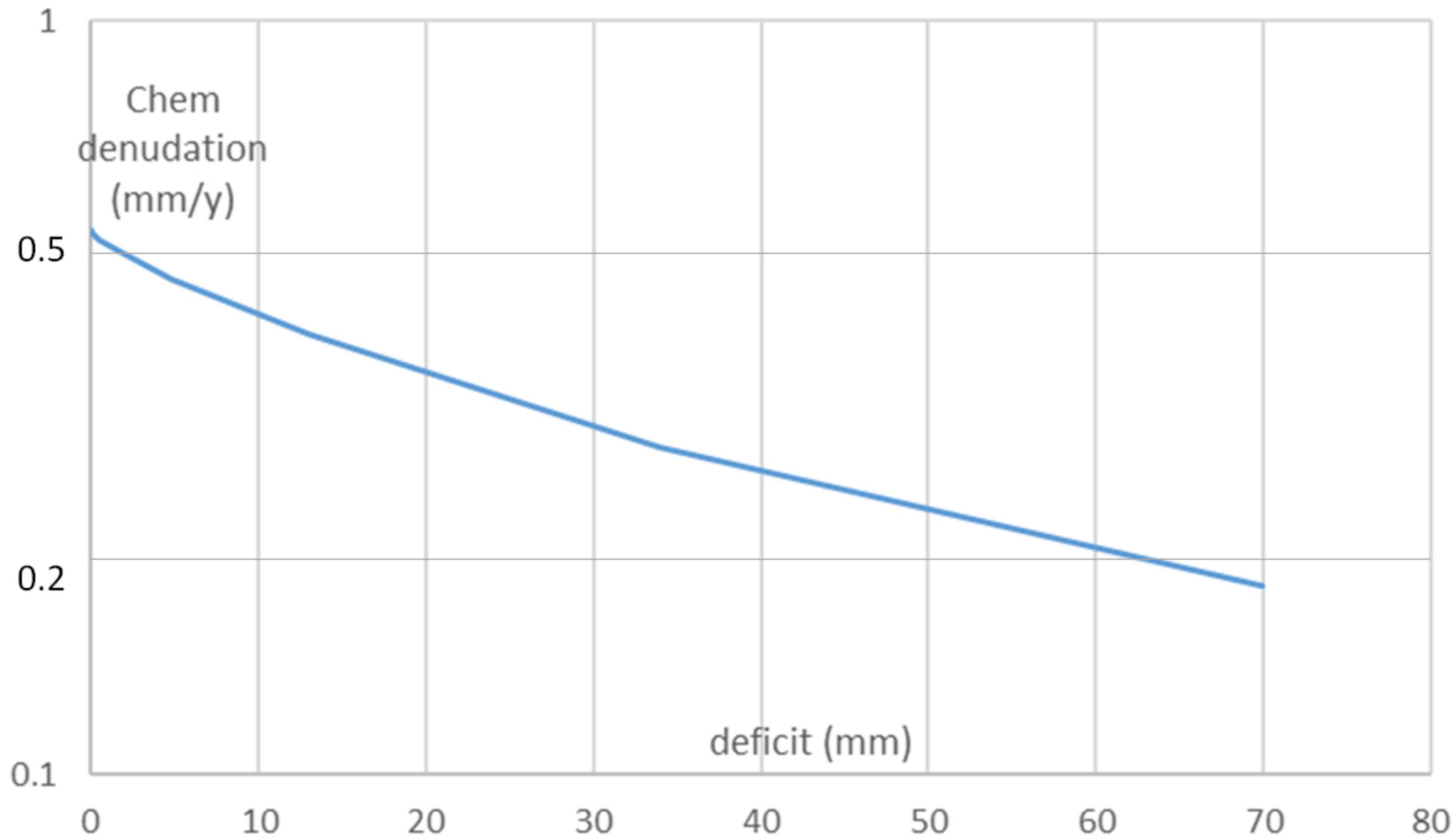


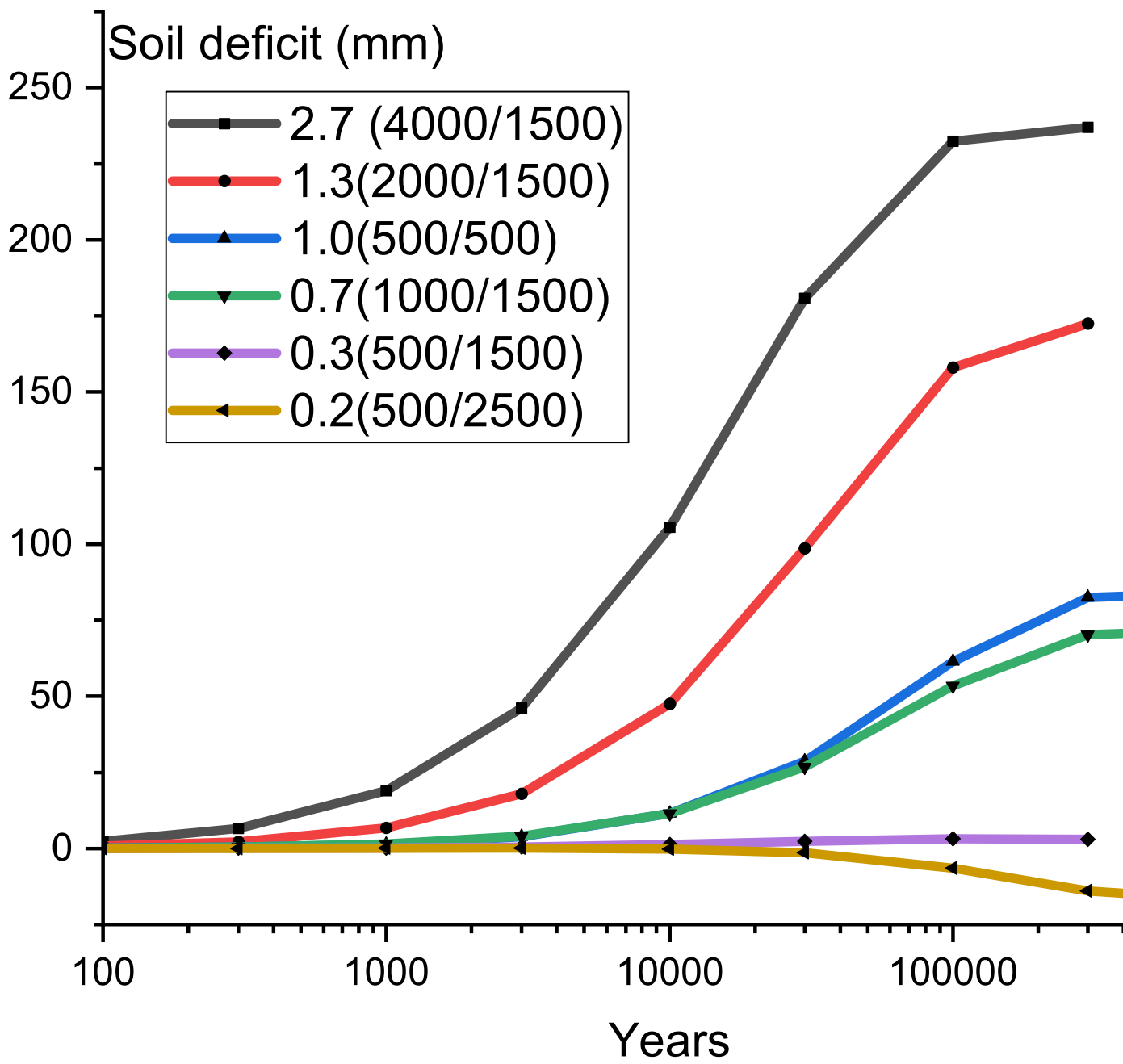
Fluxes

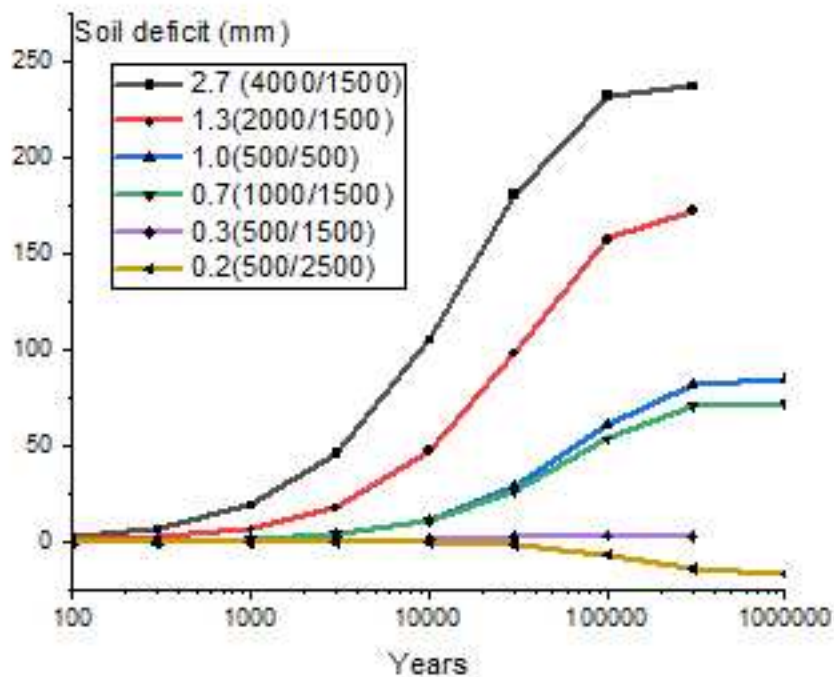


GAINS & LOSSES

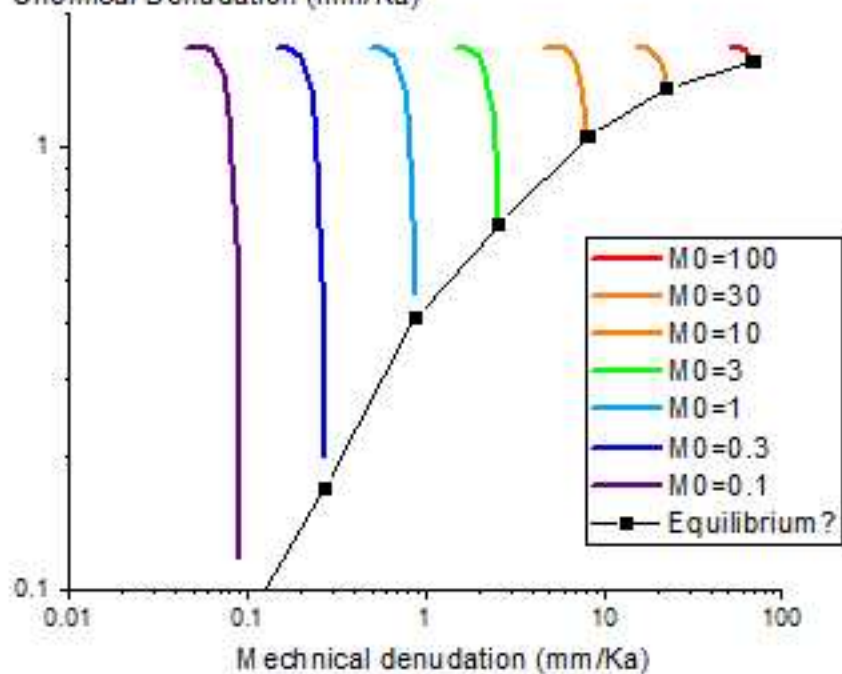








Chemical Denudation (mm/Ka)



Soil deficit (mm)

400

300

200

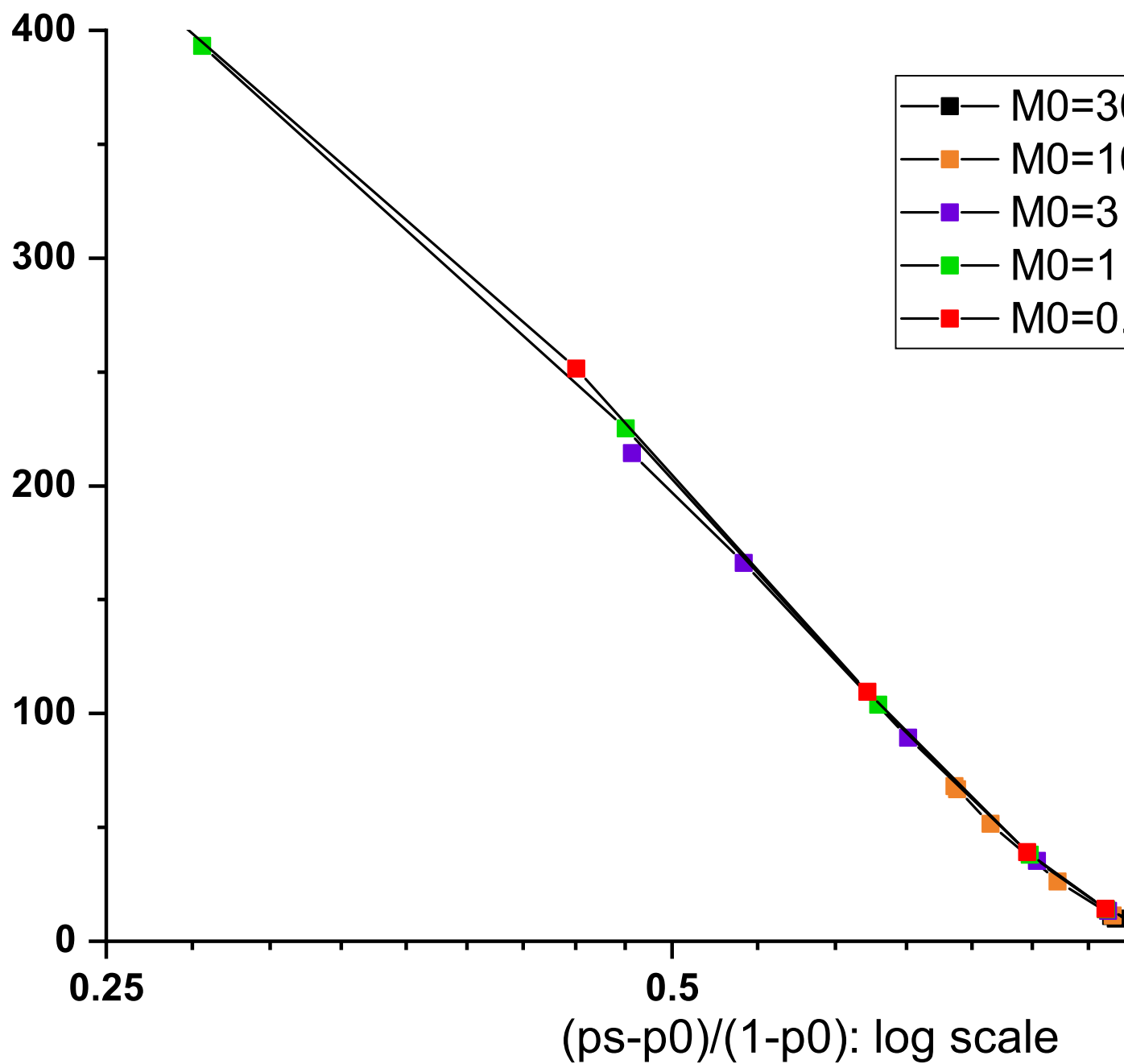
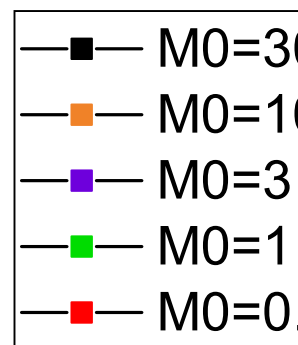
100

0

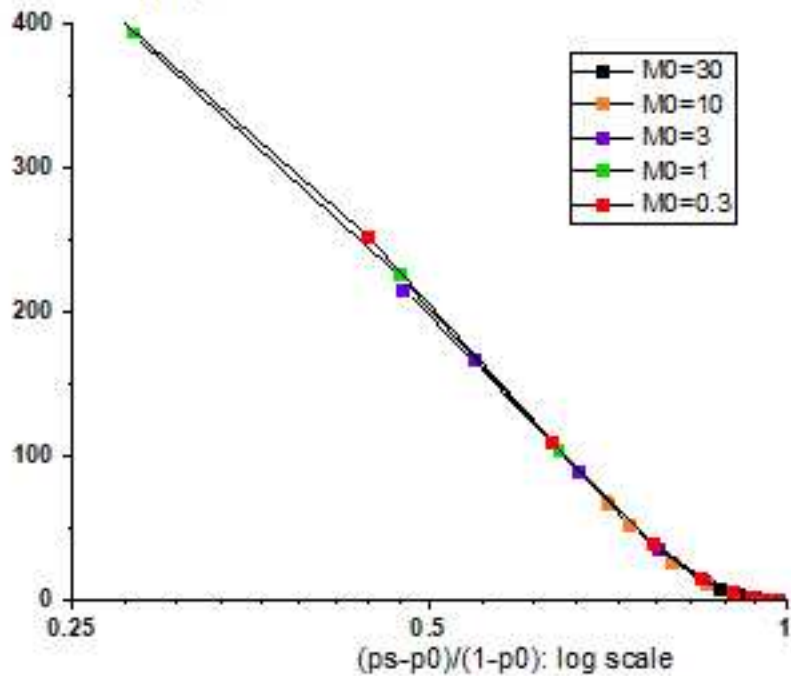
0.25

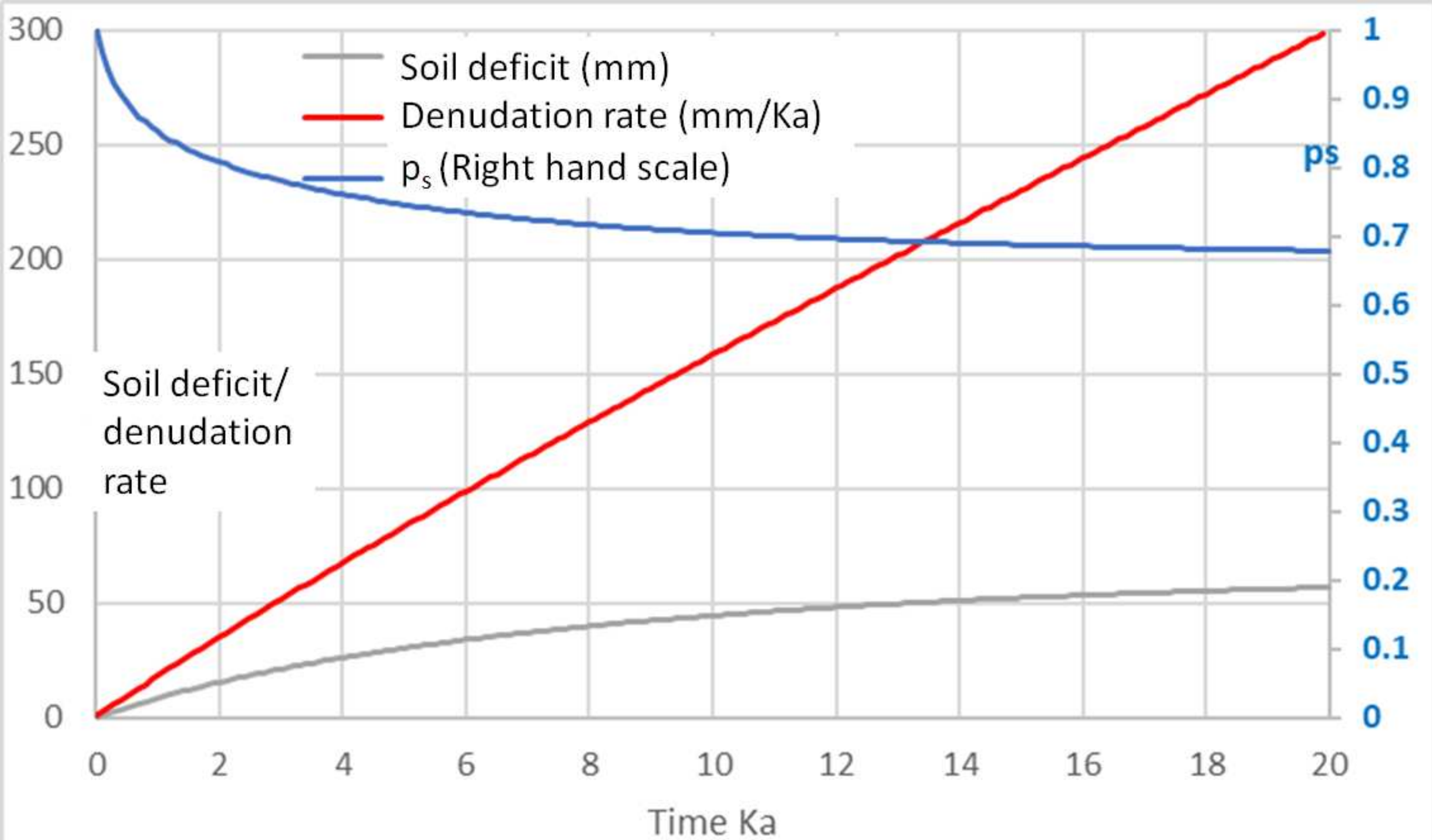
0.5

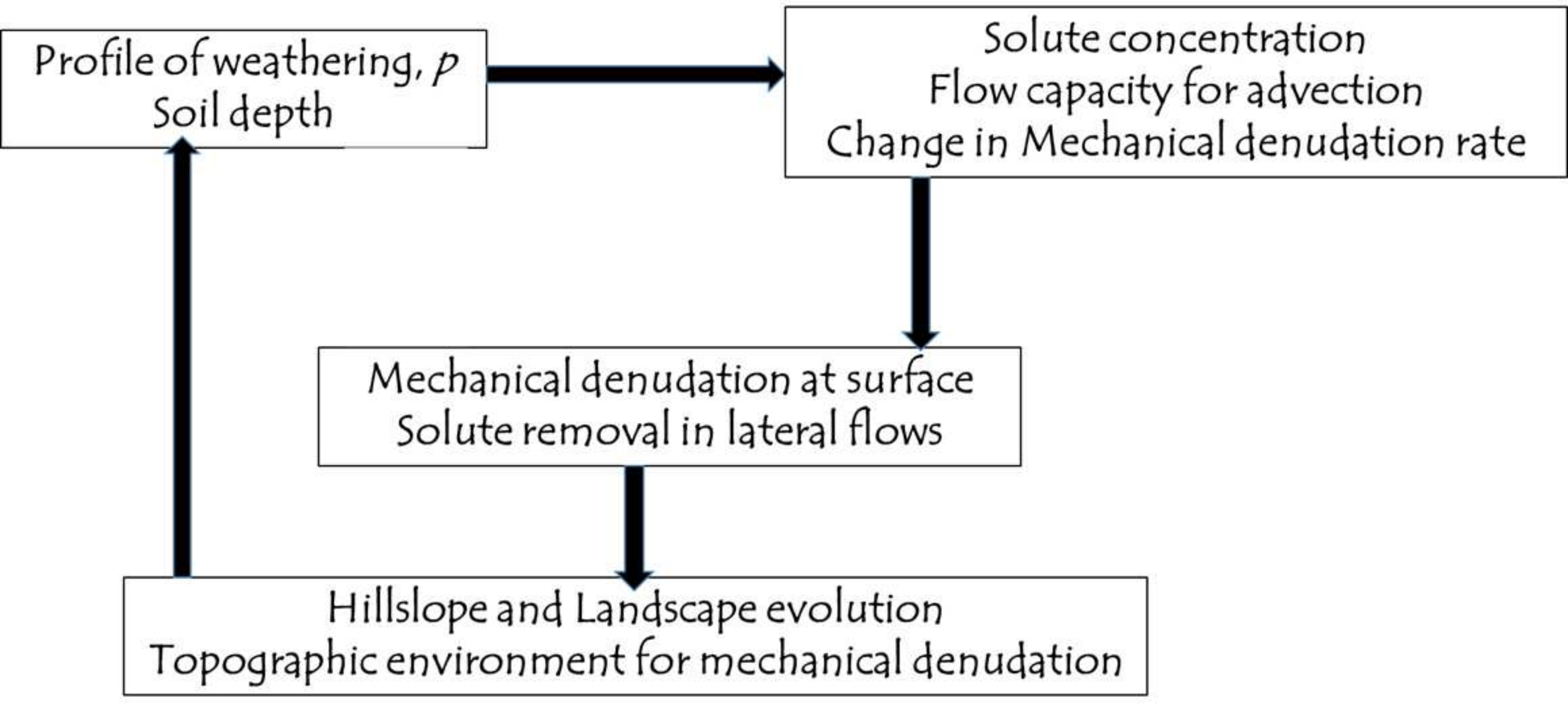
$(ps-p0)/(1-p0)$: log scale



Soil deficit (mm)

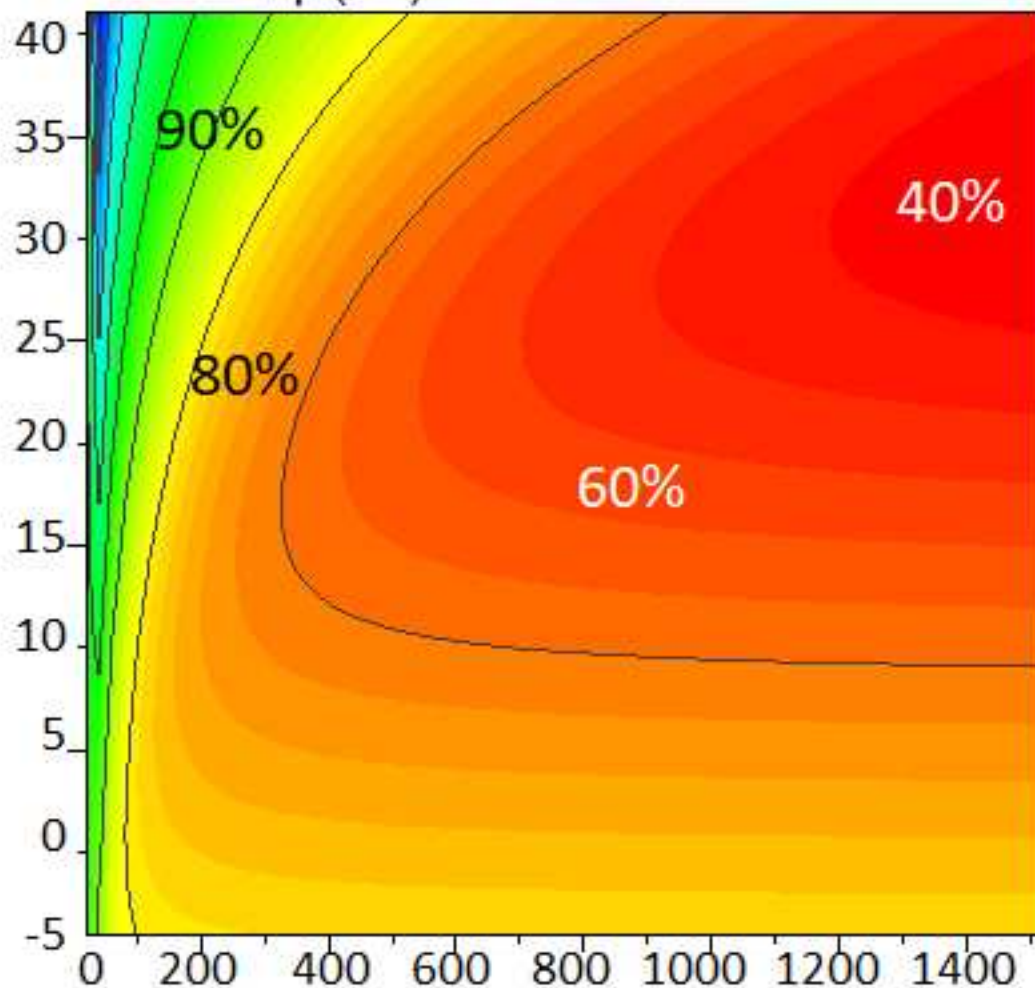






Mean Ann Temp (oC)

Color Scale



Precipitation (mm/a)

40%

60%

80%

90%

100%