

Working Paper 301
THE BASIS FOR SOIL PROFILE MODELLING
IN A GEOMORPHIC CONTEXT

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List of Equations for
W.P. 301

Page 1:

$$v(z) = \int_z^{\infty} (1-p) dz' \quad (1)$$

Page 2:

$$F(z) = F_0 + K \frac{z}{x} v(z) = F_0 + K v(z) \quad (2)$$

Page 3:

$$\begin{aligned} \log \left[\frac{K_1 S_1 O_2^{0.50}}{K_2 S_2 O_1} \right] &= \log \left[\frac{K_1 O_2}{K_2 O_1} \right] = p_1 - p_2 \\ \log \left[\frac{O_2^{0.50} S_2}{S_1} \right] &= \log \left[\frac{O_2}{S_1} \right] + 2D_{O_2} = p_2 K \\ \log \left[O_2^{-0.50} \right] &= pH = 7.0 \\ \log \left[\frac{NO_3^{-eq}}{NO_3} \right] &= \log \left[\frac{O_2}{NO_3} \right] + pH = 7.0 \\ \log \left[NO_3^{-eq} \right] &= 5/4 \log \left[\frac{O_2}{NO_3} \right] + pH = 8.9 \end{aligned} \quad (3)$$

Page 4:

$$\frac{-dp}{dt} = c(y) q \quad (4)$$

$$c(y) = K_p (1-p_y) \quad (5)$$

Page 5:

$$-\delta p / \delta z = \left[(P + \delta P) - (c + \delta c) - (-\delta c) \frac{O_2}{O_2 - p_o} \right] \delta z \quad (6)$$

$$\text{or } \frac{\delta p}{\delta z} = -P \left(z \right) \frac{\delta c(z)}{\delta z} \quad (6)$$

$$\frac{\delta p}{\delta t} = -P \left(z \right) \frac{\delta c}{\delta z} \quad (7)$$

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$$c(z) = k z \quad (8)$$

$$-D_I \frac{\delta c(p)}{\delta z} = -D_I k z \frac{\delta p}{\delta z} \quad (9)$$

Page 7:

$$-D_O \frac{\delta p}{\delta z} \quad (10)$$

Page 8:

$$\frac{dM}{dt} = \mu E = -\kappa M \quad (11)$$

$$n_1 = \frac{1}{2} - (1 + e^{-\frac{z}{z_0}})^{-1} \quad (12)$$

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$$\frac{\partial n_1}{\partial z} = 0/p_s \quad (13)$$

$$\frac{\partial p}{\partial z} = \frac{\partial}{\partial z} \left[D_0 \frac{\partial \psi(z)}{\partial z} + D_1 k_s \frac{\partial \psi}{\partial z} \right] = P(z) \frac{\partial \psi(z)}{\partial z} + \frac{T(z) \partial p}{P_s \partial z} \quad (14)$$

$$\frac{\partial p}{\partial z} = \frac{\partial}{\partial z} \left[(D_0(z) + D_1 k_s) \frac{\partial \psi}{\partial z} \right] = P(z) \frac{\partial \psi}{\partial z} + \frac{T(z)}{P_s} \frac{\partial p}{\partial z} \quad (15)$$

$$(D_0(z) + D_1 k_s) \frac{\partial \psi}{\partial z} = P(z) k_s (p - p_s) \text{ at } z = 0 \quad (16)$$

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$$\frac{\partial \psi}{\partial z} = \frac{\partial}{\partial z} (D_0(z) \frac{\partial \psi}{\partial z}) + \beta \pi = T(z) k \frac{\partial \psi}{\partial z} \quad (17)$$

$$\mu E (1 - e^{-\alpha z}) = T(z) k \pi - D_0(z) \frac{\partial \psi}{\partial z} \text{ at } z = 0 \quad (18)$$

$$\frac{\partial \psi}{\partial z} = \frac{\partial}{\partial z} \left[(D_0(z) \frac{\partial \psi}{\partial z}) \right] - T(z) k \frac{\partial \psi}{\partial z} + \beta \pi = \frac{\mu E}{z_0} = \pi/z_0 \quad (19)$$

$$T(z) = \pi - \frac{\mu E}{z_0} e^{-\alpha z/z_0} \quad (20)$$

$$T(z) \pi \pi - D_0(z) \frac{\partial \psi}{\partial z} = 0 \text{ at } z = 0 \quad (21)$$

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$$\frac{\partial \psi}{\partial z} = \frac{\partial}{\partial z} \left[(D_0(z) \frac{\partial \psi}{\partial z}) \right] - k T(z) \frac{\partial \psi}{\partial z} = \frac{\mu E}{z_0} e^{-\alpha z/z_0} \quad (22)$$

$$\mu E (1 - e^{-\alpha z}) = T(z) k \psi - D_0(z) \frac{\partial \psi}{\partial z} \text{ at } z = 0 \quad (23)$$

$$\frac{\partial \psi'}{\partial z} = \frac{\partial}{\partial z} \left[(D_0(z) + D_1 k_s) \frac{\partial \psi'}{\partial z} \right] - P(z) k_s \frac{\partial \psi'}{\partial z} + \frac{T(z)}{P_s} \frac{\partial \psi'}{\partial z} = \mu \frac{E}{z_0} e^{-\alpha z/z_0} \quad (24)$$

$$\mu E (1 - e^{-\alpha z}) = T(z) k_s (p' - p_s) - D_0(z) \frac{\partial \psi'}{\partial z} \text{ at } z = 0 \quad (25)$$

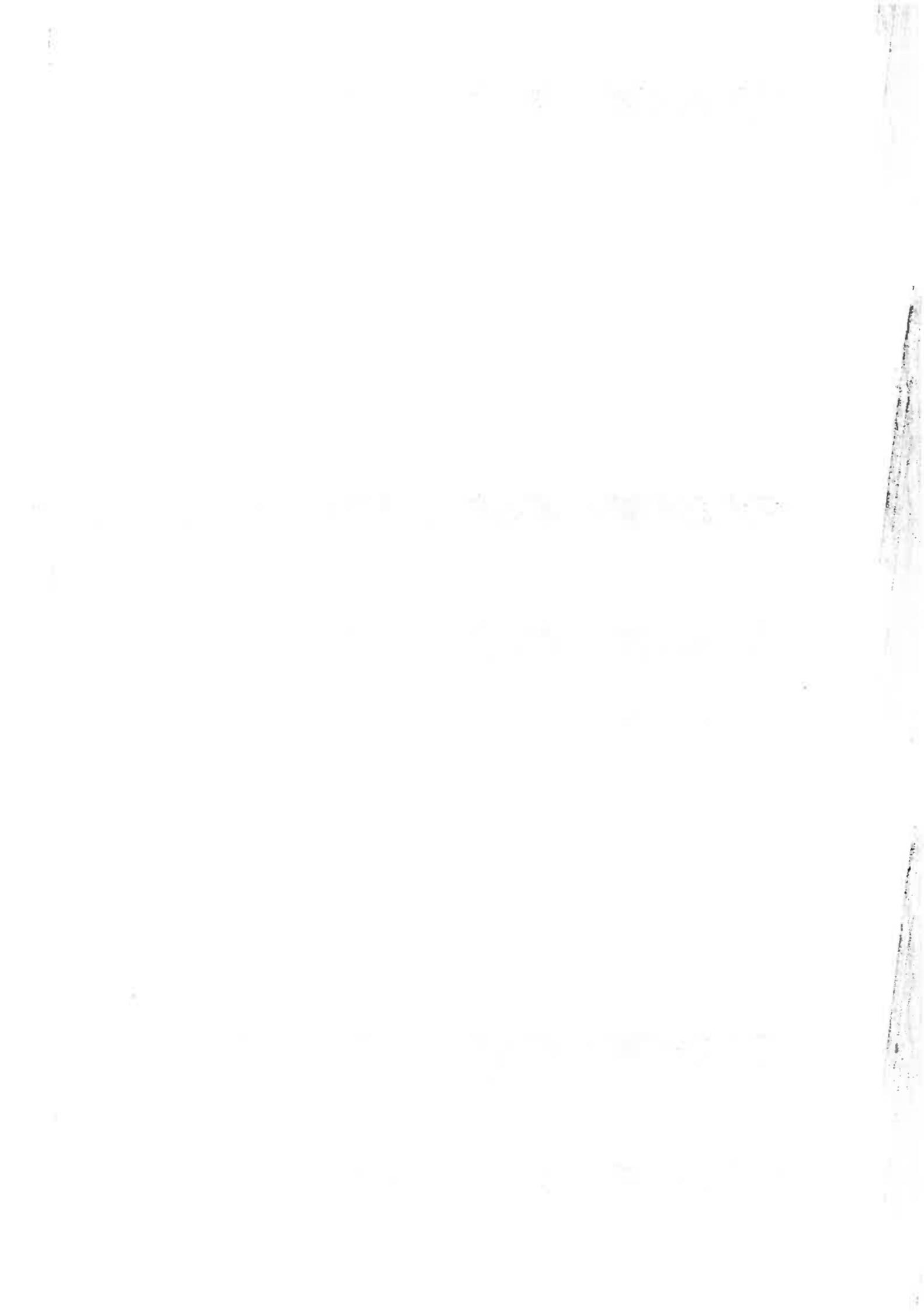
$$\frac{dp}{dz} = \frac{T(z) - T(p - p_s)/p_s}{D_1 k_s (1 - e^{-\alpha z})} \quad (26)$$

$$F C_1 = T(1 - p_s)/P_s \quad (27)$$

$$\frac{c}{c_1} > (p - p_s)/(1 - p_s) \quad (28)$$

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$$P = 1 - \frac{F k_s (1 - p_s)}{1 + F k_s (1 - p_s)} \exp \left[- \frac{(T + F k_s P_s)}{(D_1 k_s + D_0)} z \right] \quad (29)$$



$$\frac{d}{D_1 k_B T_0} \int_0^z \frac{dz}{D_2(z) + D_1 k_B T_0} = z - \left(z + z_2 \ln \left[\frac{D_1 k_B T_0 \exp(-z/z_2)}{D_1 k_B T_0 + D_2(z)} \right] \right) \left(\frac{D_1 k_B T_0}{D_2(z)} \right)^{\frac{1}{2}} \quad (30)$$

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$$\pi = \frac{n E_0}{P_K + D_2/z_2} \exp(-z/z_2) \quad (31)$$

$$z_1 = 2\delta / \left(\gamma / (r^2 k^2 \epsilon_0 \mu_0) - P_K \right)$$

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$$\tau = \frac{n E_0}{P_0 + P_K z_0} e^{-z/z_0} - \frac{n E_0}{P_0 + P_K z_1} e^{-z/z_1} \quad (32)$$

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$$-\frac{\partial U}{\partial x} = U + \frac{\partial \pi}{\partial x} \quad (33)$$

$$+\frac{\partial \pi}{\partial x} = U - \frac{1}{2\pi} \left(\frac{\partial^2 \pi}{\partial x^2} \right) \quad (34)$$

A basis for soil profile modelling in a geomorphic context

Abstract: A mathematical model for soil profile evolution is developed in three parts; respectively for the weathering profile, the organic profile and for the inorganic profile associated with nutrient cycling. Processes modelled are percolation, equilibrium solution, leaching, ionic diffusion, organic mixing, leaf fall, organic decomposition and mechanical denudation. In each component model some of these processes may be neglected, and time-scales differ, so that they may best be separated. The resulting models are all second-order linear partial differential equations with non-constant coefficients. Equilibrium profiles are derived for simplest cases, and numerical models exemplify profile evolution in more realistic cases. Modelled weathering profiles generally show a zone of near-constant composition at the surface, with saprolite developing in deep profiles. Nutrient-cycling profiles show variable development of B-horizons, and may be classified according to the relative depths of rooting, organic soil, and organic mixing processes.

A basis for soil profile modelling in a geomorphic context

The interaction between soil and hillslope development is crucial to the understanding of both separately, even though they are rarely considered together. This paper attempts to explore relatively simple ways of modelling soil profile development in a context which takes account of the geomorphic environment in the form of denudation rate; and is not too complex to couple more closely with hillslope development models. The soil model itself is composed of three linked models, representing respectively the organic soil, the inorganic profile associated with nutrient cycling, and the weathering profile. The last of these has been kept separate because of its longer time-scale, but may in principle be directly combined with the nutrient cycling profiles.

The processes modelled for the soil are nutrient uptake, leaf fall and decomposition as the vegetation cycle; mixing of the surface soil, leaching and ionic diffusion for the transfer of solutes through the soil; together with mechanical denudation at the surface. Water flow through the soil is considered to be controlled through evapotranspiration carrying nutrients to plants, through sub-surface flow diverted at a given depth as deeper layers are saturated, and with some allowance for percolation into unweathered bedrock. Each set of processes is discussed below in order to reach a simple but relevant formulation for the model.

1. Notation and definitions

The state of the soil at any depth is primarily specified through the weathering profile, and describes the 'proportion of substance remaining' (p_-) at any depth. This may be calculated in principle by reference to the reciprocal of the enrichment ratio for an insoluble rock constituent, and may be approximately obtained in practise through reference to Al_2O_3 or T_{i2}O_2 . p_- lies on a scale from zero to 1.0 (unweathered parent material) and weathering profiles normally show an increase in p_- with depth, approaching $p_- = 1.0$ asymptotically (figure 1). An important value of p_- is its surface value, $p_{-\text{s}}$, because mechanical denudation removes material through slides, wash creep etc. at essentially this value.

A summary variable for the soil is the accumulated deficit, defined as:

$$w(z) = \int_z^{\infty} (1 - p) dz' \quad (1)$$

This has the dimensions of depth. $w(0)$ is the total deficit, which may be considered as a measure of total soil depth which makes no assumption about the exact point at which the soil gives way to parent material. In this model it is assumed that weathering etc takes place without

change of position; solution losses being balanced by changes to lower density mineral forms and by increases in porosity. This assumption seems to be a close approximation except for soils developed on highly soluble rocks such as limestone, and it could be relaxed at the cost of some increase in model complexity. $w(z)$ can be used as a measure of pore-space available to percolating waters below depth z , so that percolation may respond to the developing soil profile.

For the organic soil profile and the associated inorganic profile, a more convenient unit of concentration in the soil is related to absolute units, such as kg/m^3 . In comparing the two sets of units, a bedrock density of about $2,500 \text{ kg/m}^3$ corresponding to $p=1$ should be used, \bar{w} is used for concentration (in kg/m^3) and for the associated mineral profile T is used in the same units. The latter profile is calculated relative to the weathering profile as its local datum, and it will be seen that the processes are largely additive so that this separation is justifiable.

2. The percolation process

At depth, percolation into bedrock may be assumed constant as in the examples given below, or may decrease with depth to allow for joint closing under pressure. Figure 2 illustrates the pattern of percolation in a uniform soil on a convex hillside, on which flow-lines will everywhere parallel the surface. At a distance x from the divide, the saturated flow capacity below z must be given by

$$q(z) = xF(z)$$

For the simplest, linear case, the flow law is:

$$q(z) = K.s.w(z)$$

where K is the flow velocity on unit hydraulic gradient, so that:

$$F(z) = F_0 + K \frac{s}{x} w(z) = F_0 + \lambda w(z) \quad (2)$$

where F_0 is the percolation rate at depth, and λ is a constant.

For hillslopes which are not simply convex, and for non-linear flow laws, equation (2) is a simplification, but it still seems reasonable to treat $F(z)$, the maximum percolation rate as a specified function of the accumulated soil deficit, $w(z)$.

The upper limit on percolation is set by rainfall and evapotranspiration rates, which are treated as annual totals only, without allowance for seasonal effects which may be significant for some semi-arid soils. Evapotranspiration and rainfall totals are specified as parameters of the model, and the former is considered to be extracted in proportions determined by a (fixed) distribution of roots with depth. This distribution has initially been considered as an exponential decay with a scale

depth z_0 . At any depth rainfall less accumulated evapotranspiration sets an upper limit on the amount of water available for downward percolation. The actual percolation estimate is thus the lower of the two estimates:

(i) rainfall less accumulated evapotranspiration

or (ii) percolation calculated from accumulated soil deficit as above.

Other limits might be set, for example in the presence of a regional water table level.

Where percolation is limited by the soil properties, water is considered to be diverted laterally as subsurface flow carrying solutes away at a concentration equal to that in the percolating waters.

3. The solubility model

A linear model is presented here as an approximation to the equilibrium thermodynamics of the soil water chemistry. The assumption that water residence times are long enough to approach equilibrium is a crucial assumption to the model. The approach presented is a development of that in Carson and Kirkby (1972, Chapter 9) and in Kirkby (1976), but a very much simpler form is derived from it which allows the soil model of this paper to be developed.

The assumption of equilibrium is based on experimental results which show very reduced reaction rates after a period of approximately 100 hours (eg Bricker et al 1968), which is short compared to measured residence times for most soil water systems. The equilibrium assumed is for the water. Plainly the soil does not reach equilibrium, but is thought to change slowly enough that water in contact with it may be considered to be in equilibrium with the soil composition at any time. This kind of relationship between systems with different response times is a normal one in the natural environment, as is the assumption of equilibrium for the more rapidly changing partner in such an association.

The assumption of linearity is to treat the soil solids as behaving like a mixture of their constituent oxides, with activities equal to their respective mole fractions. If empirical adjustments are made to the Gibbs' Free Energy values, then the course of weathering for a given parent material can be modelled accurately enough for present purposes; although the empirical adjustments required vary with the parent material and its mineral assemblages. With this assumption, equilibrium concentrations may be calculated from a series of equations of which the following are examples:

$$\left. \begin{aligned} \log [H_4 S_i O_4^{aq}] &= \log [S_i O_2^s] - 3.81 \\ \log [Mg^{++} aq] &= \log [Mg O^s] + 10.00 - 2pH \\ \log [OH^{-} aq] &= pH - 14.0 \\ \log [HCO_3^{-} aq] &= \log P_{CO_2}^{gas} + pH - 7.9 \\ \log [NO_3^{-} aq] &= 5/4 \log P_{O_2}^{gas} + pH - 8.9 \end{aligned} \right\} (3)$$

If the partial pressure of CO_2 gas ($p_{\text{CO}_2}^{\text{gas}}$) is treated as the independent variable and the partial pressure of oxygen and CO_2 together are assumed to sum to 0.2 atmospheres (in the presence of 0.8 Atm of N_2), then these equations may be solved to give pH and the concentration of each relevant ion. The course of weathering may be calculated in a non-spatial context, where the time scale is given by the volume of initially clean water passing through, expressed in litres of water per milligram of original rock. In terms of p , the proportion remaining, the process of weathering may be expressed as:

$$\frac{dp}{dt} = c(p) q \quad (4)$$

where $c(p)$ is the total solute concentration of p and q is the rate of flow through the soil in $l/\text{mg.y}$. It may be noted that climatic effects only enter indirectly into this formulation, through control of rates of rainfall percolation and through the vegetation cover which influences the partial pressure of CO_2 . Sample results are shown in figure 3 for simplified rocks, indicating the range of weathering sequences as the partial pressure of CO_2 varies over its likely range, while keeping the Free Energy values in equations (3) etc. constant. The values and courses shown broadly match those observed in real soils, although with rather higher base solubilities than are effective in reality. The time spans indicated by the plotted points, at intervals of 0.04 l/mg may be compared for example with percolation of 0.5 m/y of rainwater through 1m of soil. At this rate, the plotted points represent intervals of 80,000 years on average.

Compositions are uniquely associated with values of p at any of the stage of the weathering, from a given parent material, as can values of pH. For present purposes, the most useful summary of the results from this submodel are given in figure 4, which shows the total solute concentration as a function of proportion of bedrock remaining, p . It may be seen that with a good degree of approximation, the relationship may be decomposed into a series of straight lines, each of the form:

$$c(p) = k_*(p - p_*) \quad (5)$$

for constants k_* , p_* .

Such a linear expression is equivalent to solution of material, a proportion p_* of which is totally insoluble and the remainder of which has solubility k_* . It must be remembered, however, that for the rock, neither k_* nor p_* can be directly identified with particular components, but represents a composite behaviour associated with loss of constituents together in a gradually changing pH environment. Equation (5) provides an important key to the possibility of developing a simple overall soil model, in which individual constituents need not be identified separately, but composition and pH can readily be derived from the value of p_* . For most purposes, a single straight line segment of the expression is enough to model the whole soil evolution, though it may become necessary to take more than one segment into account.

Processes which are not modelled by the above simplifications include complexing and lessivage. Both have the effect of mobilizing Aluminium and Iron at the expense of Silica. In extreme cases, the curves of figure 3 might converge on the silica rather than the sesquioxide corner of the diagram, leaving a clean sand rather than a clay as the end product of weathering.

4. The leaching process

Net loss of solutes from the soil profile as a whole can only take place in water flows leaving the profile (ignoring grazing etc.), either as sub-surface flow or by percolation into the bedrock. At any depth the amount of solutes carried downwards in percolation water at depth Z is:

$$F(z)c(p)$$

Taking into account diversion as sub-surface flow (figure 5), the net loss of solutes by leaching is:-

$$-Sp.S_z = [(F+SF)(c+Sc) + (-SF)c-Fc] St$$

$$\text{or } \frac{dp}{dt} = - F(z) \frac{c(p)}{z} \quad (6)$$

which is the spatial analogue of equation (4)

Substitution of equation (5) for $c(p)$ then gives:

$$\frac{dp}{dt} = - F(z)k_* \frac{p}{z} \quad (7)$$

For the inorganic profile associated with nutrient cycling, equation (6) remains valid, but $c(p)$ must be interpreted relative to the weathering profile. It is therefore the marginal concentration for the plant nutrients. For simplicity, a linear model has been adopted for trial runs, namely:

$$c(p) = kp \quad (8)$$

Where p is the concentration of inorganic materials in kg/m^3 relative to the weathering profile

and k is a constant solubility. The value of the solubility may well be greater than k_* for the soil as a whole, since plant nutrients are a selection of soil constituents which is biased towards the more soluble, notably the bases.

For the organic profile, it has been assumed in most trial runs that the inorganic nutrients are not available for leaching until released by decomposition, so that leaching can be ignored in estimating the organic profile. It is, however, very simple to allow solution from the organic matter, either at the same rate as in equation (8) above, or at some lower rate as appropriate.

It should also be noted that the zone of nutrient cycling is one in which the partial pressure of CO_2 in the soil is increasing notably with depth, from its atmospheric level of 0.0003 Atmospheres to its equilibrium soil level of 0.01 to 0.10 Atm. There may therefore be valid reasons for allowing the relevant solubilities to decrease with depth initially.

5. Ionic diffusion

Diffusion on its own cannot export solutes from the profile, but may be important in distributing them within it. The rate of transport is proportional to the concentration gradient and in the direction from high to low concentration. Where rates of reaching are everywhere low, as at the edges of impermeable bedrock, then weathering can only proceed by diffusion, which is then at least locally the dominant process. The downward transport rate is given, for the linear expression of equation (5) by:

$$- D_I \frac{c(p)}{z} = - D_I k_* \frac{p}{z} \quad (9)$$

The diffusion coefficient, D_I , varies only slightly between different ions, but shows some dependence on the water filled porosity of the soil, ϵ , so that approximately

$$D_I = 0.06 \text{ } m^2/y \quad (\text{Lerman, 1979; p92})$$

A constant value of $0.01 \text{ m}^2/\text{y}$ has been adopted for trial runs of the model, though this may tend to over-estimate the initial rate of bedrock dissolution.

6. Organic mixing

The surface soil layers are mixed mechanically, mainly through the burrowing activities of mesofauna, most notably but not exclusively earthworms. The intensity of the mixing is related to the base status

of the soil which indirectly provides nutrition for the soil fauna. A second condition is adequate soil aeration, with very low levels of mixing in waterlogged soils. Mixing undoubtedly declines in intensity with depth, in association with these factors.

Mixing is also responsible for the production of a soil bulk-density profile, a process which is ignored in the present soil model, but one which could in principle be used to obtain an independent estimate of mixing intensity. On a slope, the same process of outward diffusion and settlement under gravity produces one form of soil creep, with a magnitude per unit slope gradient roughly equal to the surface mixing intensity, so that rates of diffusion creep provide another method of estimating mixing intensity.

Such estimates suggest that the order of magnitude of mixing lies in the range, $D_o = 10^{-4}$ to $10^{-2} \text{ m}^2/\text{y}$, with low figures for mor humus, and high figures for null humus. In fact for values as high as $10^{-2} \text{ m}^2/\text{y}$, diffusion may be the dominant cause of soil creep movement. The depth of organic mixing is plainly related to the depth of the organic and nutrient-linked inorganic profiles, and for most trial runs has been assumed to decrease with depth exponentially over scale depths of $Z_2 = 0.5\text{M}$.

Downward transport by organic mixing processes will be at rate

$$- D_o \frac{\partial p}{\partial z} \quad (10)$$

where D_o is in general treated as a decreasing function of depth. Comparison with equation (9) shows that organic mixing is very much larger than ionic diffusion near the surface, because the concentration gradient refers to the solid soil rather than to the solutes in water, but at depths of several meters ionic diffusion becomes dominant.

7. Nutrient uptake

Plant growth is not modelled seasonally, but is assumed to rely on a specified evapotranspiration stream. Because plants establish themselves relatively quickly, a fixed root distribution with depth is assumed as has been described for the percolation model above. The supply of inorganic nutrients is modelled as being at a fixed concentration for each, which is determined by the plant species or community, and is also in fixed proportion to the amount of CO_2 photosynthesised. This assumption is a great simplification of plant physiology, but one which seems to be accurate enough for the purposes of an adequate soil model.

A second assumption of the model is that plant nutrients are taken from the inorganic soil only. It is visualised that organic compounds must first release nutrients in simple ionic forms before they come available for re-use by living plants. If this assumption is incorrect, then the modelled balance between organic and associated inorganic profile will also be incorrect, but not the total nutrients in the two, taken together. For trial purposes, a figure of $\mu = 0.06$ gram of inorganic nutrients has been assumed per litre of water transpired, corresponding to approximately 1 gram of carbohydrate produced per litre of water.

8. Leaf-fall

A constant proportion of the plant biomass is assumed to fall annually, again without considering seasonal variations. The leaf-fall is added to the organic soil at the surface only, ignoring root contributions. The implicit model for above-ground plant biomass, M , is thus:

$$\frac{dM}{dt} = E - \lambda M \quad (11)$$

where E is the rate of nutrient uptake (g/l)

E is the actual evapotranspiration rate (m/y)
and λ is the proportional leaf fall (y^{-1})

For growth from zero, equation (11) has the solution for leaf-fall.

$$M = E (1 - e^{-\lambda t}) \quad (12)$$

The leaf-fall may be seen to approach an equilibrium rate balancing uptake over a time scaled to λ^{-1} years. A value of 0.04 has been used for the rate of leaf-fall, λ , in trial runs, roughly corresponding to a forest cover.

9. Decomposition

For a given soil climate, decomposition is assumed to break down organic matter at a constant proportional rate. In practice, rates vary, being highest for sugars and lowest for woody tissues, so that an average ratio under-represents both the age of the oldest organic matter in a profile and consequently its eventual depth of penetration. Despite this shortcoming, a constant proportional rate of $\lambda = 0.2$ per year has been used in trial runs, roughly corresponding to average temperatures of 10°C .

Inorganic nutrients in the organic matter are assumed to be released at the same proportional rate, and it is argued that they are mainly associated with the components which are most readily decomposed, so that a constant value of λ is more appropriate than for the organic soil as a whole. The nutrients released are added to the inorganic soil,

completing the nutrient cycle which began with their extraction by the plants.

10. Mechanical denudation

The final process considered is stripping of soil from the surface, which is considered to take place through the action of soil creep, splash, wash and mass movements. The effect of these processes together are modelled through a simple denudation rate, T , expressing the rate of surface lowering in m/y . Rates are usually in the range from zero to $20 m/y$, so that their impact can usually be ignored except over the long time periods relevant to the development of weathering profiles. Denudation rates are expressed in equivalent depths of parent material, so that the loss of surface elevation is:

$$\frac{dz_s}{dt} = T/p_s \quad (13)$$

where z_s is total loss of elevation from the initial surface and p_s is the proportion remaining at the surface.

11. The model equations

For the weathering profile, the nutrient cycling is ignored, and the near-surface percolation limit can normally be set at a level corresponding to (rainfall less evapotranspiration), although in shallow soils it may be necessary to take account of the transpiration losses progressively. If the profile is calculated relative to the eroding surfaces, these assumptions lead to:

$$\frac{\partial p}{\partial t} = \frac{\partial}{\partial z} \left\{ D_I \frac{\partial c(p)}{\partial z} + D_o(z) \frac{\partial p}{\partial z} \right\} - F(z) \frac{\partial c(p)}{\partial z} + \frac{T(t)}{p_s} \frac{\partial p}{\partial z} \quad (14)$$

which may be combined with the solution model of equation (7) above to give:

$$\frac{\partial p}{\partial t} = \frac{\partial}{\partial z} \left[(D_o(z) + D_I k_*) \frac{\partial p}{\partial z} \right] - F(z) k_* \frac{\partial p}{\partial z} + \frac{T(t)}{p_s} \frac{\partial p}{\partial z} \quad (15)$$

This equation must be solved for suitable expressions for D_o , F and T , subject to the boundary condition of zero solute transport at the surface, namely:

$$(D_o(z) + D_I k_*) \frac{\partial p}{\partial z} = F(z) k_* (p - p_*) \text{ at } z = 0 \quad (16)$$

A normal initial condition is likely to unweathered bedrock, that is:

$$p = 1 \text{ for all } z \text{ at } t = 0$$

but others may be appropriate in particular cases. Equation (13) above is also needed to give the progress of the surface erosion.

For the organic profile, surface erosion and ionic diffusion may usually be ignored and the concentration \bar{W} (in kg/m^3) is given by:

$$\frac{\partial \bar{W}}{\partial t} = \frac{\partial}{\partial z} \left(D_o(z) \frac{\partial \bar{W}}{\partial z} \right) - \bar{W} + F(z)k \frac{\partial \bar{W}}{\partial z} \quad (17)$$

The last term in this expression has been ignored in most trial runs (section 4 above), and is generally negligible relative to the diffusion term in the presence of even low rates for organic mixing. The appropriate boundary conditions are a given leaf-fall input at the surface:

$$F(z)k \bar{W} - D_o(z) \frac{\partial \bar{W}}{\partial z} \text{ at } z = 0 \quad (18)$$

and $\bar{W} \rightarrow 0$ as $z \rightarrow \infty$

A normal initial condition is normally an absence of vegetation or organic soil, so that:

$$\bar{W} = 0 \text{ for all } z \text{ at } t = 0$$

For the inorganic profile linked to the nutrient cycling, the concentration \bar{T} (in kg/m^3) is given by:

$$\frac{\partial \bar{T}}{\partial t} = \frac{\partial}{\partial z} \left(D_o(z) \frac{\partial \bar{T}}{\partial z} \right) - F(z)k \frac{\partial \bar{T}}{\partial z} + \bar{W} - \frac{E}{Z_0} \quad (19)$$

where the final two terms are respectively the addition of nutrients from organic matter decomposition, and nutrient extraction distributed with evapotranspiration in accordance with the depths of roots. The percolation in this zone may usually ignore limits set by the permeability of slightly weathered rock, so that:

$$F(z) = R - \frac{E}{Z_0} e^{-z/Z_0} \quad (20)$$

The surface boundary condition is of zero transfer, so that

$$F(z)k\bar{T} - D_o(z) \frac{\partial \bar{T}}{\partial z} = 0 \text{ at } z = 0 \quad (21)$$

and at depth:

$$\bar{T} \rightarrow 0 \text{ as } z \rightarrow \infty$$

As before, the relevant initial condition is usually that

$$\bar{T} = 0 \text{ for all } z \text{ at } t = 0$$

If the solubility for nutrients held in the organic soil is the same as in the associated unorganic soil, instead of being negligible; then equation (17) and (19) may be summed to give the combined profile concentration, $\bar{W} + \bar{T}$:

$$\frac{dp}{dt} = \frac{1}{Z} \left[(D_o(z) \frac{dp}{dz}) - kF(z) \frac{p}{Z} \right] - \frac{E}{Z_0} e^{-z/Z_0} \quad (22)$$

subject to the surface boundary condition:

$$E(1-e^{-kt}) = F(z)k_* - D_o(z) \frac{p}{Z} \text{ at } z = 0 \quad (23)$$

If the solubility for these profiles is the same as in the weathering profile as a whole, then with a change in the units for k and F to make them comparable to the p of the weathering profile, summing again gives, for a modified variable $p' = p + \psi$

$$\frac{dp'}{dt} = \frac{1}{Z} \left[(D_o(z) + D_I k) \frac{dp}{dz} - F(z)k_* \frac{p}{Z} + \frac{T(t)}{ps} \frac{dp}{Z} \right] - \frac{E}{Z_0} e^{-z/Z_0} \quad (24)$$

with a surface boundary condition equivalent to (23) above:

$$E(1-e^{-kt}) = F(z)k_* (p' - p_*) - D_o(z) \frac{dp}{Z} \text{ at } z = 0 \quad (25)$$

In this case $F(z)$ must be calculated to take account of both transpiration and weathering effects. In the trial runs, the three sets of equations have generally been solved separately, but the differences in principle and practice are slight.

12. The weathering profile

If $F(z)$, $T(t)$ are held constant, the weathering equation (14) may be integrated to give, with the boundary condition (16), and for equilibrium ($\partial/\partial t = 0$):

$$\frac{dp}{dz} = \frac{\frac{T}{ps}(p-ps)}{D_I \frac{dc}{dp} + D_o(z)} \quad (26)$$

For a meaningful profile the denominator must remain positive, and non-zero except as $Z \rightarrow \infty$, when

$$FC_1 = \frac{T}{ps} (1-ps) \quad (27)$$

where C_1 is the value of $C(p)$ at $p = 1$.

or

Substituting for T/ps in the denominator of (26), the required condition is:

$$\frac{C}{C_1} > \frac{p-ps}{1-ps} \quad (28)$$

which may be interpreted (figure 6) as showing that the lowest valid value for p_s is that for which the solid line is tangent to the curve for C/C_1 (broken lines). Referring back to figure 4, it may be seen that the lowest value for p_s for which equilibrium profiles exist is the threshold value p_* corresponding to the steepest (right-hand) leg of the $c(p)$ curve in figure 4. This is an important result, in that it shows that the behaviour of this leg of the curve has a dominant influence on profile development, and underlines the reasonableness of neglecting the remaining legs of the relationship for many purposes.

Substituting equation (5) for $C(p)$ into equation (26) leads, for constant D_o , to:

$$p = 1 - \frac{Fk_*(1-p_*)}{T+Fk_*(1-p_*)} \exp \left[- \frac{\left\{ T-Fk_*p_* \right\}}{\left\{ D_1 k_* + D_o \right\}} Z \right] \quad (29)$$

If D_o is allowed to vary, then the term

$$\frac{Z}{D_1 k_* + D_o}$$

in the exponent should be rescaled to

$$\frac{Z}{D_o(Z) + D_1 k_*}$$

Thus for example, if the rate of organic mixing declines exponentially with depth, at rate:

$$D_o \exp(-Z/Z_2)$$

then Z in equation (29) should be replaced by

$$Z = \left\{ Z + Z_2 \ln \left[\frac{D_1 k_* + D_o \exp(-Z/Z_2)}{D_1 k_* + D_o} \right] \right\} \frac{\left\{ D_1 k_* + D_o \right\}}{\left\{ D_1 k_* \right\}} \quad (30)$$

Figure 7 shows examples of the equilibrium profiles produced for declining organic mixing with $D_o = 10^{-2} \text{ m}^2/\text{y}$ in (a), $10^{-4} \text{ m}^2/\text{y}$ in (b) and with no organic mixing, and hence constant diffusion in (c). It may be seen that the effect of organic mixing is to shift the weathering front down into the soil, with a zone of rather constant composition near the surface. The importance of the surface mixing zone in producing soils of any appreciable depth cannot be overstressed. If F , the (constant) percolation rate is increased, and T , the denudation rate increased in proportion, then the degree of surface weathering (p_s) is unaltered, but the equilibrium profile becomes shallower as indicated by the broken lines in figure 7(a), most notably in the neighbourhood of the weathering front.

If the dog-legs of figure 4 are included in the expression for solute concentration, then, the effect for the broken-line curve is to provide a different solution for $p < 0.55$ which is matched to the original solution with a very slight discontinuity in slope gradient d (arrowed). Thus for lower portion of the curve:

$$C(p) = 17.1 \times 10^{-6} (p-0.2)$$

giving the overall solution shown for one example by the dotted line in figure 7a. It may be seen that the effect is to shorten the constant composition zone in this example by about 0.9 m. The discontinuity is more marked in the absence of organic mixing (fig 7c:dotted line).

It should be noted that all the profiles discussed so far are associated with marked mixing in the upper part of the profile, so that profiles are not characterised by the formation of saprolite or volumetric weathering features in which the detailed structures of the parent material are preserved intact. To look for such features it is necessary to turn to non-equilibrium profiles which will be seen to show zones of almost constant composition which extend well below the zone of effective near-surface mixing.

Figures 8, 9 and 10 show computed profiles which show the evolution of the weathering profiles towards either equilibrium or indefinite accumulation, as well as incorporating the hydrological routines described in section 2 above to calculate the percolation rate. The parameter values used in these trial runs are listed in the table below.

TABLE 1

All runs	$\left\{ \begin{array}{l} k_* = 4 \times 10^{-5} \\ p_* = 0.4 \end{array} \right.$	$D_1 = .01 m^2/y$
		F_o (into bedrock) = 0.1 m/y

Run 1	Figure 8	$D_o = 10^{-4}(e^{-z/0.5}) m^2/y$	$\lambda = 1 m/y$	$F_{max} = 1 m/y$	(Eqn. (2))
Run 2	Figure 9	$D_o = 10^{-4}(e^{-z/0.5}) m^2/y$	$\lambda = 10 m/y$	$F_{max} = 0.5 m/y$	
Run 3	Figure 10	$D_o = 10^{-2}e^{(-z/0.5)} m^2/y$	$\lambda = 10 m/y$	$F_{max} = 0.5 m/y$	

In the cases when soils approach equilibrium, profiles bear some comparison with the equilibrium profiles of figure 7 above, but differ because of the varying percolation rates with depth. Runs 1 and 2 differ only in this respect, which can be seen to have some influence on the detailed profile form predicted. At depths of 2.7m (Runs 1 and 2) or 5.0m (Runs 3) the organic mixing falls to levels as low as that due to ionic mixing, so that in deep and old soils, mostly those of indefinite

accumulation, a constant composition zone of in-situ saprolite weathering develops, usually after periods of 10^5 or more years, which is thought to be comparable to many deeply weathered lateritic profiles.

In these weathering profiles, the overall rate of solution is a dependent variable. For shallow soils, total percolation is limited by the permeability, so that solution rates increase with soil depth. For deep soils, the flattening of the weathering front with age diverts more of the subsurface flow to depths with more weathered (lower p) soils which are less soluble, so that there is some reduction in solution rates beyond a peak value. This effect is less marked in runs 2 and 3 because of their higher permeability and lower total subsurface flow. Figure 11 illustrates this effect at three stages in the evolution of a soil in Run 1, where it may be seen that the total solution loss (the area enclosed within the broken-line) is greatest at the intermediate time, when percolation is at a maximum and the weathering front still covers a relatively broad band of depths.

Figure 12 generalises this effect for Run 1, showing that, for given climatic and parent material controls, solution loss can be expressed largely in terms of total soil deficit (i.e. soil depth) over a wide range of denudation rates. Some divergencies may however be seen as profiles approach their equilibrium forms. The curves for Runs 2 and 3, for more permeable soils, show a less marked decline from the peak solution rate as soils thicken. The very restricted cross-overs shown in figures 8 to 10 suggest that, for the set of condition within each run of the model, there is almost a single valued expression relating the total deficit, w to the degree of surface weathering, ps, with minimal ambiguity (figure 13). Relationships such as those shown in figure 12 and 13, if treated as empirical inputs to slope development models, provide one way of modelling soil catenas which can be linked implicitly with the models described here.

13. The Organic profile

Equilibrium organic profiles may readily be derived from equation (17) above in the case where $D_0(Z)$ and $F(2)$ are held constant with depth giving:

$$E = \frac{Fk \cdot E}{Fk + D_0/Z_1} \exp(-Z/Z_1) \quad (31)$$

$$\text{where } Z_1 = \sqrt{\frac{2D}{F^2 k^2 + 4D_0}} - Fk$$

It was suggested above that k should be zero, so that z , then simplifies to D_0/Z_1 , and is the characteristic scale depth for organic material in the soil. For the carbon profile, the total can be made up from the sum of such curves with different values of β according to the durability of compounds, so that there will be a long tail to the distribution corresponding to the lignified components. In this paper however, the profiles are drawn for the inorganic constituents of the organic soil, so that a single value of β is thought sufficient.

As for the weathering profiles, the results from several trial runs are presented, and common run numbers indicate compatible sets of values as follows:

Table 2 $D_0 (\text{m}^2/\text{y})$ Z₀= scale depth
 for rooting (m) Rainfall (m/y)

Run 2	$10^{-4} e^{-(Z/0.5)}$	0.5	1.0
Run 3	$10^{-2} e^{-(Z/0.5)}$	0.5	1.0
Run 4	10^{-2}	0.5	1.0
Run 5	$10^{-4} e^{-(Z/0.2)}$	0.5	1.0
Run 6	$10^{-2} e^{-(Z/.5)}$	0.1	0.5

All Runs: E = Actual evapotranspiration = 0.5 m/y

(R-E) = Residual percolation = 0.5 m/y

β = rate of nutrient uptake = 0.06 kg/m³

α = rate of leaf-fall = 0.04/y

ω = rate of decomposition = 0.2/y

k = solubility of constituents in organic soil = 4×10^{-5}

(zero for organic soil)

The profiles in figure 14 (a) to (d) have been drawn on semi-logarithmic graph paper to emphasise the slight departures from the exponential decay form of equation (31) in consequence of the decline in mixing rate with depth. It may be seen that the value at the surface is dominant, especially where this value is low as in runs 2 and 5((a)and (d)). The approach to equilibrium is only roughly an exponential decay to saturation values, with time scale very similar to that for the vegetation; 25 years ($= \frac{1}{\beta_0}$) in these examples.

14. The inorganic profile associated with nutrient cycling

The third component of the model is controlled by equation (19) or (22) above. Once more it may be solved for constant $F(2)$, $D_0(2)$ at equilibrium. For the case of zero solubility within the organic matter, and for the case of equal solubilities in organic and inorganic profiles:

$$T = \frac{Ez_0}{D_0 + Fkz_0} e^{-z/z_0} - \frac{Ez_1}{D_0 + Fkz_1} e^{-z/z_1} \quad (32)$$

This solution has constant sign for all z ; and is all-positive if $z_0 > z_1$; all negative if $z_0 < z_1$. The eventual state of the profile is one of overall enrichment if the roots (z_0) penetrate beneath the organic soil zone (z_1), and overall depletion if the roots lie within the organic soil. The former appears to be the normal strategy for humid-zone plants, typified in the extreme case by tropical rain forest. In such equilibrium humid soil nutrients from decomposing organic soil are leached down the profile and caught by the root system which brings them back to the surface again, with what is generally considered to be a very high efficiency, and often in the effective absence of other nutrients within reasonable reach of root penetration. It is suggested that the reverse strategy, with roots less deep than the organic soil and a deficit at equilibrium, is a similar strategy for plants in very base-rich environments, including the adaptation of non-phreatophyte arid plants. In this case, the deficit zone provides a zone of lower base status which the plants can better tolerate. Examples of equilibrium profiles are given in figure 15. It will be seen that for likely values of D_0 , F , k the organic diffusion terms generally dominate the result, so that $\frac{\partial T}{\partial z} > 0$ and $\frac{\partial z}{\partial T} > 0$ at the surface. Examples are drawn with $z_0 > z_1$ in all cases. If z_0 and z_1 are exchanged, profiles are the same but with opposite sign.

During the evolution of the inorganic profile, a net deficit builds up which is eventually equivalent to the total quantity of nutrients stored in the plants and organic soil. In the 'humid' case ($T > 0$ at equilibrium), this deficit delays the onset of equilibrium for what may be a long time,

and is certainly longer than the times needed for the vegetation and organic soil to equilibriate. Figure 16 shows trial runs with the values given in Table 2 for Runs 2 to 5. It may be seen that the extent of the deficit built up varies considerably, and appears to depend strongly on the effective depth to which organic mixing descends. In Run 4 (Fig 16C) organic mixing (D_o) is constant with depth, so that it has an infinite scale depth (Z_2), and in particular its scale depth is greater than that for either rooting (Z_o) or the organic soil (Z_1). In this case ($Z_2 > Z_o > Z_1$), it may be seen that the deficit zone is relatively short lived, being completely removed after 350 years and with an approach to equilibrium in under 1,000 years. In runs 2 (fig. 16a) and 3 (fig. 16b) $Z_2 = Z_o > Z_1$ and the deficit zones persists for periods in excess of 500 and 2,000 years respectively, with equilibrium approached over periods 3 to 4 times greater. In run 5 (fig. 16d), with $Z_o > Z_2 > Z_1$, the deficit is very much greater and is still increasing in magnitude after several thousands of years, leading to equilibrium only over periods of 10^4 to 10^5 years. It can be shown that in the absence of leaching, no equilibrium is reached unless Z_2 is not the largest of the three scale depths. Even the low leaching rates will however allow convergence on an equilibrium profile eventually.

It is argued that the processes incorporated in this model are showing a tendency towards B - horizon formation, although it is recognized that several other sets of factors are also acting, of which the most important are thought to be variations in solubility down the profile and mechanical washing of clay particles (lessivage). The partial pressure of CO_2 in the soil rises in response to decomposition of organic carbon in the soil. The CO_2 produced diffuses out, setting up a partial pressure gradient which approaches a saturation level (for a soil of uniform gas-filled porosity) with a scale depth equal to that for the organic soil (Z_1). Reduction in CO_2 partial pressure goes in part to increase soil pH and in part to reduce solubility of bases. Similarly changes in concentration of bases, if large in absolute terms (including the background value from the weathering profile) tend to decrease solubility for $T > 0$ and increase it for $T < 0$. Other solubility changes may be due to complexing, which will tend to allow mobilization of Al and Fe in the deficit regions of figure 16. These changes will, overall, reduce the magnitude of the deficit zone to a small extent. Lessivage will tend to be most effective in the organic soil zone where percolation velocities are highest, so that there is likely to be some translocation of clays to depths of about 0.2m (Runs 2 and 5) or 1.0m (Runs 3 and 5). In the case of Runs 3 and 5, this movement is likely to reinforce the deficit zone to what may be an appreciable extent

in soils with low clay (and Al) contents.

The magnitude of the effects shown in figures 14, 15 and 16 varies with the level of uptake within the plant, so that it is greatest for bases and least for sesquioxides. The background level for bases is reduced much more by weathering than for sesquioxides, so that the combined effect is for the deficit zones shown in figure 16 to be associated with high ratios of sesquioxides to bases, as is illustrated by the following table.

Table 3: Hypothetical calculation of Sesquioxide: Base ratio

$p = 1 \text{ to } 2,500 \text{ kg/m}^3$; all values converted to concⁿ in kg/m^3 below

	Bedrock ($p = 1$)	Weathered Soil ($p = 0.75$)	Deficit zone	Surface Acc Zone
Sesquioxides	625. (25%)	600 (34%)	600 - .2	600 + 6
Bases	625 (25%)	87 (5%)	87 - 30	87 + 360
Ratio	1.0	6.9	10.4	1.4

It may readily be seen that the deficit zone is a zone showing pronounced nutrient loss and a dominance of clay forming minerals. On parent materials which initially have a low base status, the effect will be most pronounced; and it is those same soils which exhibit more humus with very shallow depths of organic mixing, so that several of the features of podsolization are predicted.

Figure 17 shows an inorganic soil for the "arid" case with $Z_0 < Z_1$, and rainfall = evapotranspiration = 0.5 m/y. The overall deficit associated with the biomass in this case encourages rapid convergence on an equilibrium, so that there is not the same tendency for a local surplus to become large. However, many arid soils are already almost saturated with calcium carbonate, so that even a small surplus may be sufficient to encourage its precipitation as a caliche layer below the rooting zone. It should be noted that Z_1 (organic soil depth) is unlikely to be greater than Z_2 (organic mixing depth) since the latter is one major control on the former. The case ($Z_1 > Z_2, Z_0$) is therefore inherently implausible so that the conditions for accentuating the surplus zone (analogous to the accentuated deficit zone in Run 5) are not normally met.

It is suggested that the relative values of the three scale depths provide a basis for a categorization of major soil types (figure 18).

(i) If $Z_0 < Z_1$, a deficit zone will develop in the surface inorganic soil, only partly offset by the nutrient in the organic soil. A weak surplus zone may develop for a while, associated with calichification in strongly

calcified parent material. If $Z_0 > Z_1$, a surplus zone will eventually develop in the surface inorganic soil. (ii) If $Z_2 > Z_0 > Z_1$, a transient B-horizon will develop but will never be a major feature of the profile. (iii) If $Z_2 > Z_1$, Z_0 then the B-horizon will be much more marked and long-lasting.

15. Links to hillslope modelling

These slope models may be incorporated in a downslope sequence linked by a slope sediment transport model in one of two ways. The first is to derive appropriate versions of figures 12 and 13 above for the correct climatic and parent material values. These diagrams express the surface degree of weathering, ps and the rate of solute denudation, U in terms of the total soil deficit, w . The second method is to incorporate the model more directly in a model with two independent space dimension; of distance downslope and depth within the soil.

The first method implies solution of the continuity equations:

$$-\frac{\partial y}{\partial t} = U + \frac{\partial S}{\partial x} \quad (33)$$

$$+\frac{\partial w}{\partial t} = U - \frac{\partial}{\partial x} \left\{ \frac{Sps}{1-ps} \right\} \quad (34)$$

where y is elevation (in rock equivalent units),

x is distance downslope

S is downslope mechanical sediment transport,
and other variables are as previously defined.

By expressing U and ps in terms of soil deficit, w via the equivalents of figures 12 and 13; and by expressing S in terms of distance from divide (x) and slope gradient ($-\partial y/\partial x$), these equations can in principle be solved numerically, leading to slope profiles with associated soils.

The second method for combining slope and soil models requires an explicit hydrological model which routes sub-surface flow downslope and supersedes equation (2) above. The two models may then be formally linked, calculating flows and solute transport within the two-dimensional soil layer on the hillside. Sediment transport may also then be modelled in terms of calculated overland flows etc. Plainly this type of combined model has theoretical advantages, although at a considerable cost in added complexity. The present paper does not attempt to pursue the link, but merely suggests that this soil model is not too complex to allow combination.

16. Conclusion

This paper clearly draws on a considerable body of other work, but has not been fully referenced because the levels of, for instance, geochemistry,

which are drawn on are not above a level of broad acceptance. The contribution made here is to attempt to simplify the geochemical changes to a simple empirical relationship, and to build this relationship into an integrated model. The questions raised by this attempt relate to the validity of assumptions and simplifications made; to the selection of dominant processes to include; and to the extent to which the model leads to a range of realistic profile forms.

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Figure 1 : Definition sketch for soil profile terminology

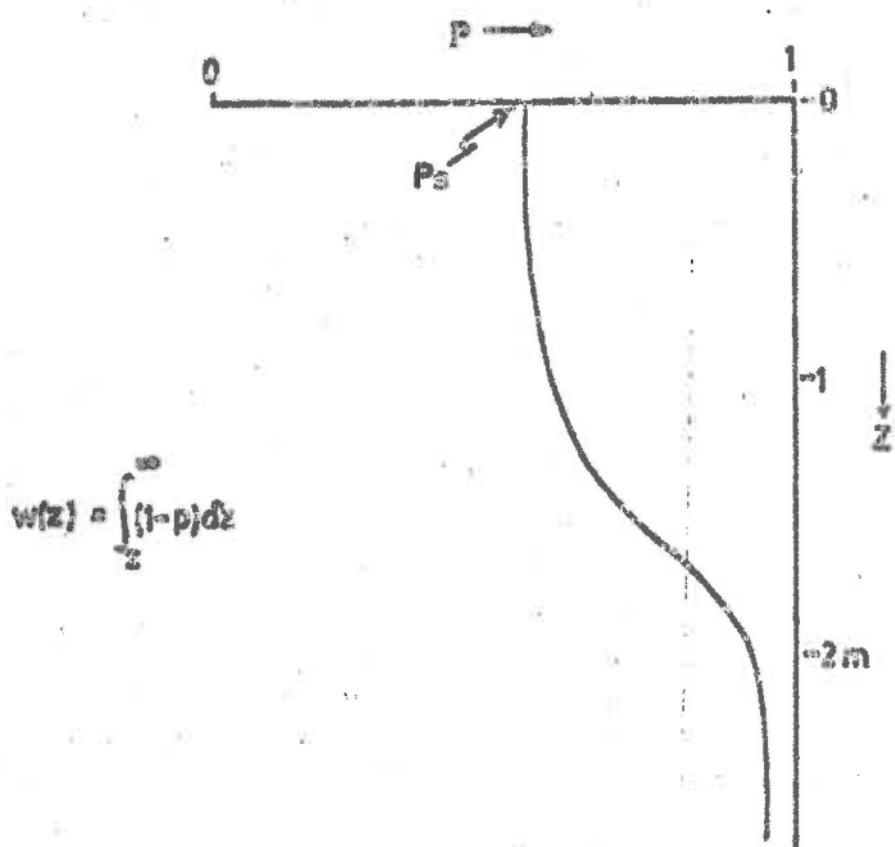
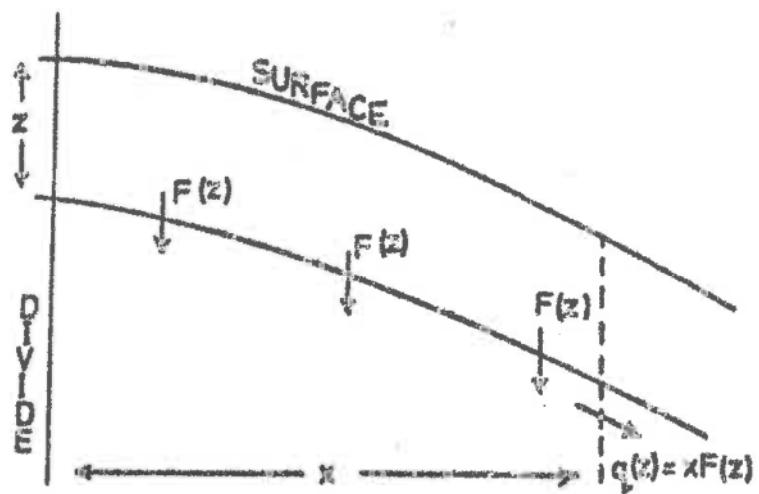


Figure 2 : Percolation at depth z on a convex hillside



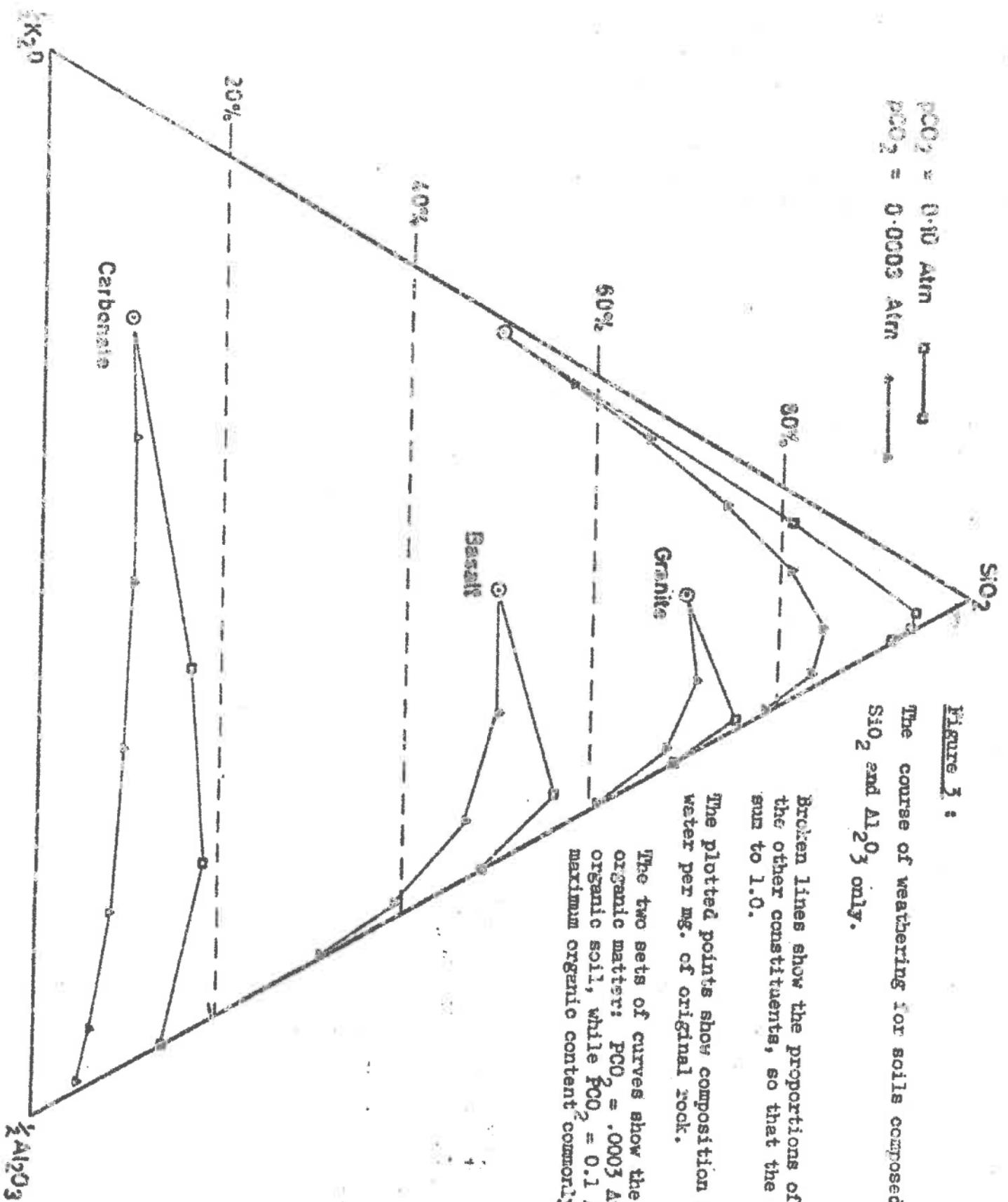


Figure 2 :

$pCO_2 = 0.10 \text{ Atm}$ $\text{---} \circ \text{---} \circ$
 $pCO_2 = 0.0003 \text{ Atm}$ $\text{---} \square \text{---} \square$

The course of weathering for soils composed of mixtures of K_2O , SiO_2 and Al_2O_3 only.

Broken lines show the proportions of SiO_2 ; and similarly for the other constituents, so that the three proportions always sum to 1.0.

The plotted points show composition at intervals of 0.041 water per mg. of original rock.

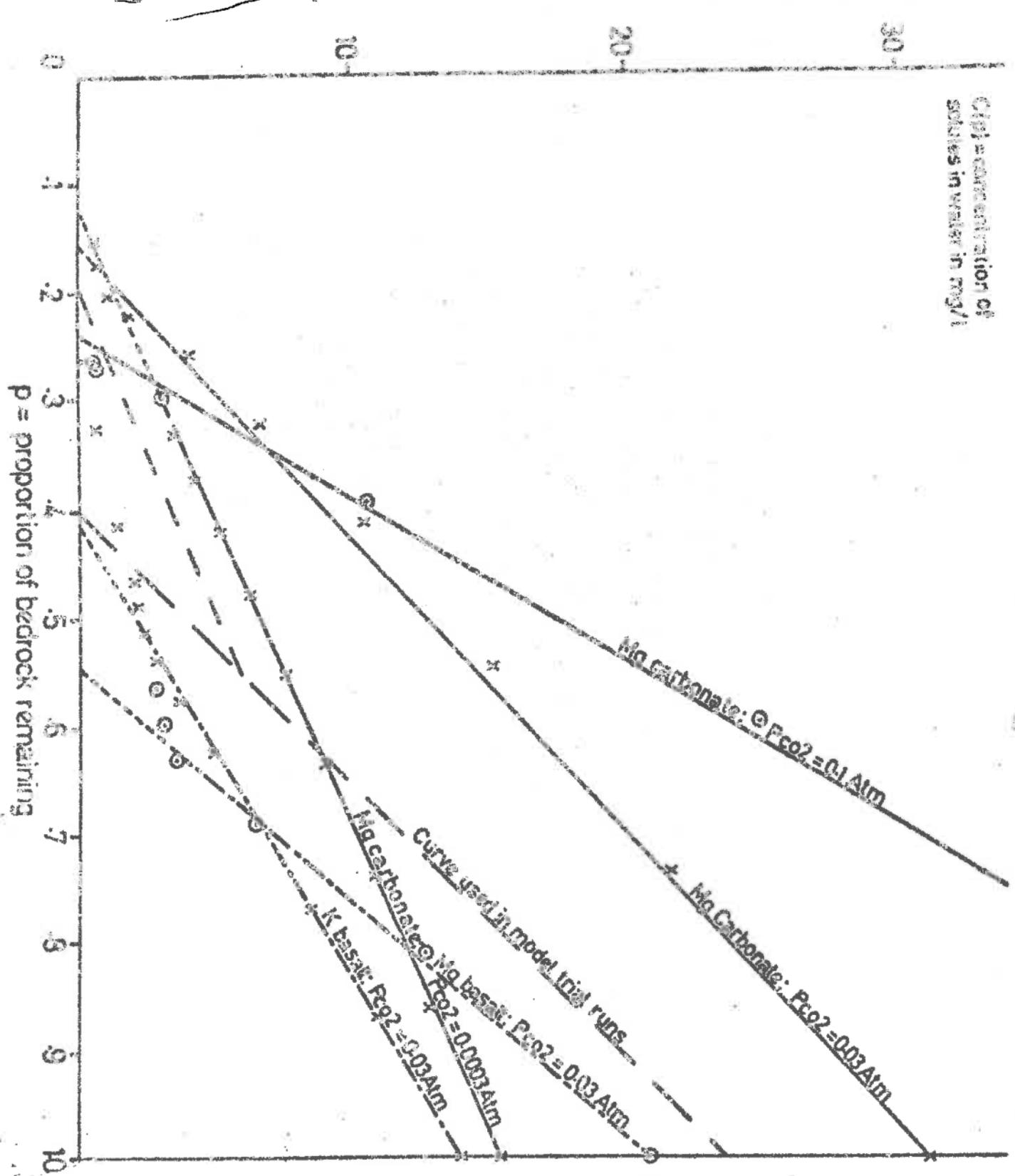
The two sets of curves show the influence of soil organic matter: $pCO_2 = .0003 \text{ atm}$ corresponds to no organic soil, while $pCO_2 = 0.1 \text{ atm}$ corresponds to maximum organic content commonly found.

Granite

Basalt

Carbonate

$C(p)$ = concentration of
solute in water in mg/l



The relationship between
total solute concentration,
 $C(p)$ and total proportion
remaining p .

Figure 4 :

Figure 5 : Definition sketch for leaching loss

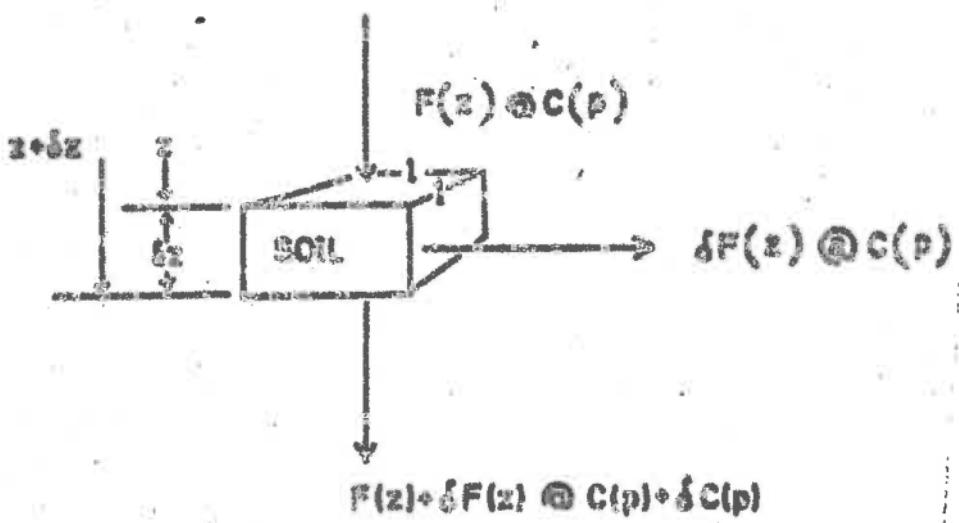


Figure 6 : Hypothetical relationship between minimum ps value and the C(y) curve.

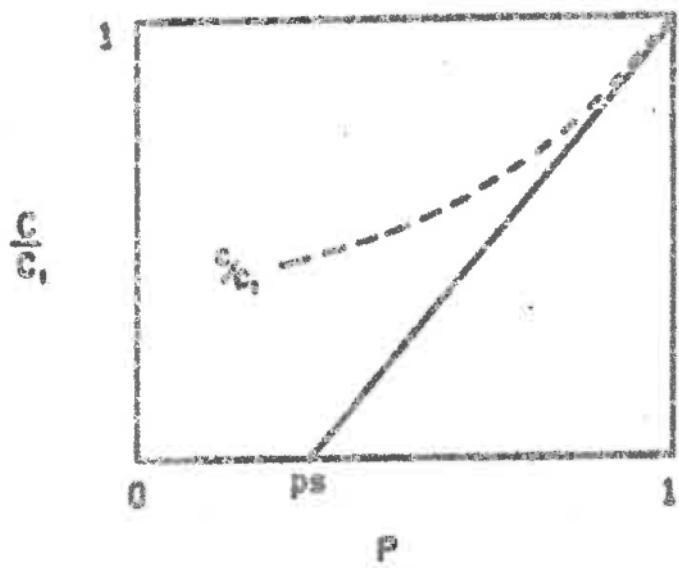
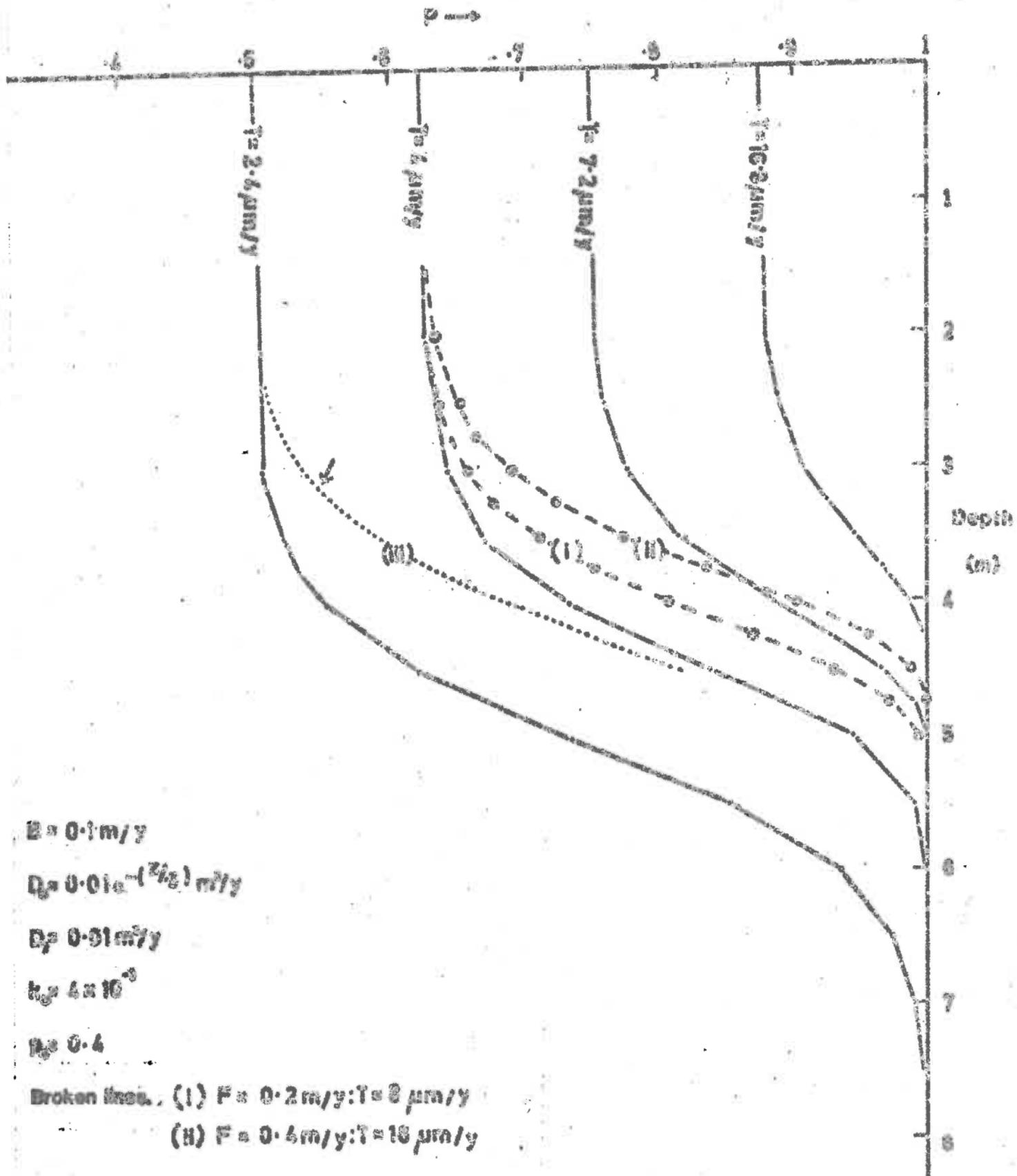
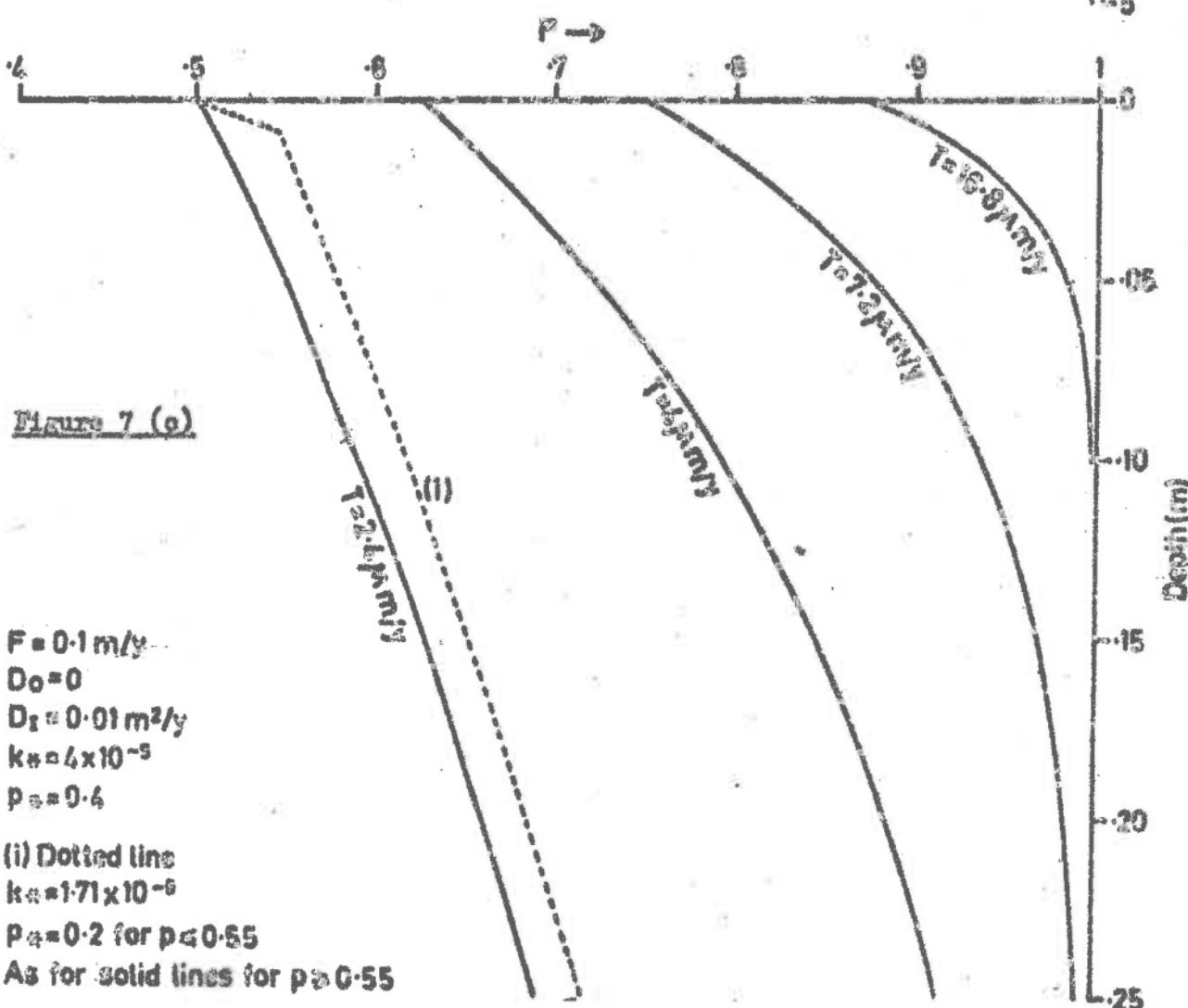
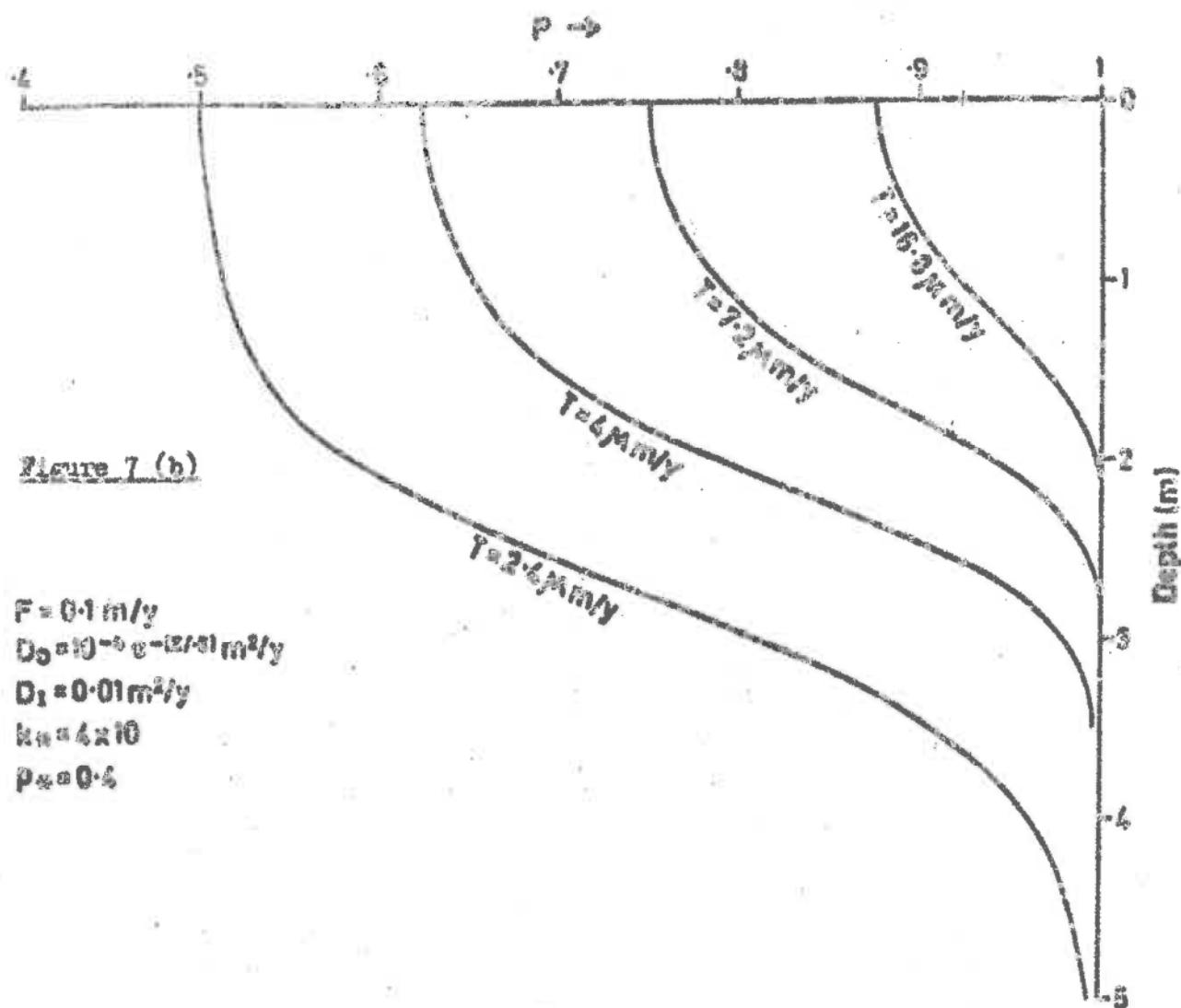
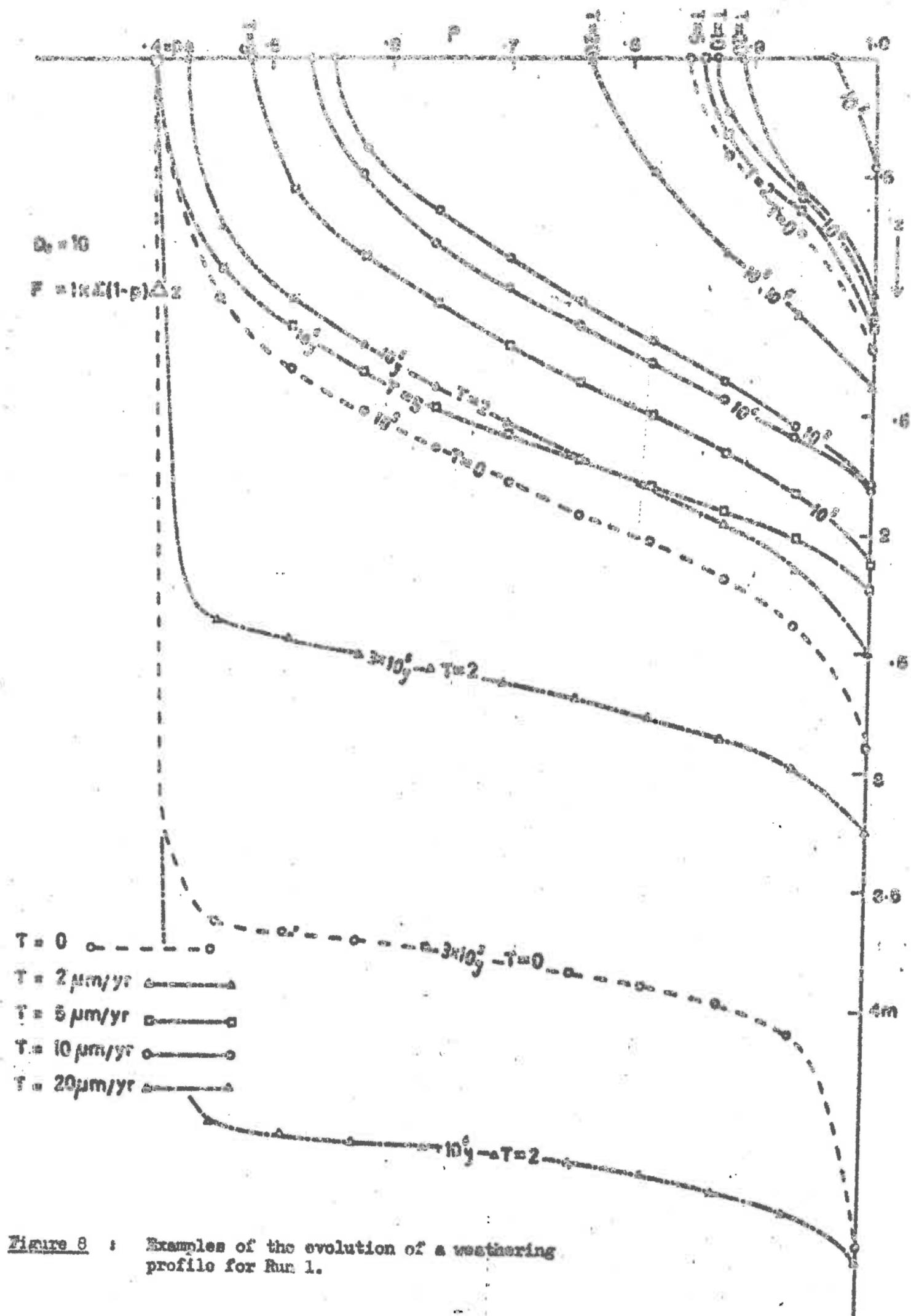


Figure 7(a) : Examples of equilibrium weathering profiles computed from equations (29) and (30).







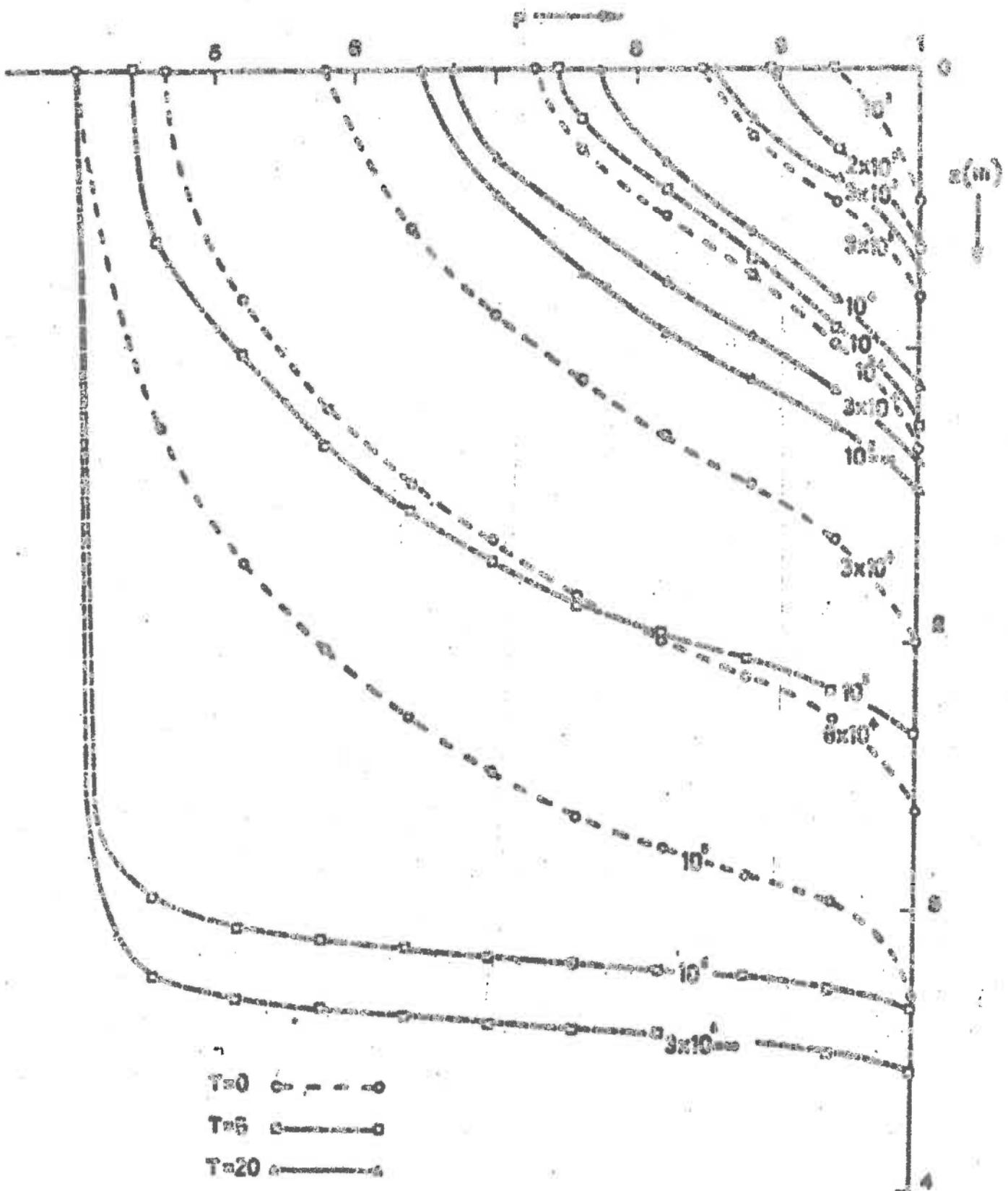
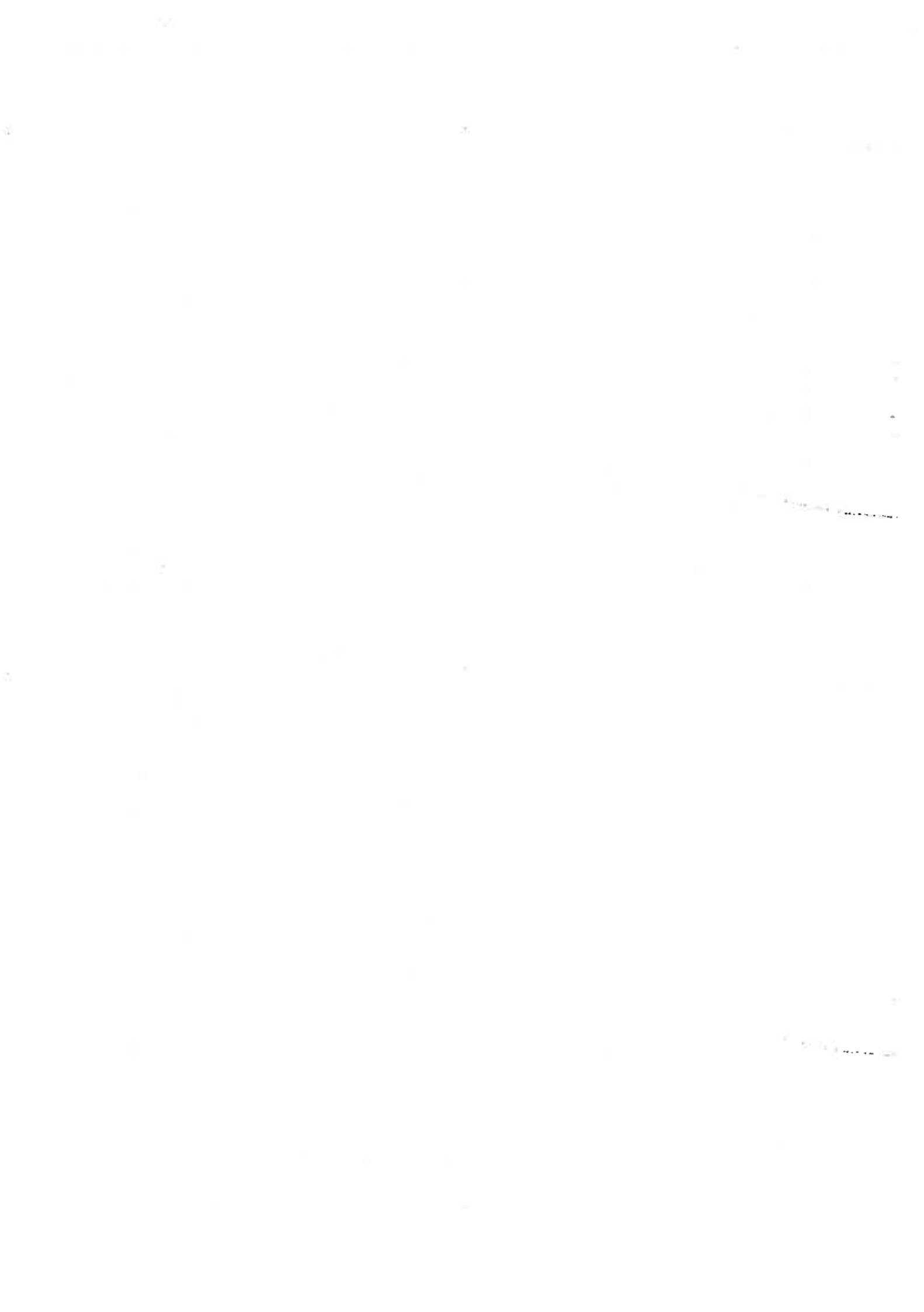
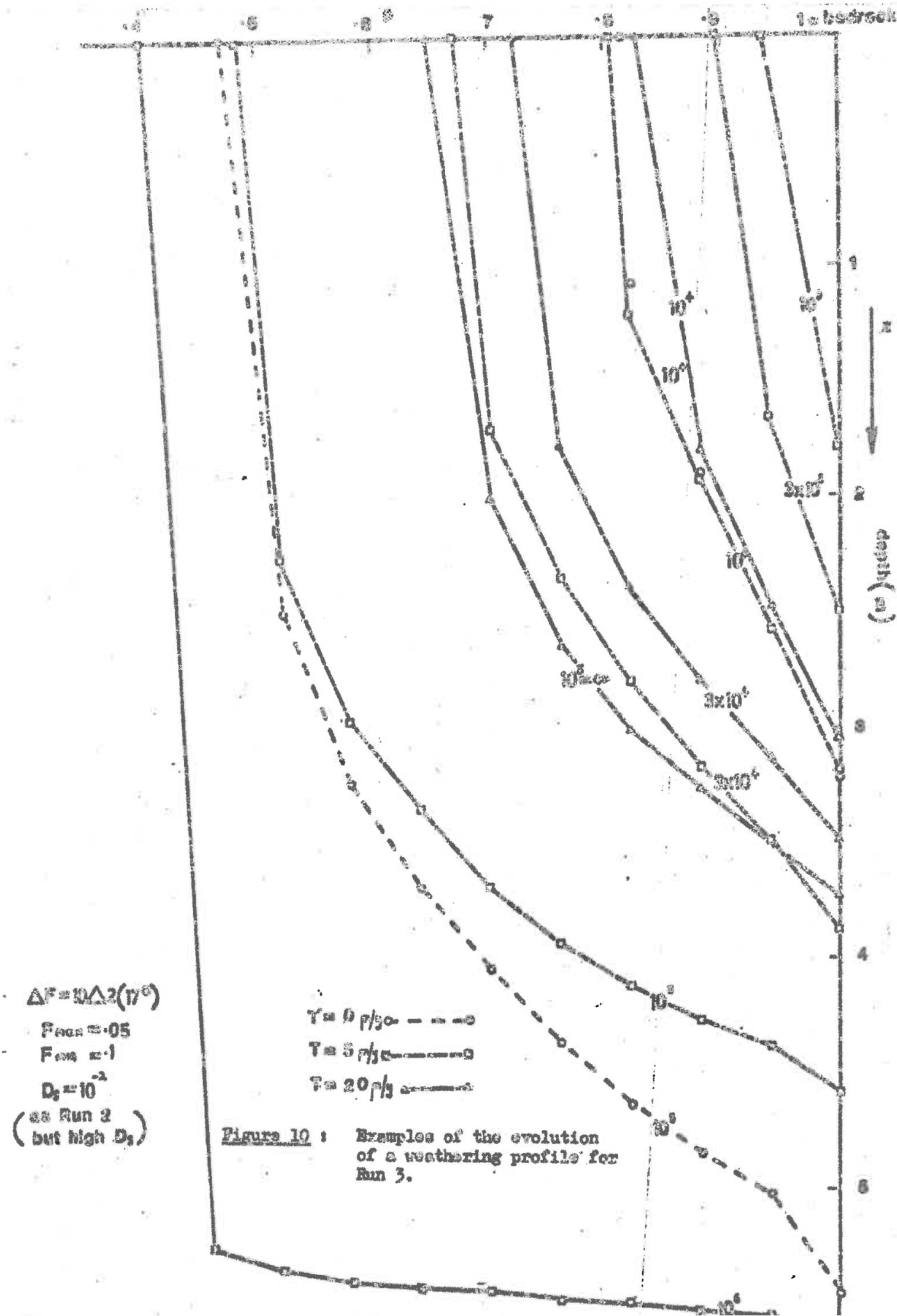


Figure 9 : Examples of the evolution of a weathering profile for Run 2.





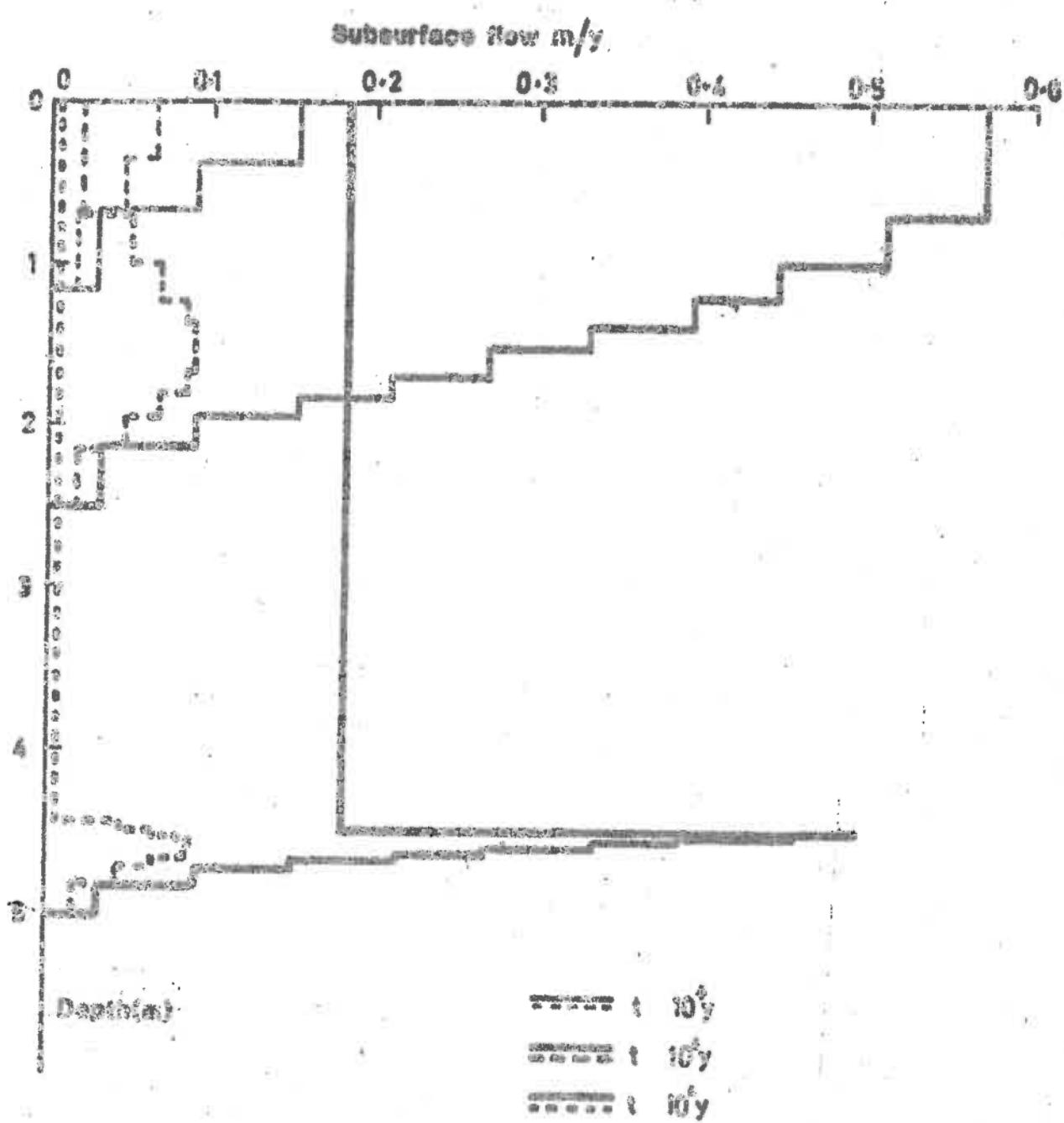


Figure 11 : Modelled distribution of sub-surface flow (solid lines) and solution loss (broken lines) at three stages in Run 1 with $T = 2$ per/yr.

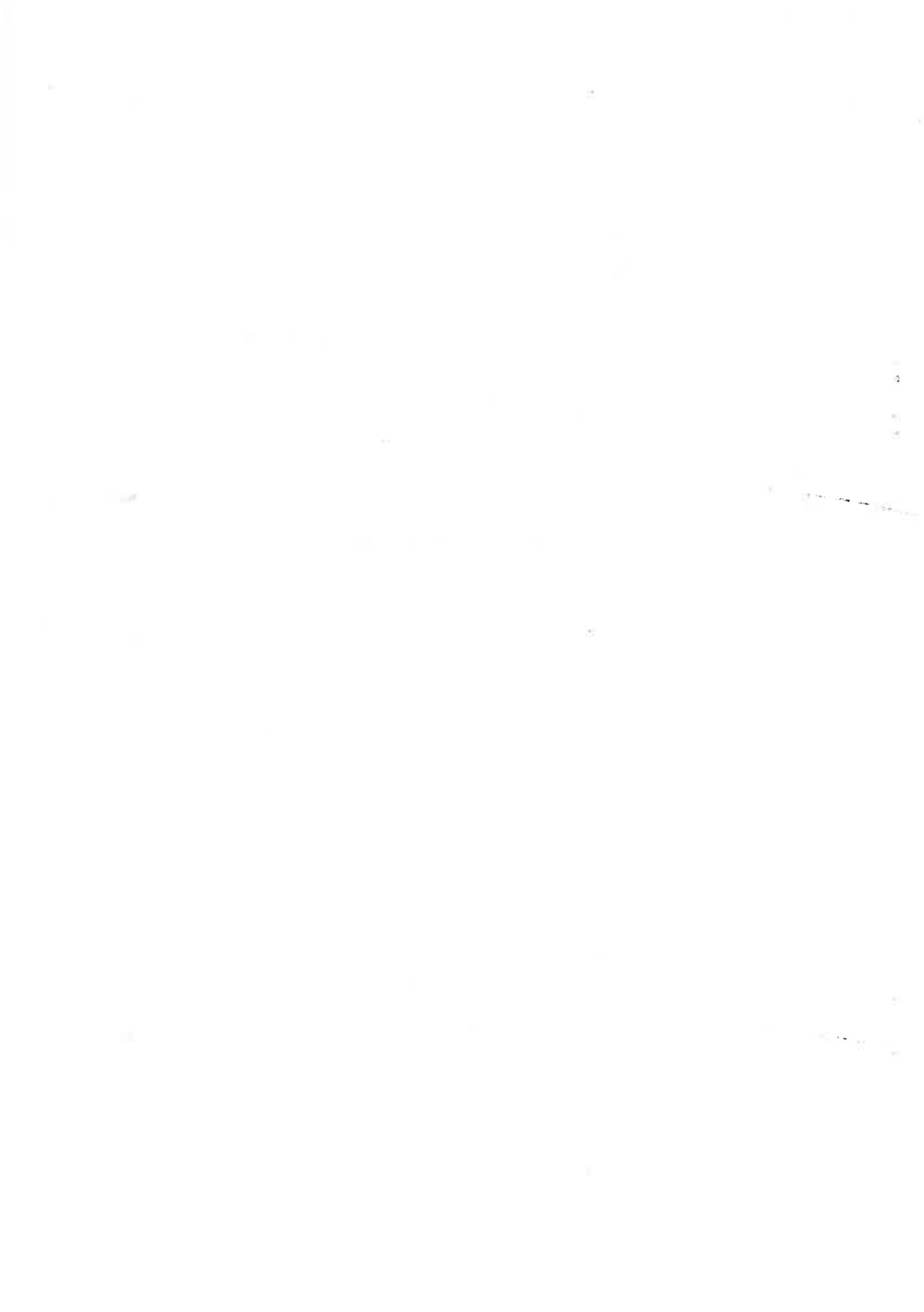
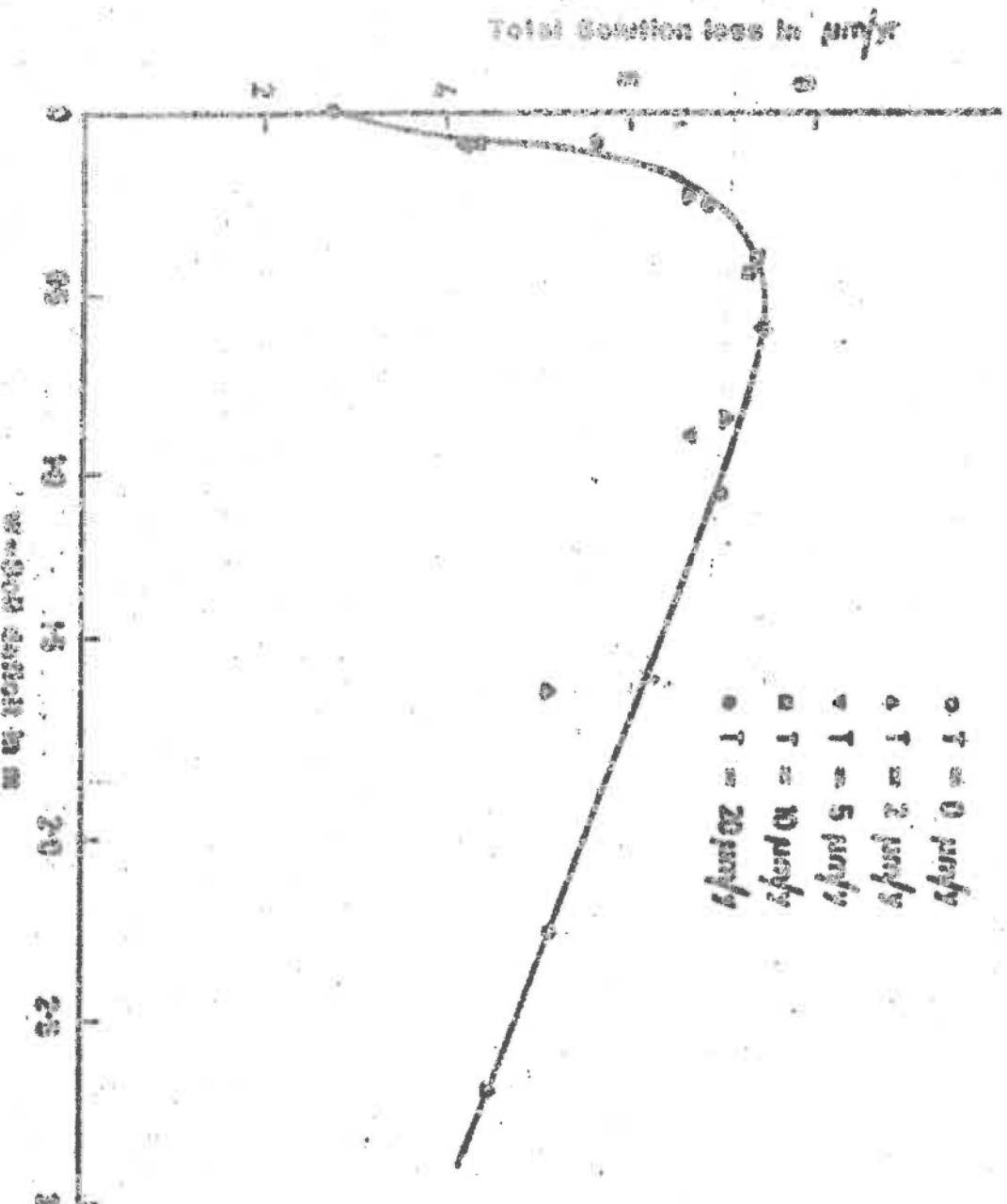


Figure 12

Total solution loss as a function of soil depth, for a range of mechanical denudation rates, for Run 1 conditions.



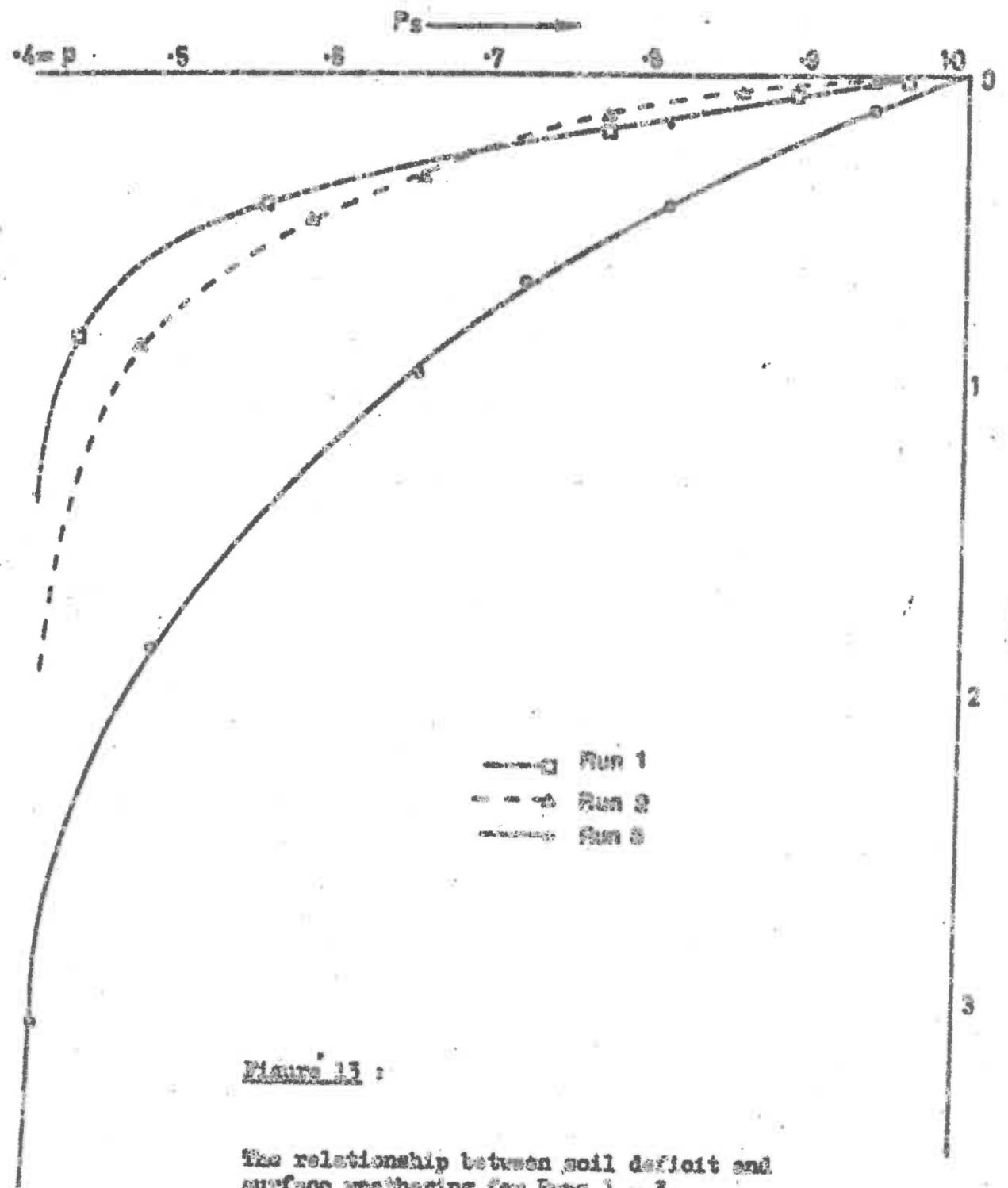


Figure 13:

The relationship between soil deficit and surface weathering for Runs 1 - 3.



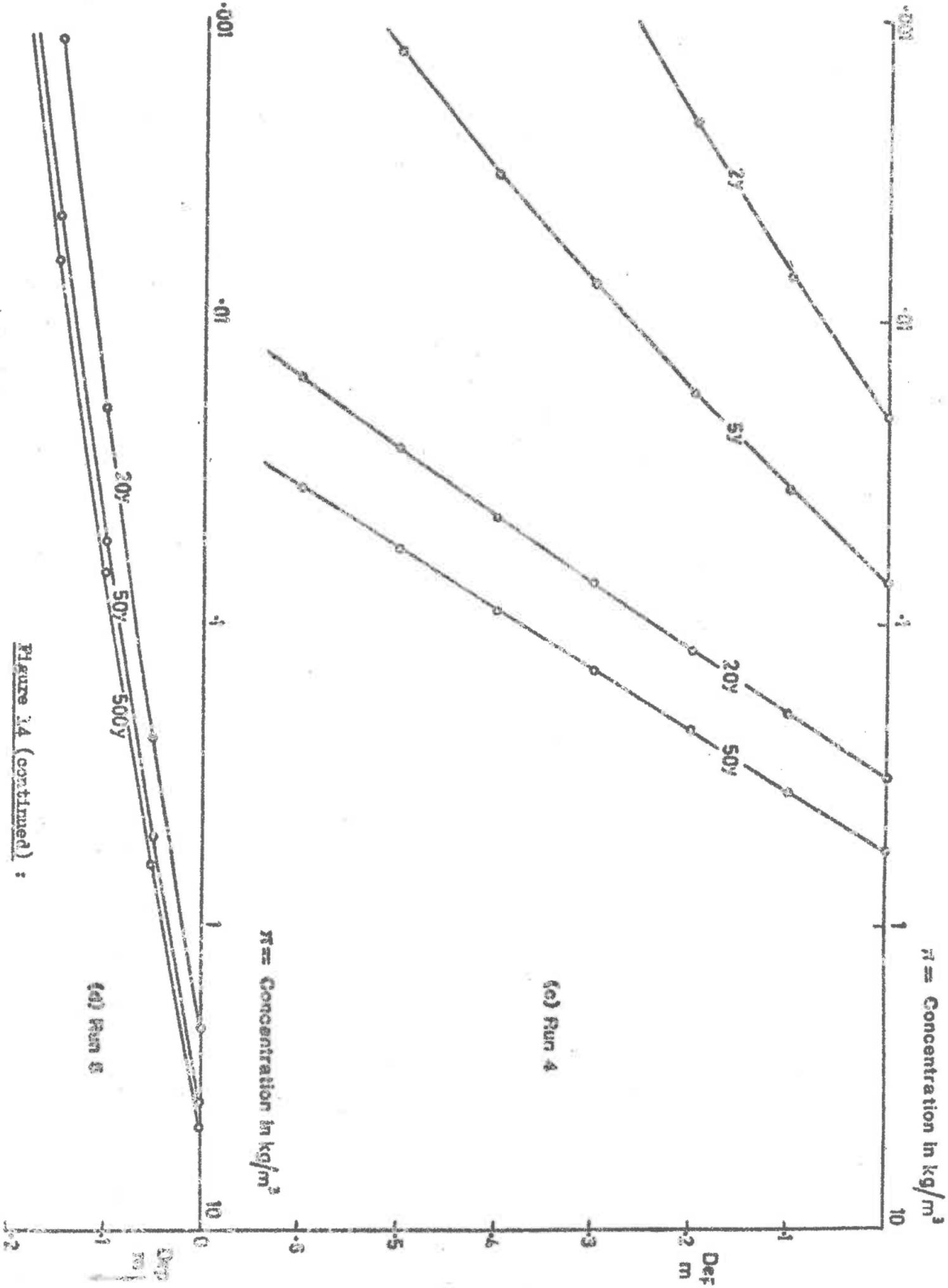
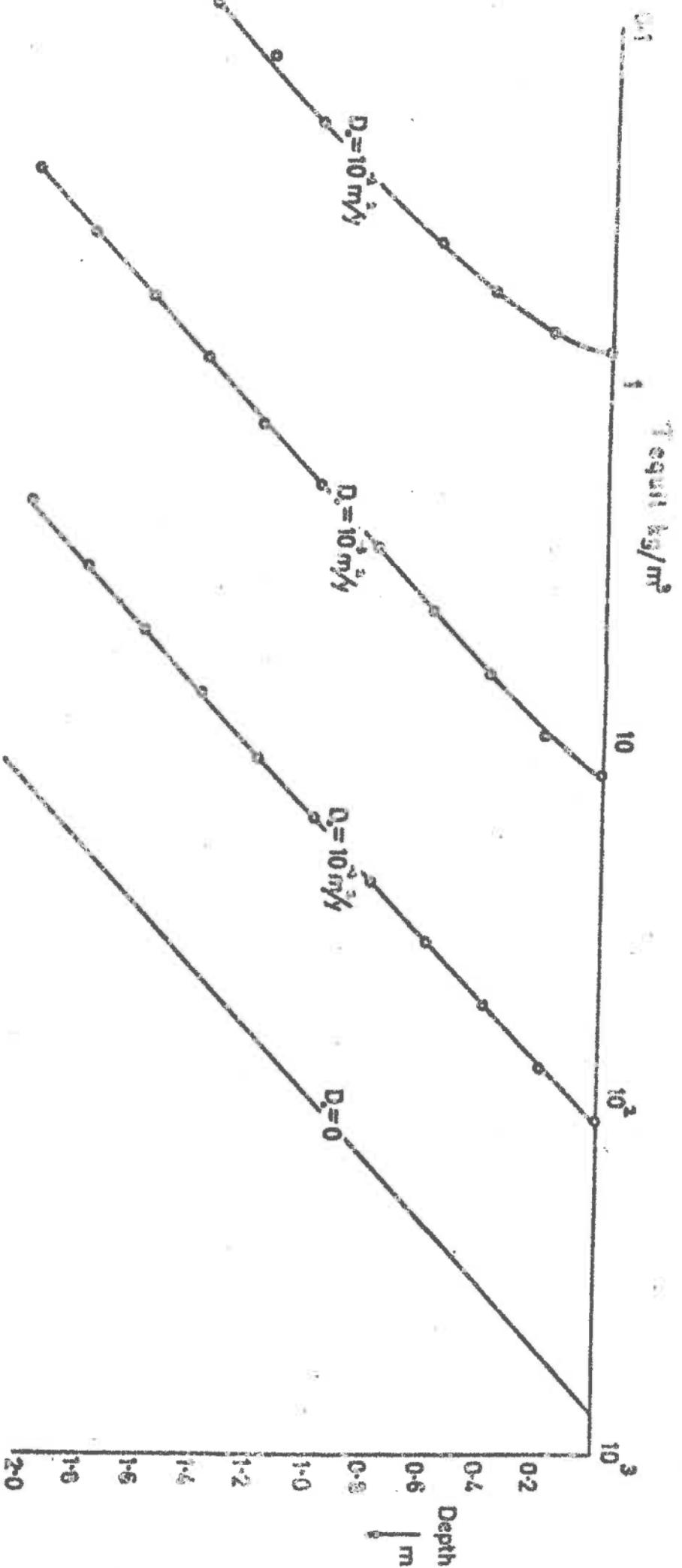


Figure 34 (continued) :

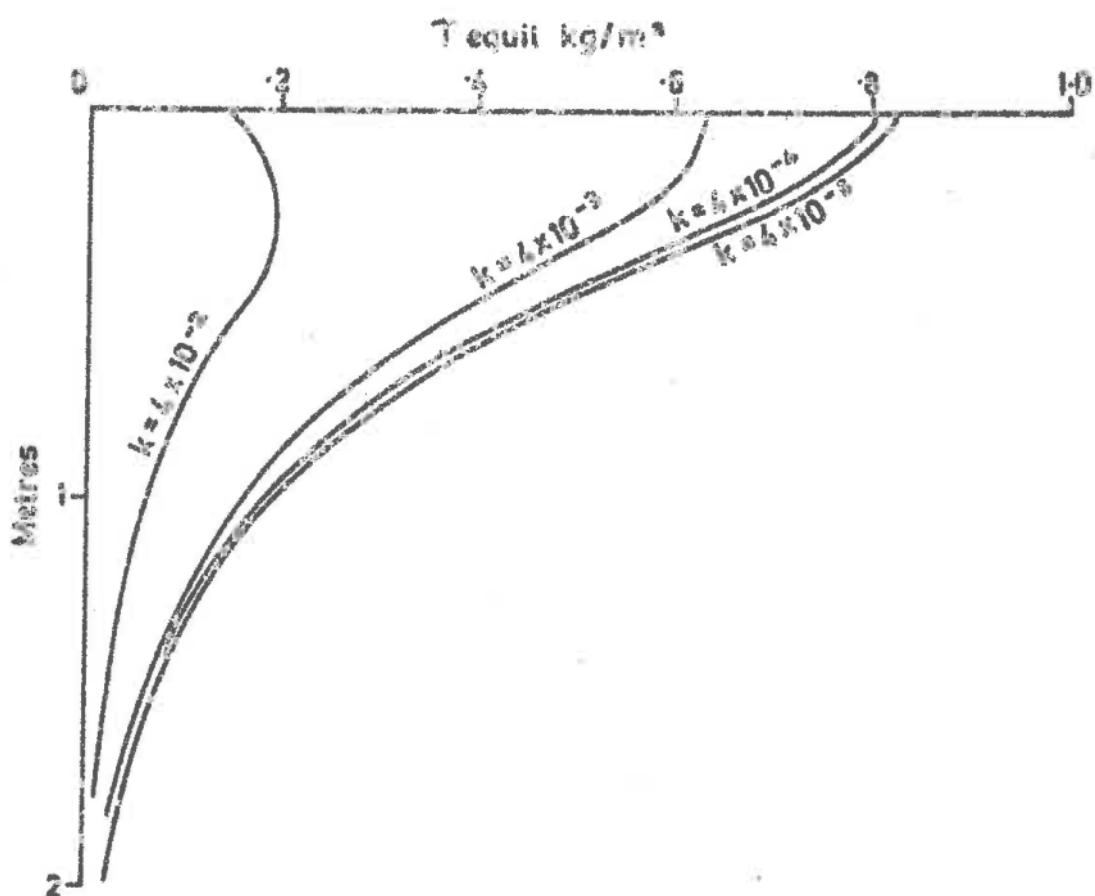


$$Z_0 = 0.5 \text{ m} \quad F = 1 \text{ m/s} \quad \beta = 0.2 \text{ /s}$$

(a) $k = 4 \times 10^{-3}$ $Z_i = D_r / \beta$ as D_r varies (semi-log plot, $Z_i > Z_0$ in all cases)

FIGURE 15 : Equilibrium inorganic soil profiles associated with nutrient cycling

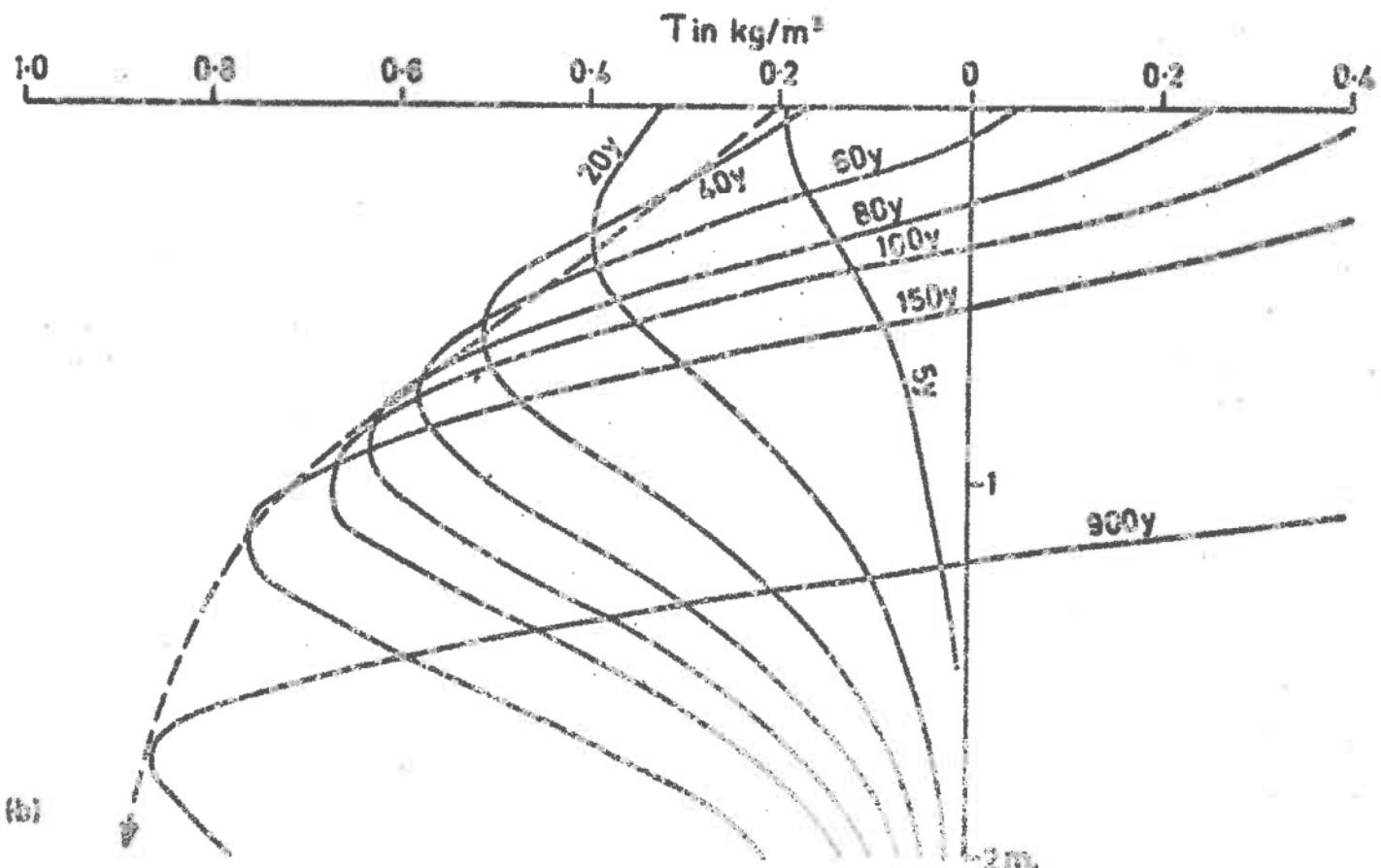
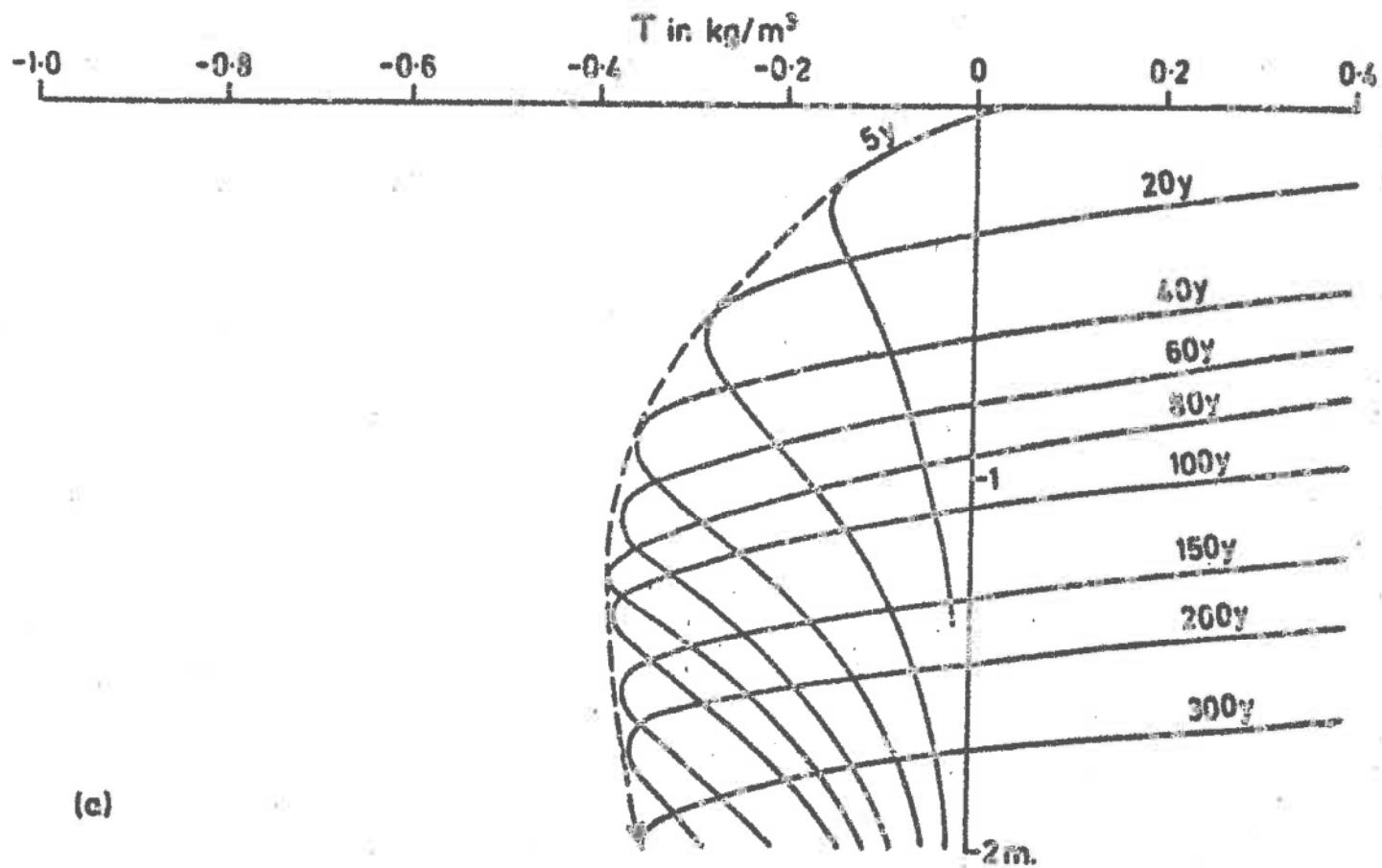
Figure 15 (continued)



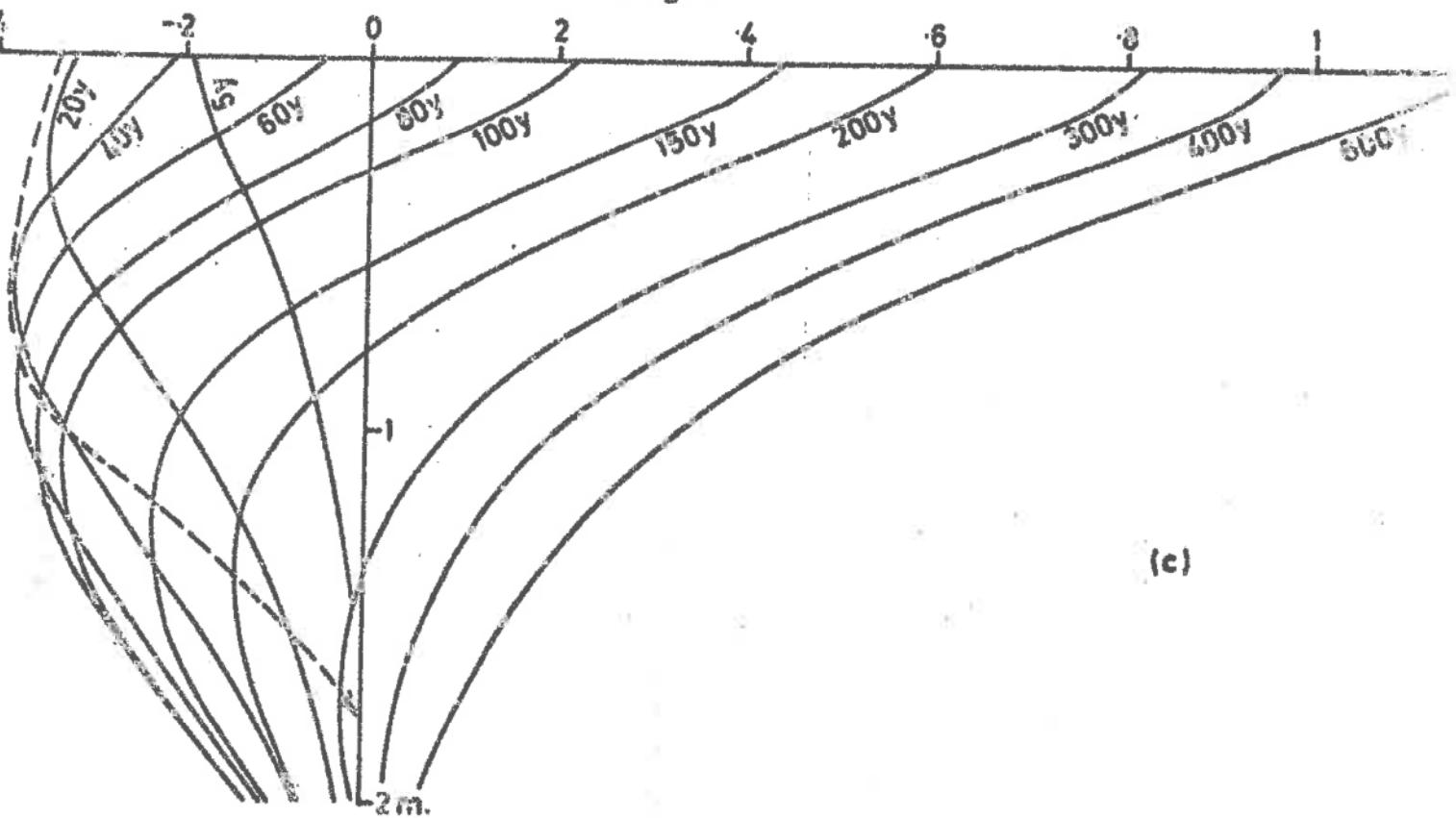
- (b) $D_0 = 0.01 \text{ m}^2/\text{y}$
 $Z_1 = \sqrt{D_0/\theta} = 224 \text{ m}$
as k varies (emphasizing influence of percolation as k increases)

Figure 16: Inorganic Profiles associated with nutrient cycling

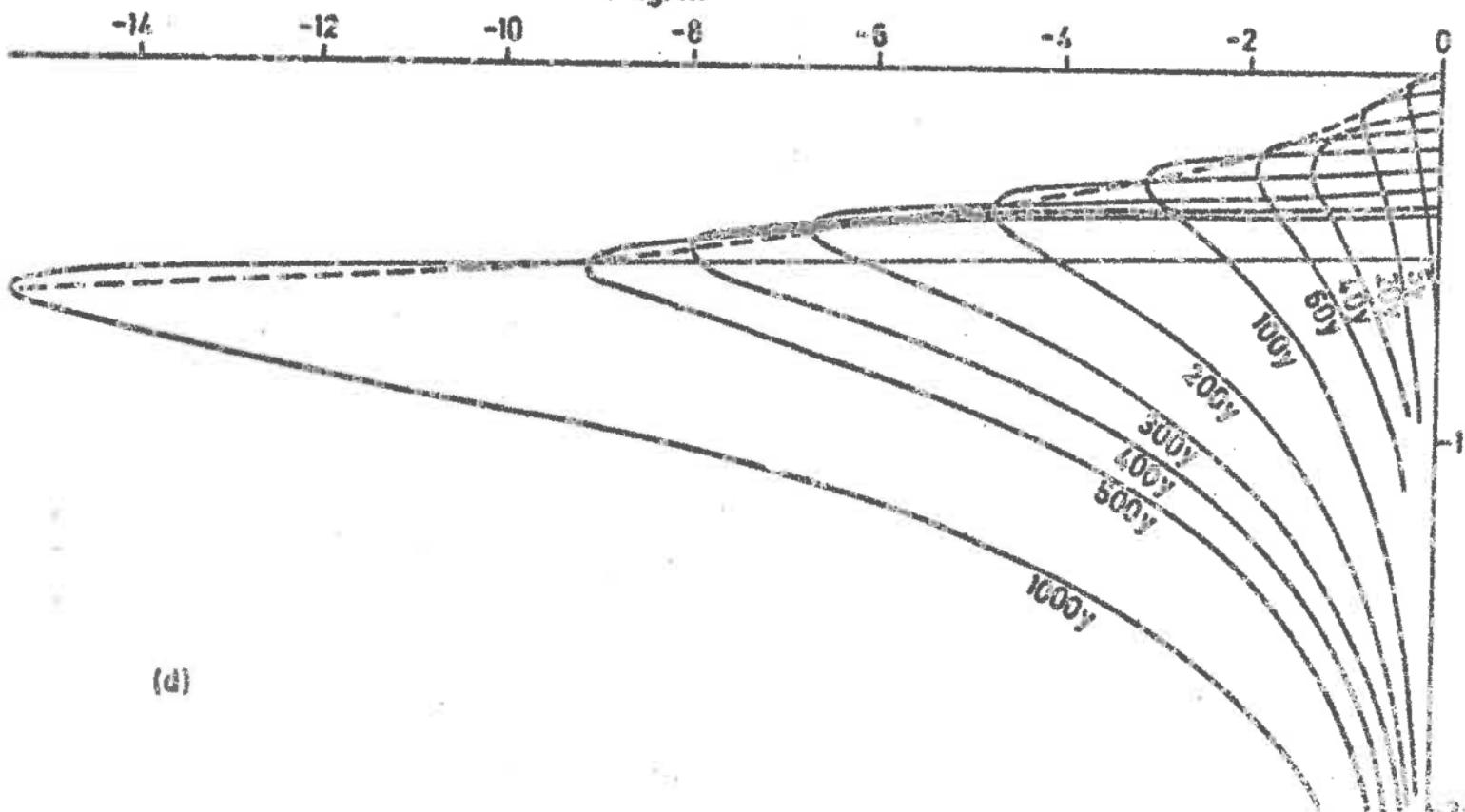
Evolution over time (for $Z_0 > Z_1$ cases)

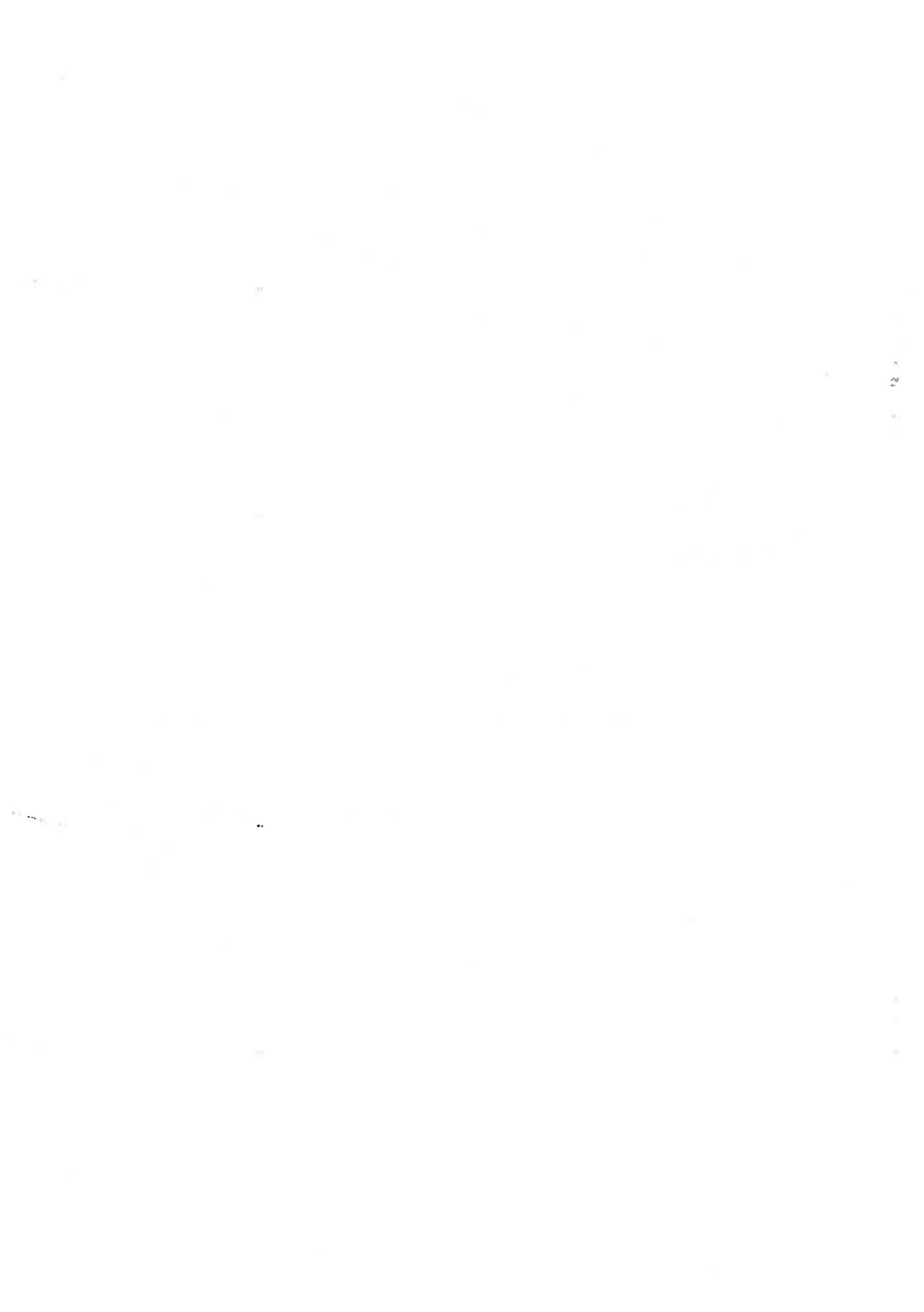


T in kg/m^3



T in kg/m^3





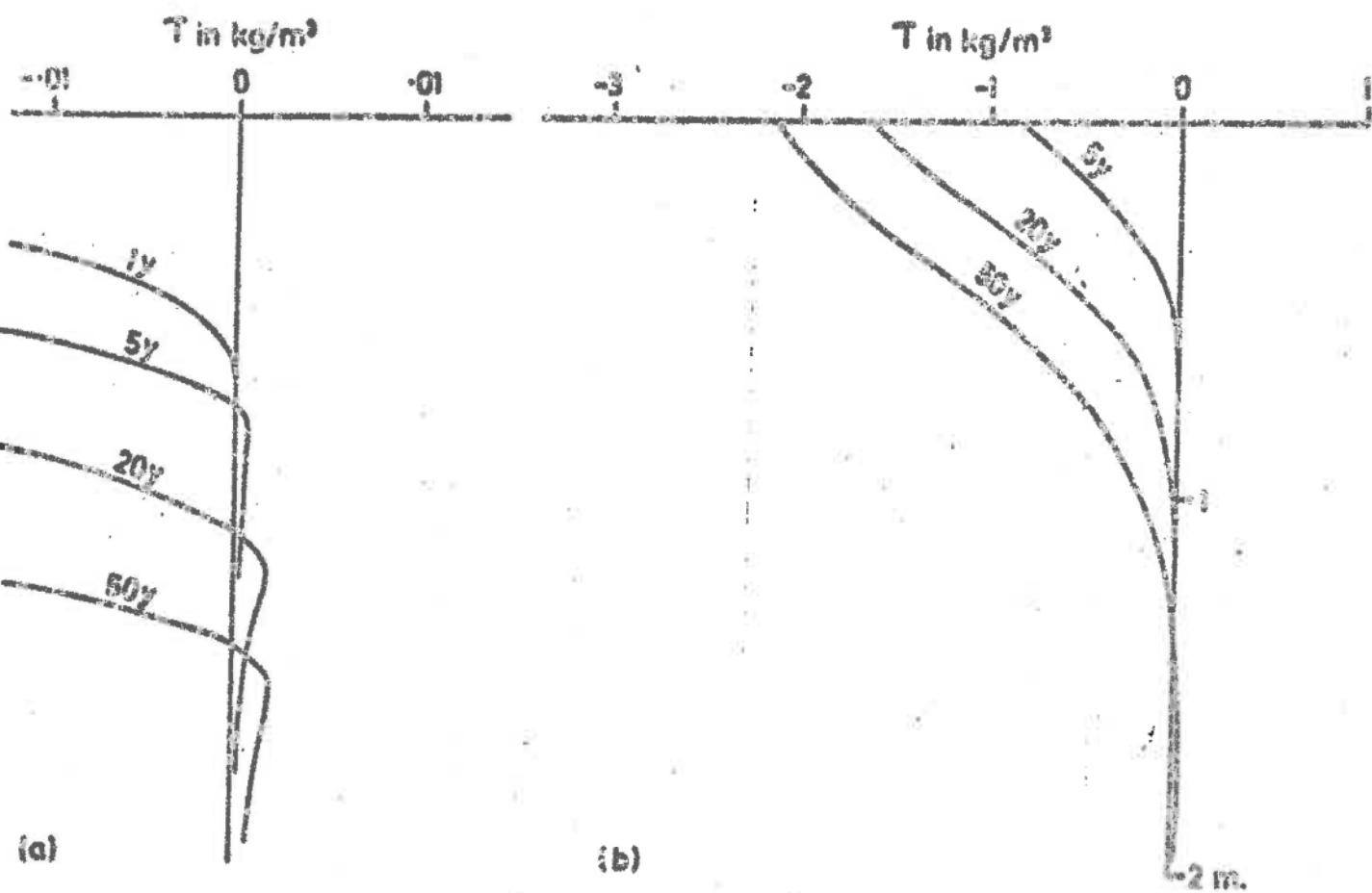


Figure 17 : Inorganic soil associated with nutrient cycling

Run 6 : an example of an "arid" evolution ($S_0 < Z_1$)

(a) enlarged scale showing detail of surplus zone
 (b) overall profile at reduced scale for same data.

Figure 18 : Basis for classifying soil horizontal types on basis of scale depths

