# EROSION AND NITROGEN LOSS FROM FOREST SOILS IN RELATION TO SOIL STRUCTURE

by

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## Declarations

This thesis contains no material that has been accepted for a degree or diploma by the University of Tasmania or any other institution. To the best of my knowledge and belief this thesis contains no material previously published or written by another person except where due acknowledgment is made in the text of the thesis.

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***Abstract***

Soil erosion by rain and runoff causes the loss of potential productivity from natural landscapes and the deterioration of the quality of surface water. Most studies of soil erosion have focused on agricultural soils, but little work has been done on forest soils. Understanding the causes and effects of soil erosion and the resultant loss of nutrients is essential to the development of sustainable soil management practices.

This study considers the soil properties that influence the susceptibility of soils to erosion, as well as the external factors that promote erosion. The combined effects of these factors on soil and nitrogen (N) loss were examined. Experiments were conducted in field and laboratory plots using simulated rainfall. In the laboratory, three forest soils of contrasting structure were subjected to simulated rainfall of constant intensity. For each soil, there were four erosion treatments. These permitted comparison of erosion due to varying kinetic energy (KE) of rain at a constant slope and varying slope at a constant KE, and due to the presence or absence of a drying cycle.

Erosion of soil and size-distribution of sediment were strongly influenced by the mechanical stability of soil aggregates. Indices based on the Mean Weight Diameter of uneroded soils provided a good indication of the susceptibility of soils to erosion and the characteristics of sediment.

As direct measurement of erosion is expensive, information on erodibility can be useful in prediction of erosion. The possibility of inferring soil erodibility from physical properties of soil was explored with the erosion model GUEST (Griffith University Erosion Systems Template). This model estimates parameters that describe

erodibility due to rainfall- or runoff-only erosion processes. Results showed dependence of sediment concentration on soil strength. However, the relationship between erodibility parameters and soil strength indicated that information on other physical properties may be also required in prediction of erosion.

The bonding strength of soil aggregates provides a good measure of the susceptibility of soils to aggregate breakdown and erosion. The process of aggregate breakdown was studied through the application of a known amount of disruptive energy to soil-water suspensions using ultrasound. A simple dynamic model was used to estimate the rate of aggregate breakdown and dispersion from the data on size-distribution of soil aggregates. The rates of aggregate breakdown for various soils were well related with their mechanical stability and the erosion rates. It was shown that this method was sufficiently robust to indicate the mechanical stability of soils and erodibility.

The loss of N due to erosion was directly related with the amount of soil loss. Irrespective of the soils and erosion treatments, the N-concentration of sediment was found to be greater than that of the uneroded soil. Production of N-enriched sediment presumably arose from a combination of uneven distribution of N within aggregates, aggregate breakdown including raindrop stripping, abundance of organic matter and residue, and deposition, rather than due to a single mechanism and/ or specific soil condition promoting nutrient enrichment of sediment. A strong dependence of measured N-loss with soil loss indicated that N-loss can be predicted from soil loss, without requiring information about N in soils or sediment.

## Acknowledgments

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Andrew Gibbons and Linda Ballard were very helpful with their technical support and competence. Mike Laffan introduced me to the Tasmanian soils, their distribution and classification, David Wright of the Department of Primary Industries (Launceston) kindly provided the rainfall simulator; Prof. Michael Clark (Department of Biochemistry) facilitated the use of the ultrasonic equipment; Brian Denton built some useful equipment to use with the rainfall simulator; Dr. S. Raine provided useful discussions related to soil dispersion by ultrasound.

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And finally, the most important of all, I wish to dedicate this thesis to my best friend, companion, and source of motivation, my husband Nuno Borralho, who was always there, no matter what. And to my son, Tomás who still does not know what a PhD means, except that it requires lots of work and little time to play.

### ***Preface***

The Chapter 4 of this thesis was previously published as: Teixeira, P.C., and Misra, R.K, 1997, Erosion and sediment characteristics of cultivated forest soils as affected by the mechanical stability of aggregates, *Catena*, 30, 119-134. This paper was then adapted to accomplish the style and form of this thesis.

## TABLE OF CONTENTS

### *List of Symbols*

| | |
|---|---|
| $\alpha_1$ | intercept of regression equation between $S_N$ and $S_L$ |
| $\alpha_2$ | slope of regression equation between $S_N$ and $S_L$ |
| $\beta$ | erodibility parameter (dimensionless) |
| $\theta$ | volumetric water content (%, by volume) |
| $\rho$ | density of water ( kg $m^{-3}$) |
| $\sigma$ | wet-density of soil (kg $m^{-3}$) |
| $\tau$ | shear stress associated with runoff water (Pa) |
| $\tau_a$ | shear strength after erosion (kPa) |
| $\tau_b$ | shear strength before erosion (kPa) |
| $\phi$ | depositability of soil (m $s^{-1}$) |
| $\Omega$ | stream power (W $m^{-2}$) |
| $\Omega_0$ | threshold stream power (W $m^{-2}$) |
| $a$ | rainfall detachability (kg $m^{-3}$) |
| A | area ($m^2$) |
| $a_d$ | rainfall re-detachability (kg $m^{-3}$) |
| ASM | antecedent soil moisture (%, by weight) |
| $a_t$ | slope of linear regression (°C $s^{-1}$) between change in temperature ($\Delta$T, °C) and duration of sonification ($\Delta$t, s) |
| $b$ | intercept of linear regression (°C) between change in temperature ($\Delta$T, °C) and duration of sonification ($\Delta$t, s) |
| BD | bulk density (Mg $m^{-3}$) |
| $c$ | sediment concentration (kg $m^3$) |
| $\bar{c}$ | mean sediment concentration (kg $m^{-3}$) |
| $\overline{c_t}$ | sediment concentration at transport limit due to runoff-only processes (kg $m^{-3}$) |
| $c_d$ | sediment concentration due to rainfall-only processes (kg $m^{-3}$) |
| $c_{eq}$ | sediment concentration at equilibrium (kg $m^{-3}$) |
| $C_L$ | rate of soil loss (kg $m^{-2}$ $min^{-1}$) |
| $C_R$ | concentration ratio of nitrogen (dimensionless) |
| $c_s$ | sediment concentration at the source limit (kg $m^{-3}$) |
| $\mathrm{c_s}$ | specific heat of air-dry soil (J $g^{-1}$ °$C^{-1}$) |
| $c_t$ | sediment concentration at the transport limit (kg $m^{-3}$) |

| | |
|---|---|
| $c_v$ | specific heat of container (J $g^{-1}$ °$C^{-1}$) that holds soil-water suspension for ultrasonification |
| $c_w$ | specific heat of water (J $g^{-1}$ °$C^{-1}$) |
| d | diameter of a raindrop (m) |
| $D_c$ | extent of dispersion of clay by ultrasonification (%, by weight) |
| $d_i$ | rate of deposition (kg $m^{-2}$ $s^{-1}$) of sediment of settling velocity of class *i* |
| $D_p$ | packing density (dimensionless) |
| $D_w$ | depth of water on the surface of soil (m) |
| $E$ | total soil loss (kg $m^{-2}$) |
| $E_c$, $E_c'$ | energy (J) lost by conduction from a soil-water suspension, with undispersed and fully-dispersed soil, respectively |
| $e_{di}$ | rate of rainfall re-detachment (kg $m^{-2}$ $s^{-1}$) of sediment of settling velocity of class *i* |
| $E_h$, $E_h'$ | energy (J) used in heating a soil-water suspension, with undispersed and fully-dispersed soil, respectively |
| $e_i$ | rate of rainfall detachment (kg $m^{-2}$ $s^{-1}$) of sediment of settling velocity of class *i* |
| $E_i$ | accumulated energy (J) applied to a soil-water suspension |
| $E_R$ | measured enrichment ratio of nitrogen (dimensionless) |
| $E_R'$ | predicted enrichment ratio of nitrogen (dimensionless) |
| $E_t$, $E_t'$ | energy (J) lost by transmission of wave energy from a soil-water suspension, with undispersed and fully-dispersed soil, respectively |
| F | fraction of excess stream power |
| $f$ | efficiency with which re-entrained sediment can slide down the walls of a rill (dimensionless) |
| $F_{ij}$ | rate of aggregation/disaggregation (g $s^{-1}$), or flux of soil material from size-class i to size-class j |
| $f_{ij}$ | rate constant of flux ($s^{-1}$) of soil material from size-class i to size-class j |
| $g$ | acceleration due to gravity (m $s^{-2}$) |
| I | number of size-classes of soil with equal mass |
| $J$ | specific energy of entrainment (J $kg^{-1}$) |
| $k$ | constant (J °$C^{-1}$) that depends on the mass and specific heat of soil, water, and the container with the soil-water suspension |
| $k_d$ | constant dependent on the density of water to convert mass of water to volume |
| L | slope length (m) |
| $L_g$ | energy (J) used in disaggregation and dispersion of soil aggregates |

| | |
|---|---|
| m | dimensionless exponent to indicate laminar/ turbulent water flow |
| M | mass of raindrops (kg) |
| $m$ | slope of regression equation between $E_R$ and $S_L$ |
| MAX | size-distribution of soil (%, by weight) after 15 minutes of sonification |
| $M_s$ | mass of air-dry soil (g) |
| $M_v$ | mass of container (g) that holds a soil-water suspension for ultrasonification |
| $M_W$ | mass of water (g) |
| MWD | Mean Weight Diameter (mm or μm) |
| $MWD_d$ | MWD after dry-sieving (μm) |
| $MWD_w$ | MWD after wetting, but without dispersion of soil (μm) |
| $MWD_w'$ | MWD after wetting and dispersion of soil (μm) |
| $n$ | Manning's roughness parameter ($m^{-1/3}$ s) |
| N | number of rills per metre width of the erosion plot |
| $N_S$ | measured concentration of nitrogen in sediment (kg $kg^{-1}$) |
| $N_U$ | concentration of nitrogen in the uneroded soil (kg $kg^{-1}$) |
| OC | organic carbon (%, by weight) |
| OM | organic matter (%, by weight) |
| P | rainfall rate (mm $h^{-1}$ or $m^3$ $m^{-2}$ $s^{-1}$) |
| $P_a$ | penetrometer resistance after erosion (kPa) |
| $P_b$ | penetrometer resistance before erosion (kPa) |
| PD | particle density (Mg $m^{-3}$) |
| PSA | size-distribution of soil (%, by weight) after particle-size analysis |
| $q$ | volumetric flux of runoff water per unit width of erosion surface ($m^3$ $m^{-2}$ $s^{-1}$) |
| Q | runoff rate (m $s^{-1}$) |
| $\bar{Q}$ | mean runoff rate ($m^3$ $m^{-2}$ $s^{-1}$) |
| Q' | runoff rate (mm $h^{-1}$) |
| $Q_{eq}$ | runoff rate at equilibrium ($m^3$ $m^{-2}$ $s^{-1}$) |
| $q_{si}$ | sediment flux (kg $m^{-1}$ $s^{-1}$) |
| $r$ | rate constant of heat loss ($s^{-1}$) |
| $R_1$ | ratio of the width of the deposited layer to the wetted perimeter in a rill |
| $r_{ei}$ | rate of runoff entrainment (kg $m^{-2}$ $s^{-1}$) of sediment of settling velocity of class $i$ |
| $r_{ri}$ | rate of runoff re-entrainment (kg $m^{-2}$ $s^{-1}$) of sediment of settling velocity of class $i$ |

| | |
|---|---|
| $S$ | slope, taken as the sine of the angle of inclination |
| $s$ | proportion of sand (>20 μm) in the soil sample (%, by weight) |
| $S_D$ | susceptibility of soil to dispersion (dimensionless) |
| $S_d$ | saturation of soil (%) |
| $S_L$ | quantity of sediment produced in erosion (kg $ha^{-1}$) |
| $S_N$ | amount of nitrogen lost due to erosion (kg $ha^{-1}$) |
| $S_N'$ | predicted amount of nitrogen lost due to erosion (kg $ha^{-1}$) |
| $S_W$ | susceptibility of soil to wetting (dimensionless) |
| $S_{WD}$ | susceptibility of soil to wetting and dispersion (dimensionless) |
| t | time (s) |
| T | temperature (°C) |
| $T_0+T_1$ | temperature of soil-water suspension (°C) at time t=0 |
| $T_h$, $T_h'$ | temperature of soil-water suspension (°C) due to heating during sonification of undispersed and fully-dispersed soil, respectively |
| $u$ | intercept of regression equation between $E_R$ and $S_L$ |
| UC | uniformity coefficient (%) |
| V | velocity of runoff flow (m $s^{-1}$) |
| v | velocity of raindrops impacting the soil surface (m $s^{-1}$) |
| $v_i$ | settling velocity (m $s^{-1}$) of soil of size-class i |
| $v_L$ | volume of water (L) |
| W | rill width (m) |
| $w_i$ | proportion of the total soil sample (by weight) of the size-fraction *i*. |
| $x$ | distance downslope (m) of a bare plot of known area |
| $\overline{x_i}$ | mean diameter of aggregates (μm) of size-fraction *i*. |
| $X_i$ | amount of soil (g) in size-class i |

# Chapter 1

## INTRODUCTION

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### *1.1 BACKGROUND AND SCOPE*

In order to provide reserve of native forests and to reduce the demand for wood from natural and regrowth forests, the area established with plantations has been increasing steadily in Australia. To ensure the development of these intensively managed plantations is sustainable, the conservation of soil and water is of primary importance (Resource Assessment Commission, 1992). Minimising soil erosion is fundamental to soil conservation and the sustainable productivity of plantations, but this has received little attention from forest growers. During establishment of a plantation, the site is cleared to remove any existing vegetation and then cultivated before planting. At this stage of plantation establishment, the site is therefore essentially bare and the soil is loose. Bare and loose soils are prone to erosion by both water and wind. In some situations, planting may be delayed for up to six months after the site is ready for planting. Such a delay extends the period during which the soil is most vulnerable to erosion. The lack of aerial cover for several months after planting is also likely to extend the period during which the soil is more exposed to erosion.

The research reported in this thesis focuses on the period when plantation soils are at their highest risk of erosion, *i.e.* when they are bare and loose

It is well known that nitrogen (N) is normally the most important factor limiting the growth of trees In forest ecosystems the soil is the most important reservoir of N (Charley, 1981; Mahendrappa *et al.*, 1986). Generally, more than 98 % of the N in soils is organically bound, and most of the organic N is concentrated within the upper 60 cm of the soil profile (Bremner, 1965; Charley, 1981). Therefore, the loss of top soil due to erosion may have a significant influence on the loss of N. The importance of soil erosion in the loss of N was emphasised by White (1986) and Rose and Dalal (1988) who reported that more than 90 % of the N removed in erosion is associated with the eroded sediment, whereas 1-10 % of N is lost by leaching, in runoff, by denitrification or by volatilisation.

The experimental work reported in this thesis focused on the forest soils of Tasmania (Australia) in which plantations have been established recently. Compared with other Australian States, Tasmania has the highest proportion of its land (44 %) covered with forests (Forestry Commission of Tasmania, 1995) This report also indicated that the Tasmanian forest-based industries are the third largest employer group in the state, and the second largest export income earner, contributing substantially to the economy of the State. Due to the strong impact of forestry in the economy and ecology of Tasmania, and a lack of knowledge on the impact of plantation establishment on soil erosion, this research was undertaken to provide a baseline measurement of the erosion rates in the state as well as to provide an indication of potential erosion in temperate Australia.

## *1.2 HYPOTHESES*

The physical properties of soils are considered important factors in determining the susceptibility of a soil to erosion (erodibility), and hence its behaviour under erosive conditions (more specifically, the breakdown of soil aggregates). It was hypothesised that:

1. For a constant site (slope, slopelength and cover) and hydrologic regime (rain and runoff), the variation in erodibility and mechanical stability of soils (that affect aggregate breakdown) could be attributed to structure and strength of soils.

2. The characteristics of sediment (*i.e.*, its size-distribution) may be a factor affecting soil erodibility, the extent of aggregate breakdown, and the erosive conditions.

3. The amount of soil lost during erosion, the N status of the soil and sediment, and the size-distribution of soil and sediment were the dominant factors affecting the loss of N due to erosion.

Consequently, erodibility of soils, the mechanical stability of aggregates, and the structure and strength of soils were considered to be key elements that influence soil erosion and N loss. The relationships between these key elements are shown in Fig. 1.1, indicating the corresponding chapter of the thesis where these are described. The objectives of the work described in this thesis were to test the hypotheses summarised above.

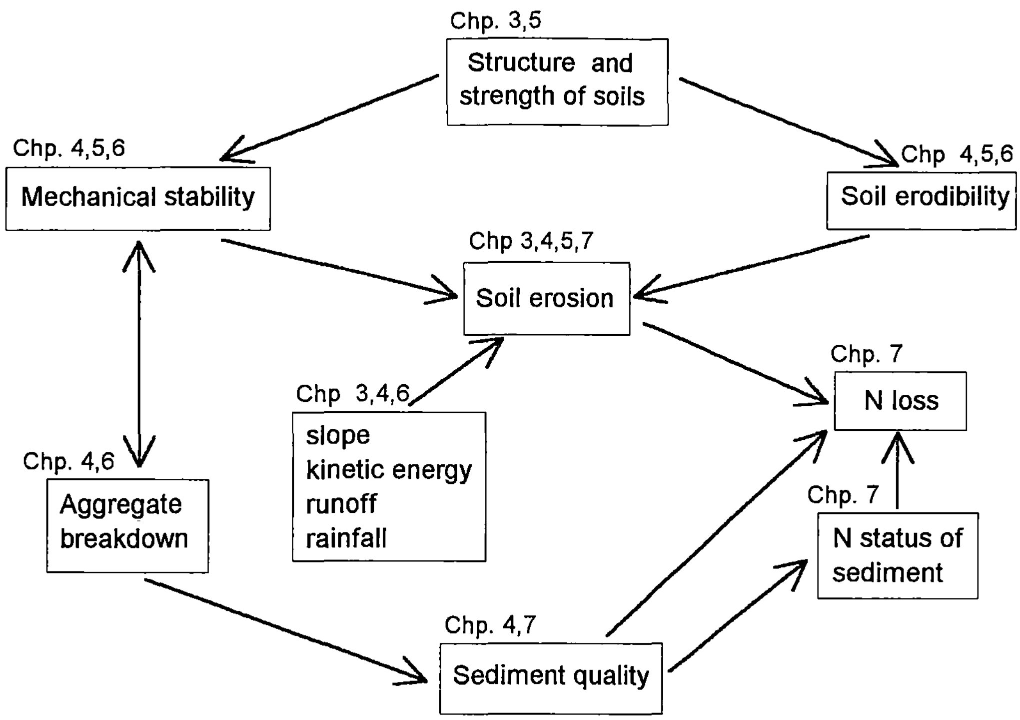


**Fig. 1.1** - Schematic diagram of the relationships between key elements of soil erosion and N loss which were the subjects of this study. The numbers of the Chapters where the key elements were analysed are also shown in the diagram.

## *1.3 OUTLINE OF CHAPTERS WITH MAJOR OBJECTIVES*

The thesis is divided into 8 chapters. Each of the Chapters 3 to 7 focus on specific aspects of research and, therefore, contain a review of literature that is relevant to the chapter. **Chapter 1** provides a general introduction, indicating the scope and background of the project and an overview of the hypotheses and objectives.

**Chapter 2** provides a general description of the materials and methods used in the research.

**Chapter 3** details the *in-situ* erosion experiments in the field which had the objectives:

- to determine the feasibility of conducting erosion experiments in the field.
- to study the temporal variation in erosion and in the characteristics of sediment.

**Chapter 4** describes erosion experiments in the laboratory under simulated rainfall, for three forest soils with the objectives:

- to examine variation in erosion and sediment characteristics of cultivated forest soils with varying slope and rainfall kinetic energy.
- to develop indices of the mechanical stability of soils to wetting and dispersion.
- to assess the importance of the mechanical stability of soils on erosion and size-distribution of sediments.

**Chapter 5** includes a detailed interpretation of the data on sediment concentration obtained from erosion experiments and an analysis of erodibility parameters with the objectives:

- to evaluate the influence of soil structure and strength on sediment concentration.
- to study the variation in soil erodibility parameters with structure and strength of soil.

**Chapter 6** describes the dynamics of aggregate breakdown under disruptive forces of ultrasound with the objectives.

- to test the adequacy of the ultrasonification method of soil dispersion in determining the energy required to break down soil aggregates of variable structure.

- to study the dynamics of aggregate breakdown in soil-water suspensions exposed to disruptive forces arising from ultrasonification.

**Chapter 7** provides estimates of nitrogen loss due to erosion from three forest soils with the objectives:

- to examine variation in the loss of N due to erosion with variation in soil structure, slope, and kinetic energy of rain.
- to investigate factors that determine the enrichment of N in sediments, when compared with the uneroded soil.
- to test the capability of models to predict the loss of N due to erosion from forest soils.

**Chapter 8** is a synthesis of the main findings of the thesis and it provides conclusions and future research needs arising from this research.

# Chapter 2

## MATERIALS AND METHODS

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### *2.1 CHARACTERISTICS OF SOILS USED*

Three forest soils were used in most of the experiments. The soils were chosen to represent sites which have been useful for commercial plantations of eucalypts with a broad range in structure and texture, and from different geographical locations in Tasmania. These sites had recently been cleared in order to establish new plantations. Soils were located at Dover, SE Tasmania (43° 22' S, 146° 57' W), Ridgley, NW Tasmania (41° 10' S, 145° 50' W), and Maydena, central part of Southern Tasmania (42° 42' S, 146° 41' W).

These soils will be referred to by their locations, *i.e.* Dover (D), Ridgley (R), and Maydena (M). A brief account of some of the pertinent characteristics of these soils is given in Table 2.1. Soil D was a poorly aggregated loamy sand, soil R a strongly aggregated clay, and soil M a clay loam with low aggregate stability when wet.

## *2.2 RAINFALL SIMULATIONS*

### 2.2.1 Features of the rainfall simulator

A portable rainfall simulator was used in all erosion experiments. The simulator was mounted on a two-wheeled trailer to be towed by a vehicle. Some components of the simulator could be retracted, which reduced the size of the unit allowing safe transport. The doors and roof of the simulator unit were unfolded to expose the nozzle during rainfall simulations (Fig. 2.1).

Table 2.1

Selected taxonomic and physical properties of the soils used in erosion experiments.

| Soil properties | Soil D (Dover) | Soil R (Ridgley) | Soil M (Maydena) |
|---|---|---|---|
| Parent material | Permian sandstones and mudstones | Basalt | Alluvial |
| Suborder [a] | Redoxic Hydrosol | Red Ferrosol | Grey Kurosol |
| Principal Profile Form [b] | Dy 5.21 | Gn 4.31 | Dy 5.41 |
| Sand % (2.0-0.02 mm) | 80.2 | 22.4 | 40.5 |
| Silt % (0.02-0.002 mm) | 11.1 | 23.8 | 24.2 |
| Clay % (< 0.002 mm) | 8.7 | 53.8 | 35.3 |
| Texture | loamy sand | clay | clay loam |
| Organic carbon (%) | 3.63 | 9.25 | 2.16 |
| pH [c] | 4.59 | 4.77 | 4.44 |
| EC (dS $m^{-1}$) [c] | 0.107 | 0.184 | 0.092 |
| Dispersion class[d] | 3 (4) | 3 (2) | 2 (1) |

[a] According to Isbell (1996)
[b] According to Northcote (1979)
[c] 1:5 soil water ratio
[d] According to Emerson (1967) with sub-classes in brackets according to Craze and Hamilton (1992).

**Fig. 2.1** - Rainfall simulator used in erosion studies. The doors (white) were flipped back to provide support for the roof where the nozzle was mounted. The water reservoir and water pump are on the trailer (red).

A detailed description of this rainfall simulator was given by Grierson and Oades (1977). The simulator consisted of a stationary pressurised spraying nozzle (Spraying Systems Fulljet 1 ½ H30 nozzle) positioned 2 m above an erosion tray. The nozzle was connected to a water reservoir of 400 L volume. Water was supplied to the nozzle by a pump of 2.5 hp (1.9 kW) capacity. A valve and pressure gauge enabled adjustment of the water pressure reaching the nozzle. There was a rotating disc with a radial slot directly below the nozzle which intercepted the spray of water. Excess water not used in spraying was redirected back to the reservoir through drainage pipes.

In field experiments, the electric motor used for rotating the disk (1/8 hp = 93 W) was powered with a generator, while in laboratory experiments the power was supplied directly from an AC power outlet. The angle of the radial slot on the rotating disc, together with water pressure, allowed regulation of rainfall rate. This type of simulator with a rotating disc is known to produce rainfall with intensity and kinetic energy similar to that of natural rainfall (Hudson, 1971).

In the erosion experiments, rainfall rate varied from 110 to 120 mm $h^{-1}$, which was within the range used in studies of erosion, and related aggregate breakdown and nutrient loss using simulated rainfall, e.g. Sharpley (1985): 60 and 120 mm $h^{-1}$; Loch (1989): 100 mm $h^{-1}$; Palis *et al.* (1990a,b): 100 mm $h^{-1}$ ; Ghadiri and Rose (1991a,b). 100 mm $h^{-1}$; Schultz and Malinda (1994): 100 mm $h^{-1}$; Torbert *et al.* (1996)· 125 mm $h^{-1}$.

### 2.2.2 Characteristics of the simulated rainfall

Average rainfall intensity was measured from the volume of water collected on a tray of 1 $m^2$ after a simulation of 15 minutes. Spatial variation of rain was obtained by measuring volume of water collected in 144 steel cans (of 7.5 cm diameter and 11.7 cm height) placed on a 12 × 12 grid, within the 1 $m^2$ tray. Uniformity coefficient of rain (UC, %) was calculated as:

$$\mathrm{UC} = 100\left(1 - \frac{\sum |v_L|}{\overline{v_L} \cdot n}\right), \tag{2.1}$$

where $|v_L|$ is the absolute value of the deviation of individual observations of volume of water (L) from the mean volume of water measured ($\overline{v_L}$), and $n$ is the number of observations ($n$ = 144).

Several 15-minute simulations were made at varying water pressure and slot angle of the rotating disc, to achieve the optimum combination for a rainfall rate ≈ 120 mm $h^{-1}$ with the minimum spatial variation (highest possible UC). Tests were made for water pressures of 35, 50 and 70 kPa, combined with slot angles of 5, 10, 20, and 30°. Results of the tests for rainfall rate and uniformity are presented in Tables 2.2 and 2.3 These tables showed that a desired rainfall rate of 100 - 120 mm $h^{-1}$ with a satisfactory UC could be obtained with a water pressure of 50 kPa and a slot angle of 30°. This setting was used in all erosion experiments.

Table 2.2

Rainfall rate (mm $h^{-1}$) for various combinations of water pressure and slot angle of the rotating disc

| | Water pressure | | |
|---|---|---|---|
| Slot angle (°) | 35 kPa | 50 kPa | 70 kPa |
| 5 | 8 | 14 | 16 |
| 10 | 19 | 25 | 35 |
| 20 | 60 | 70 | 78 |
| 30 | 102 | **115** | n.m |

n m.: not measured

Table 2.3

Rainfall uniformity (UC, %) for various combinations of water pressure and slot angle of the rotating disc.

| | Water pressure | | |
|---|---|---|---|
| Slot angle (°) | 35 kPa | 50 kPa | 70 kPa |
| 5 | 40 | 61 | 79 |
| 10 | 72 | 81 | 83 |
| 20 | 28 | 37 | 48 |
| 30 | 67 | **74** | n.m. |

The measured rainfall uniformity was lower than that reported by Grierson and Oades (1977) for the same rainfall simulator and nozzle Further cleaning of the nozzle and a check of mechanical components of the simulator did not improve UC.

### 2.2.3 Raindrop size and kinetic energy

Raindrops of two sizes were used in laboratory erosion experiments to vary kinetic energy (KE) while keeping the rainfall rate constant. KE could be varied by changing the size of raindrops before they reached the soil surface. The median size of raindrops for this type of simulator has been reported to be 2.6 mm (Grierson and Oades, 1977) This raindrop size represented the high KE in erosion experiments. To obtain low KE, the size of raindrops was reduced by allowing interception of large raindrops by a net (with apertures 1.0 × 1.5 mm) placed 0.5 m above the soil surface. The median drop size for low

KE was calculated as 0.65 mm (Fig. 2.2), obtained using a 2·1 mixture of engine oil (STP brand) and heavy mineral oil with the oil method of Eigel and Moore (1983).

Assuming raindrops to be spherical, the kinetic energy (KE) associated with raindrops for a given rainfall event is

$$\mathrm{KE} = \frac{1}{2}\mathrm{Mv}^2, \qquad (2.2)$$

where M and v are respectively the mass (kg) and velocity (m $s^{-1}$) of raindrops. Assuming the density of water to be a constant (1000 kg $m^{-3}$), the mass of a raindrop (M) could be converted to size (d) via the volume of raindrop. Therefore, equation (2.2) can be stated as

$$\mathrm{KE} = k_d \mathrm{d}^3 \mathrm{v}^2, \qquad (2.3)$$

where $k_d$ is a constant arising from the consideration of density in conversion of mass to volume of raindrops.

Two different KE were used in the erosion experiments: low (with a drop size of 0.65 mm) and high (with raindrop size of 2.60 mm). It should be noted that KE of rain can be estimated from rainfall intensity (Wischmeier and Smith, 1978; Park *et al.*, 1983), but such methods of estimation could not be used when KE was varied at a constant rainfall intensity

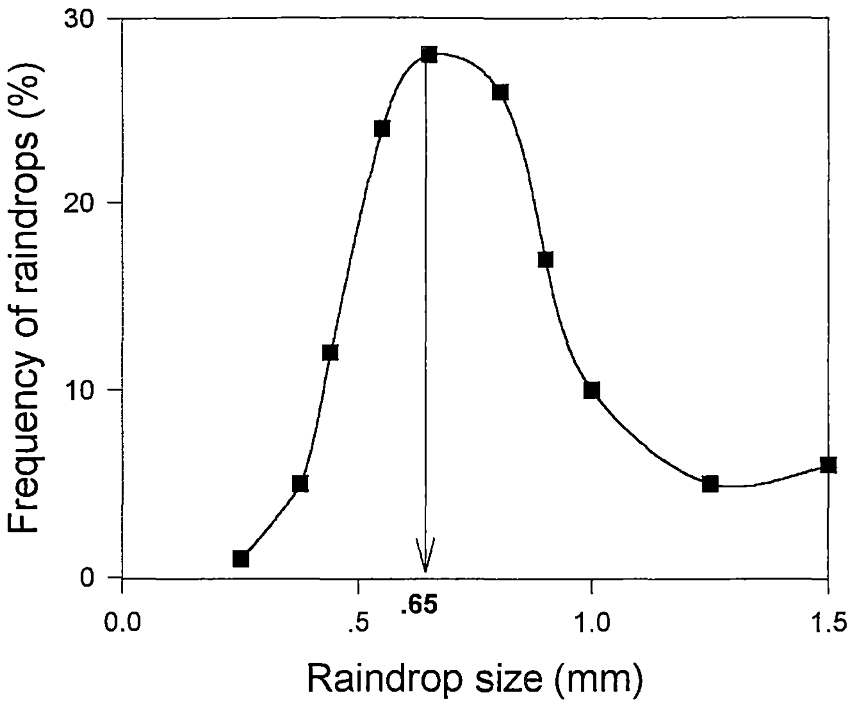


**Fig. 2.2** - Distribution of raindrop size for rainfall with low KE. The median raindrop size for a sample of 134 raindrops is indicated with an arrow.

### *2.3 FIELD EROSION STUDIES*

#### 2.3.1 Erosion site

Field erosion studies were carried out on a site near Dover. Soil from this site (soil D) was also used in other experiments (Section 2.4). The site was located on Old Hastings Road, South of Strathblane, in SE Tasmania. Annual rainfall of this area is 1300 mm.

The site was cleared from a regrowth forest dominated by *Eucalyptus obliqua*. Experiments with simulated rainfall commenced one month after the forest had been harvested and the site cleared to establish eucalypt plantation. The site was ripped and mounded (mound orientation was up and down the slope) prior to planting. Typical slope of a cultivated mound was 15-16°. Although there was little vegetative cover at the time experiments were conducted, some organic debris covered the surface.

Experiments with simulated rainfall were conducted during January to June 1995, depending on weather conditions. All rainfall simulations were carried out on mounds only, but successive simulations were never repeated at the same location. For each simulation, care was taken to select mounds without any vegetation, but with a similar slope.

### 2.3.2 Erosion plots and runoff collector

A mound with a uniform slope of 16 °, free from any vegetation, was located and marked with pegs. Pieces of woody residue ≥ 5 mm in size were removed from the marked area. Apart from that, there was no other soil disturbance within the marked area. The rainfall simulator was then placed at the desired position, and held at that position using the adjustable legs of the simulator unit to improve stability The side doors of the simulator were opened and the roof was slid out to expose the nozzle. During simulations, a retractable tarpaulin cover was used to reduce any interference from wind which could influence spatial variation of rain (Fig. 2.3). Rainfall simulations were never carried out in windy conditions

Before each rainfall simulation, the simulator was run for 10 minutes to collect rain in a 1.5 $m^2$ tray to measure rainfall intensity. A plumb line was used to locate the projected nozzle position on ground. A runoff plot was established by enclosing a 1 $m^2$ area such that the nozzle of the simulator was at the centre of the plot. Three steel plates (1 m long, 25 cm wide) were inserted 10 cm into the ground to make up three sides of the erosion plot, and a runoff collector was placed at the downslope end The runoff collector was 1 m long to cover the full width of the plot, and consisted of a 10 cm deep trough with a gentle slope to collect all runoff and sediment, and divert these to sample plastic bottles (5 L capacity) through a flexible hose. A tilted roof over the collector prevented direct entry of rain. Each sample bottle was placed within a small hole just outside the runoff plot (down the slope) for quick transfer of runoff from collector to sample bottle.

Four simulated erosion experiments were conducted during 23 January to 1 May 1995, each consisting of a single rainfall event. The first event lasted 27 minutes from the time runoff started. The duration of remaining events was 40 minutes. All runoff and sediment were sampled at 2.5 minute intervals for the entire duration of each simulation. Samples of runoff and sediment were transported in plastic bottles to the laboratory from field site in an upright position. It should be noted that transport of sediment samples in this manner has little effect on the size-distribution of sediment, as Clearly *et al.* (1987) found no significant differences in aggregate-size distribution of sediment if it was analysed immediately after rainfall simulation or after transport of samples in an upright position for 60 km

**Fig. 2.3** - Rainfall simulator used in field studies. A tarpaulin cover was attached to the side doors to reduce the effect of wind on spatial distribution of rain.

**Fig. 2.4** - Measurement of penetrometer resistance in a typical erosion tray. The grid (10 × 10 cm) was used as a guide for random selection of each point. (A figure for measurement of surface shear strength is given in Section 2.4.4).

### 2.3.3 Measurements

For each simulated rainfall event, measurements were made of soil strength before and after erosion, plus size-distribution of wet uneroded soil and selected sediment samples. Additionally, rainfall rate was measured for each event prior to erosion as described in the previous section.

The volume of runoff in each sample was measured on decanted water after the sediment had undergone deposition for at least 1 h. Runoff rate was calculated by dividing the volume of runoff over the duration of sampling (2.5 min). Sediment concentration (kg $m^{-3}$) for each sample was estimated as the ratio of oven-dry weight of sediment (110 °C) and the volume of sediment plus runoff, measured before decanting.

Soil strength was measured before and after each rainfall simulation with a pocket penetrometer (Geotester, Italy) and a Torvane shear device (Soiltest, ELE International, USA). The shaft of the pocket penetrometer was modified to attach a steel probe with a conical tip (60 ° cone angle, 4 mm basal diameter) Penetrometer resistance was measured with a resolution of 0.05 kg load (39 kPa) by pushing the penetrometer cone into the soil to a depth of 4 cm. The readings of penetrometer resistance were converted to kPa by multiplying the value in kg force by the acceleration due to gravity and dividing by the projected area of the conical probe. Surface shear strength was measured with the Torvane by rotating a vane of 49 mm diameter at a soil depth of 5 mm, i.e the full depth of the vanes The readings of this vane had a ratio of 0.2 to convert to measurements with a standard vane size. Resolution of shear strength measurements was 0.025 kg cm$^{-2}$ (0.049 kPa). Each measurement was replicated at least ten times covering the entire erosion tray. The locations of ten measurements were selected by assigning numbers to each square of a 10 $\times$ 10 cm grid and generating random numbers for 10 squares (Fig. 2.4).

For three simulated events, the size-distribution of wet sediment was determined by sieving sediment samples collected during 0-5, 12.5-17.5, and 35-40 minutes of erosion. For the fourth event, wet-sieving of sediment was made after 0-6.5, 10-16.5, and 20-27 minutes of erosion. Samples of uneroded soil taken from the vicinity of the plots were submerged in water for 40 minutes, and then sieved as for the sediment samples. Soil or sediment was separated into seven size-classes: $\geq$ 2000, 2000-1000, 1000-500, 500-250, 250-150, 150-53, and < 53 µm. The wet sediment was spread thinly on the top of a nest of 6 sieves and water was gently sprayed with a spray bottle to facilitate separation. Tap water was used because it is known to have no effect on size-distribution of aggregates obtained with wet sieving (Loch, 1989). Use of physical and chemical agents was avoided during wet sieving to ensure that further breakdown of aggregates did not occur during separation of the sediment. The amount of soil material which passed through the last sieve (< 53 µm) was estimated by (1) measuring the total volume of < 53 µm sediment suspended in water; (2) taking two sub-samples of 500 cm$^3$ from that suspension after homogenization; (3) weighing the sediment in each sub-sample after drying; and (4) converting the weight per unit volume of sub-samples to the total volume measured in (1). All samples of soil and sediment were dried at 40 °C. The weights were then converted to that at 110 °C using a conversion factor (CF) determined separately for samples of submerged uneroded soil dried at 40 °C and 110 °C.

## *2.4 LABORATORY EROSION STUDIES*

### 2.4.1 Soil Sampling, handling and storage

Three soils were collected from the top 0.2 m of three recently cultivated plantation sites as described in Section 2.1, and were spread out under shade to dry. Stone and organic debris were removed after the soil was air-dry, and clods were crushed by hand to reduce the size to ≤ 8 mm. The air-dry soil was then stored in metal drums (with lids) until used in experiments.

### 2.4.2 Erosion trays and runoff collector

Each erosion tray used in the experiments was 1.0 m long (downslope), 1.0 m wide, and 0.1 m deep. There was a buffer area of 0.1 m×1.0 m along both sides of tray (which was part of the 1.0×1.0 m tray) to reduce sediment loss by splash. No drainage was provided for soil in the trays. Each tray was mounted on a frame which could be adjusted to provide two slope angles of 2 and 16 °. At the lower end of each tray, a runoff collector was attached to allow sampling of runoff and sediment from an area of 1 0 m×0 8 m (excluding buffer). This runoff collector could be moved up or down to be leveled with the soil surface in the tray. A hinged cover over the runoff collector prevented direct entry of rain into the collector. This cover could be flipped back to gain access into the runoff collector, allowing complete recovery of sediment (Fig.

2 5). Within the collector, runoff and sediment were routed to an outflow tube into sample bottles. Sample bottles were changed regularly to obtain runoff and sediment at desired intervals within a simulated rainfall event.

A description of the experimental design and treatments is given in Section 4.2 1.

### 2.4.3 Simulation procedure

Each erosion tray including the buffer area was filled with loose air-dried soil to resemble recently tilled soil. The surface of the packed soil was levelled with a wooden plate. Prior to each simulation event, tap water was gently sprayed on the packed soil bed until ponding was visible. During pre-wetting, each erosion tray was kept in a horizontal position. The tray was then covered with a black polyethylene sheet and left overnight to allow for redistribution of water throughout the soil.

The duration of each rainfall simulation was 40 min from the appearance of runoff. All runoff and sediment were collected at intervals of 2.5 min.

### 2.4.4 Measurements

The methods used for measurements of rainfall and runoff rates, sediment concentration, soil physical condition before and after erosion, and the characteristics of sediment and soil were similar to those described for field simulations (Section 2.3.3) with some exceptions as described below. Rainfall rate was measured before each simulation for a duration of 15 minutes instead of 10 minutes. Samples of uneroded soil were collected from the buffer area of the tray before each simulation. Sieves used for wet-sieving of soils and sediments had apertures of 2000, 1000, 500, 250, and 53 μm The size-distribution of dry uneroded soil was obtained with the same set of sieves as for wet-sieving, but with the help of a sieve shaker run for 3 minutes. The size-distribution of the wet uneroded soil with dispersion was obtained by wet-sieving a suspension of 30 g air-dry soil and 100 $cm^3$ of sodium-hexametaphosphate solution (50 g $L^{-1}$) after the suspension was mechanically dispersed for 5 minutes in an electric blender.

Sub-samples of about 0.3 g from each size-class of soil and sediment were used for the analysis of nitrogen (N) Samples were digested for total-N using the sodium-thiosulfate-salicylic acid method, which is a modification of the Kjeldahl method (Rayment and Higginson, 1992). Digested samples were analysed in a flow injection analyser (QuikChem800, Lachat Instruments, USA) by the colorimetric method

**Fig. 2.5** - Erosion tray used for laboratory erosion experiments. Erosion from the tray did not include sediment from the buffer areas present on both sides of the tray. The runoff collector attached to the down slope end of tray with the hinged cover (flipped back) is also shown.

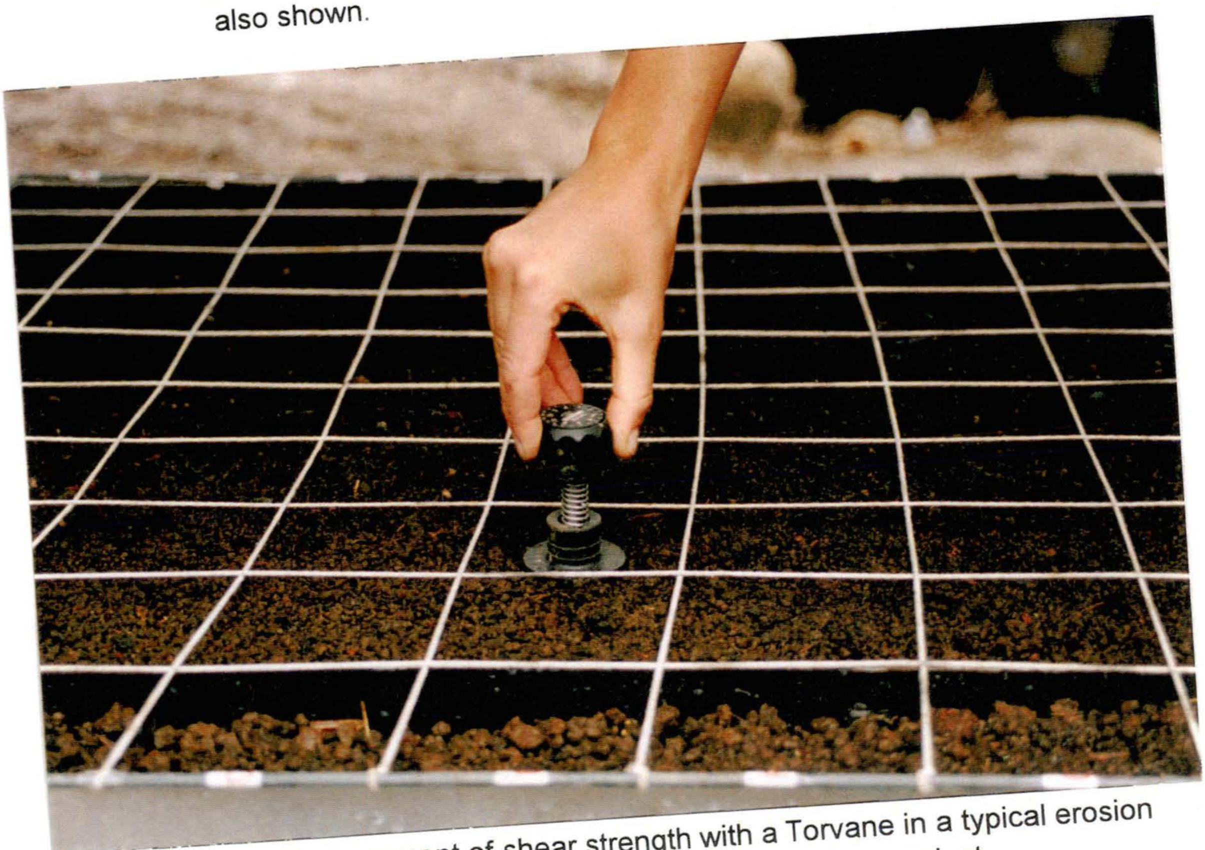

**Fig. 2.6** - Measurement of shear strength with a Torvane in a typical erosion tray. A 10 × 10 cm grid was used as a guide to select measurement points.

For each simulated event, sediment samples of all size-classes for the sampling times of 0-5, and 35-40 min were analysed for total-N These sediment samples were obtained by combining two consecutive samples, each of 2.5 min duration. This was necessary in order to obtain enough sediment of each size-class for N-analysis. For each event, total-N was analysed for the whole uneroded soil and also for various size-classes of uneroded soil. Each measurement of N was replicated three times (corresponding to any replicate of any erosion treatment) for the uneroded soil with fractionation. For the whole uneroded soil without fractionation, N-analysis was made for every sample corresponding to each soil, erosion treatment, and replicate.

## *2.5 ULTRASONIC DISPERSION OF SOIL AGGREGATES*

### 2.5.1 Equipment used

An ultrasonic probe of 22-mm diameter and 100-mm length (Measuring & Scientific Equipment Limited) was used for all ultrasonic dispersion studies. During sonification, the probe and the soil-water suspension were enclosed in a sound proof cabinet. Ultrasonic energy was applied after inserting the probe tip into a Pyrex beaker containing the soil-water suspension. The Pyrex beaker was always placed in a polystyrene cup. During sonification the temperature of suspension was measured with a thermistor, coupled to a data logger (DS93, Dataflow Systems, Queensland) Temperature (± 0.05 °C) was recorded at 1 s intervals for the entire duration of each experiment, except during cooling. The thermistor was pre-calibrated against a platinum resistance thermometer

(Temperature calibrator, model D55SE, METEK, Jofra Instruments, operating range -45 to 123 °C, maximum uncertainty in measurements ± 0 02 °C).

### 2.5.2 Measurement of energy components of the system

The output-power readings of the ultrasonic probe were not accurate enough to measure directly the energy used in dispersion of soil. Therefore, various components of the output energy were estimated indirectly. The total energy applied during sonification of a soil-water suspension ($E_i$, J) was equal to the sum of the various components of the system:

$$E_i = E_h + E_c + E_t + L_g , \tag{2.3}$$

where $E_h$ was the energy used in heating the suspension, $E_c$ the energy lost by conduction, $E_t$ the energy lost by transmission of wave energy, and $L_g$ the energy used in dispersion of soil aggregates. For a suspension containing soil which has been dispersed completely, $L_g = 0$. Therefore,

$$E_i = E'_h + E'_c + E'_t \ , \tag{2.4}$$

where $E'_h$, $E'_c$, and $E'_t$ were the energies used in heating the suspension, lost by conduction and lost by transmission, respectively. Full dispersion of soil was assumed to have occurred when the soil-water suspension was sonified for 15 minutes.

Each component of equations (2.3) and (2.4) was estimated, except that components $E_t$ and $E'_t$ were considered to be ≈ 0 for the experimental conditions of this study (Ian Newman, pers. com.; Raine and So, 1993, 1994).

$E_h$, $E_h'$, $E_c$, and $E_c'$ (with fully dispersed or undispersed soil) were estimated respectively from the change in temperature of suspension during sonification ($E_h$, and $E_h'$) and during the cooling down of suspension ($E_c$, and $E_c'$). The specific energy of the soil-water suspension during sonification was also considered for the estimates of $E_h$, $E_h'$, $E_c$ and $E_c'$

$$E_k = (M_w c_w + M_s c_s + M_v c_v)\Delta T\,, \qquad (2.5)$$

where $E_k$ was either $E_h$, $E_h'$, $E_c$, or $E_c'$; $M_w$ and $c_w$ were respectively the mass and specific heat of water; $M_s$ and $c_s$ were respectively the mass and specific heat of soil; $M_v$ and $c_v$ were respectively the mass and specific heat of the beaker which contained the soil-water suspension during sonifications. In Equation 2.5, $\Delta T$ was the change in temperature during sonification (for $E_h$ and $E_h'$), or during the cooling down of suspension (for $E_c$ and $E_c'$). The energy used in aggregate breakdown (component $L_g$ in equation 2.4) was estimated with the assumption that the cumulative energy applied by the ultrasonic probe did not change for a given power setting and it remained independent of the dispersion state of the suspension. Consequently, $E_t$ of a fully dispersed suspension was equal to $E_t$ of an undispersed suspension. Therefore, from equations (2.3) and (2.4), $E_h + E_c + L_g = E_h' + E_c'$, or

$L_g = E_h' + E_c' - E_h - E_c$.

Sonification of the suspension stopped when the temperature reached ≈ 45 °C. During sonification, the change in temperature of the suspension was recorded every second. For the measurement of losses of energy due to conduction, soil-water suspensions with fully dispersed soil and undispersed soil were sonified for 10 minutes, regardless of the maximum temperature reached by the suspension. After sonification was stopped, the suspension was allowed to cool down to room temperature. During cooling, the temperature of the suspension was recorded every 5 seconds

The heat components of equation (2.5) were obtained as follows: both $M_w$ and $M_s$ were measured before each sonification. The specific heat of the water ($c_w$) was obtained from Weast (1981): $c_w = 4.2$ J $g^{-1}$ °$C^{-1}$ at 20 °C. The parameter $c_s$ was calculated by adding the products of specific heat of each soil constituent with the proportion of that constituent in a soil sample. The specific heat of mineral constituents of soil is 0.73 J $g^{-1}$ °$C^{-1}$, and that of organic matter is 1.93 J $g^{-1}$ °$C^{-1}$ (Van Wijk, 1963). Organic matter for these soils was converted from organic carbon (OC) (Table 2.1) as 1.72×OC. Three soils were used in these studies. soils D, R, and M (details in Section 2.1). Table 2.4 gives the composition of each soil, in terms of mineral and organic constituents and water content (for air-dry soil), and the specific heat of soils. The heat capacity of the ultrasonic probe was not determined as its contribution to changes in temperature of the system was assumed negligible.

Information on $c_v$ for the beakers (Pyrex borosilicate glass) was obtained from the distributor of the containers (BIBY Sterilin Ltd , 1995). The specific heat of the two containers used in the experiments was 0.750 J $g^{-1}$°$C^{-1}$ The mass of the beakers were 48.268 g for container no. 1, and 49 905 g for container no 2.

Table 2.4

The composition and specific heat of air-dry soils used in the experiments.

| Soil | Mineral constituent (%) | Organic constituent (%) | Water content (% by weight) | Specific heat of soil, $c_s$ (J $g^{-1}$ °$C^{-1}$) |
|---|---|---|---|---|
| D | 92.78 | 6.26 | 0.96 | 0.838 |
| R | 80.67 | 15.95 | 3.38 | 1.038 |
| M | 95.44 | 3.72 | 0 84 | 0.804 |

### 2.5.3 Experimental conditions

Several studies have shown that some properties of the soil-water suspension could affect the results obtained for soil dispersion by ultrasound. These properties include the concentration of suspension, the volume of suspension, the temperature, and the gas saturation of suspension. Before starting experiments with the ultrasonic probe, tests were performed to determine the optimum experimental conditions relating to the characteristics of equipment and soils. These tests were similar to those of Raine and So (1994).

The depth of insertion of the ultrasonic probe into the suspension affects inversely the power dissipated into the system To avoid any change in dissipation of power in various experiments, a fixed depth of 3 mm was used in all experiments, as a thorough mixing of the suspension could be observed with this depth of insertion

When a suspension is exposed to ultrasound, the volume of suspension affects the efficiency of dispersion via mixing. A volume of 50 $cm^3$ was found to provide a complete mixing of the suspension.

The concentration of suspension may influence the power available for soil dispersion via particle abrasion. A thorough mixing of suspension during sonification has been reported to occur at a soil:water ratio of 1:5 to 1:10 (Christensen, 1992; Raine and So, 1994). For the equipment and soils used in the present experiments, a good mixing of suspension could be obtained at a soil:water ratio of 1:6.

An increase in the amount of air entrapped in a suspension can increase the dissipation of energy as it enhances the effect of sonification in dispersing soil. Similarly, degassing the suspension can reduce the rate of energy dissipated into the system. As it was difficult to control precisely the amount of air entrapped in the suspension, water used in all suspensions was degassed. For all experiments, deionised water (in 1 L bottles) was degassed in an ultrasonic bath for 20 minutes. The lids of the bottles were kept open to release entrapped air. However, during the addition of degassed water to make up the soil-water suspensions there was a possibility of some air entrapment, which may have reduced the effects of previous degassing.

*2.5.3.1 Sonification of soil-water suspension*

A100 $cm^3$ Pyrex beaker containing $\approx$ 8.3 g of air-dry soil ($\pm$ 0.001 g) and 50.0 g of degassed, deionized water was placed in a polystyrene cup for insulation and to reduce loss of energy via conduction and transmission ($E_c$ and $E_t$). The beaker, including the polystyrene cup, was placed inside the soundproof cabinet. The ultrasonic probe was then inserted to a depth of 3 mm into the suspension, and positioned in the middle of the container. A thermistor probe (6-mm diameter) was placed inside the suspension (without touching the probe). The thermistor was connected to a datalogger placed outside the cabinet. The datalogger was turned on 5 s before the ultrasonic probe was switched on. The temperature of the suspension was recorded continuously for the entire duration of sonification.

There were three replicates of each soil D, R, and M, which were chosen randomly for sonification. A total of 8 durations of ultrasonic energy were applied to the soil-water suspensions: 15, 60, 100, 180, 360, 540, 720, 900 seconds. The temperature of suspension was never allowed to exceed 45 °C (except during the measurement of $E_c$ and $E_c'$), in order to avoid any possible effect of high temperature on the stability of soil aggregates. Whenever the temperature approached 45 °C, sonification was interrupted for the suspension to cool down, and then sonification continued for the remaining time Sonification was interrupted as many times as necessary to avoid temperatures > 45 °C.

After sonification, the suspension was transferred to a plastic container and kept in a constant temperature room (20 °C) until particle-size analysis (PSA) was made. Time between sonification and PSA varied from 6 to 36 h.

*2.5.3.2 Particle-size analysis (PSA)*

Following sonification, the suspended soil material of < 53 μm was transferred to a 1 L measuring cylinder by passing the suspension through a sieve of 53 μm aperture via a funnel. Excess deionized water was used to wash the soil through the sieve, and to fill the cylinder up to the 1 L mark. The soil retained on the sieve (fraction ≥ 53 μm) was transferred to tared aluminium containers and dried at 105 °C.

The PSA of < 53 μm was measured with the pipette method of Gee and Bauder (1986) as follows: the cylinder containing the < 53 μm fraction in suspension was placed once again in the constant temperature room (at 20 °C) for the suspension to equilibrate to 20 °C. After equilibration, the cylinder was shaken end-over-end by hand for 60 seconds during which the top of the cylinder was temporarily sealed with a piece of polyethylene film (Cling Wrap). As soon as the cylinder was returned to an upright position, a stopwatch was started to record time. Two 25 $cm^3$ samples were collected with a pipette at 4 min 48 s, and at 8 h 38 min, to obtain silt (≤ 20 μm) and clay (≤ 2 μm) fractions. Each sample was drawn from a depth of 10 cm from the top of the suspension. The duration of sample collection did not exceed 12 seconds, as advised by Gee and Bauder (1986). All samples were dried at 105 °C and weighed. The proportion of silt and clay was calculated for each suspension. PSA was also made for soil samples simply immersed in water but without sonification, to determine the effect of immersion-wetting on the proportion of silt and clay.

A previous PSA was made for each wet-dispersed soil (Section 2.4.4) where complete dispersion was achieved with the use of chemical agents and mechanical action. The proportions of silt and clay obtained with this method were taken as the maximum dispersible material for each soil

Oven-dry soils $\geq$ 53 μm after sonification, immersion-wetting and after complete dispersion were further sieved through a set of sieves (of apertures 2000, 1000, 500, 250, and 53 μm) after 3 minutes of shaking with a mechanical sieve-shaker. Soil retained in each sieve was weighed.

# Chapter 3

## FIELD STUDIES OF EROSION

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### *3.1 INTRODUCTION*

Soil erosion in field conditions has been the subject of various studies. However, few have focused on forest soils. In a small watershed in Arkansas (USA), on shallow soils derived from sandstone, Miller et al. (1988) measured the effects of three silvicultural treatments (clearcutting, selection cutting, and no disturbance) on soil erosion and off-site sediment. Sediment yield from clearcut compared to no disturbance was 20-fold in the first year after site preparation, and 3-fold after 3 years, with a mean soil loss varying from 240 to 180 kg $ha^{-1}$ $yr^{-1}$ in the first and third years, respectively Thongmee and Vannaprasert (1990) compared erosion in several land use treatments on a forest soil in Thailand. They measured a soil loss of 74 Mg $ha^{-1}$ $yr^{-1}$ in the bare soil treatment, compared with 0.4 to 1 Mg $ha^{-1}$ $yr^{-1}$ in a young eucalypt plantation, after a total accumulated rainfall of 1150 mm, in plots of 4 m width, 20 m length (downslope) on a 8-10 % slope Blackburn *et al.* (1990) compared erosion in undisturbed and cultivated forest watersheds in East Texas and Northern Mississippi (USA). They found that undisturbed watersheds had less erosion due to less storm flow and lower peak-discharge rates than watersheds with clearcut harvesting and mechanical site preparation. They attributed this increase in erosion to the removal of vegetation that reduced evapotranspiration and increased soil moisture A review of Moffat (1991) on erosion studies in forested areas of UK, indicated erosion rates to range from 210 to 1300 kg $ha^{-1}$ $yr^{-1}$, with the highest rates occurring in areas subject to site preparation or harvesting.

In Australia, Wilson and Lynch (1992, 1993, and 1994) studied the impact of recent logging activities on water quality and soil erosion of two catchments with different soil characteristics in north-east Tasmania. They studied two forest soils that had different susceptibilities to erosion: a very erodible granite soil and a less erodible dolerite soil. Field plots of 300 $m^2$ had two erosion treatments: logged-burnt and unlogged-unburnt. The plots were subjected to simulated rainfall of intensities 40, 75, and 150 mm $h^{-1}$ during 10-40 min. Erosion from the plots varied from about 100-2000 kg $ha^{-1}$, and 2-100 kg $ha^{-1}$ in the logged-burnt and unlogged-unburnt treatments, respectively. This study also illustrated a marked difference in erosion between the more erodible granite soil and the less erodible dolerite soil. Also in Tasmania, Davies and Nelson (1993) studied the effects of steep-slope logging on fine sediments of ephemeral and perennial streams. They considered four size-fractions of sediments in the river beds: < 125, 125-250, 250-500, and 500-1000 μm. For all size fractions, sediment yield was significantly greater on logged stream sites than on unlogged sites. Median sediment yield for sites logged increased compared with unlogged sites by a factor of 3:1. Laffan *et al.* (1996) described a method to assess soil erodibility of Tasmanian forest soils. Soil resistance to aggregate breakdown was rated on the basis of aggregate stability (expressed as % water-stable aggregates > 250 μm), approximate soil strength (expressed as the difficulty in pressing the thumb or fist into moist soil), stoniness, and thickness of surface horizons. The resistance to aggregate breakdown was combined with classes of permeability and drainage to establish finally the erodibility rating of soils. This method provided a qualitative separation of soil erodibility into eight classes which ranged from low to very high. However, this method has not been fully tested yet and, therefore, it can not provide a quantitative estimate of potential soil erosion.

The research described in the preceding paragraphs showed that cultivated bare soil is significantly more prone to erosion than undisturbed bare soil. When a plantation is established, the site is essentially bare from the time it is cleared until substantial growth of trees or weeds occurs to provide some aerial cover. Site preparation before planting often includes cultivation and mounding. A typical mound has a slope of around 16 ° with an inter-mound distance of 2.5 - 4 0 m. When the soil is bare, mounds may contribute to localised erosion, and when such mounds are on a slope, and their orientation is up and down the slope, the scope for erosion is further enhanced.

Although field studies of soil erosion under natural rainfall are desirable, they require a great amount of time, labour, and capital, which are not always available. Furthermore, the variables of erosion (e.g. slope angle, slopelength, rainfall) are difficult to control under field conditions, preventing further use and transportability of the data from field experiments. In order to determine the contribution of various erosion variables and their interactions to sediment yield, some control of the variables which determine erosion is helpful. The use of simulated rainfall in field experiments provides control of variables such as rainfall intensity, duration, and raindrop size. Unless erosion plots are small, *in-situ* erosion experiments with simulated rainfall would still require considerable resources. In addition, care is needed in interpreting the estimates of soil loss from simulated rainfall studies due to possible differences in the characteristics of rainfall in simulated events compared to natural rainfall events.

In this Chapter, erosion from field experiments under simulated rainfall is reported as a first step in understanding the *in-situ* dynamics of erosion from cultivated forest soils. In addition, spatial and temporal variation in erosion is also examined.

## *3.2 METHODS*

### 3.2.1 Experimental conditions

All field experiments of erosion were carried out on one site only (Dover). Soil from this site (soil D) was also used in laboratory erosion experiments. The detailed features of the portable rainfall simulator and other aspects of the erosion experiments were described in Chapter 2. There were four simulated erosion experiments carried out on the mounds of recently cultivated soil, within the same plantation site. The locations of experiments were close to each other (≈ 10 m). Attempts were made for the erosion plots to resemble each other in terms of slope (14 - 16 °), aspect, soil cover (no vegetation, and removal of large woody debris prior to each experiment), and microtopography. These experiments are referred to by their numbers: #1, #2, #3, and #4. There were differences in rainfall rate from one experiment to another. The duration of sampling of runoff in the first experiment was different from the other three experiments. Measurements of soil strength were also made after erosion in the first experiment. Table 3.1 summarises various aspects of field experiments.

Table 3.1

Time and other details of erosion for field experiments.

| Experimental details | Experiment #1 | Experiment #2 | Experiment #3 | Experiment #4 |
|---|---|---|---|---|
| Date of experiment | 23.1.95 | 7.3.95 | 14.3.95 | 1.5.95 |
| Rainfall rate (mm $h^{-1}$) | 101 | 84 | 93 | 64 |
| Duration of runoff (min) | 27 | 40 | 40 | 40 |
| Slope (°) | 16 | 14 | 14 | 16 |
| MWD (µm) | 2048 | n.a. | 1260 | 1690 |

n.a. - not available.

As the drainage of the site was poor, trafficability was not good when the site was wet; therefore, experiments could not be carried out in quick succession. Due to the interval between experiments and variation in weather conditions, antecedent soil moisture may have varied from experiment to experiment, but soil moisture content was not measured.

For each erosion event, runoff volume and sediment were collected every 2.5 minutes, and calculations were made to obtain runoff rate (mm $h^{-1}$) and sediment concentration (kg $m^{-3}$). Soil strength (penetrometer resistance and shear strength) was measured before and after erosion. Measurements of the size-distribution were made for soil samples collected prior to erosion, and for sediment samples from selected sampling times.

### 3.2.2 Data analysis

Mean Weight Diameter (MWD) of aggregates was calculated from the data on size-distribution of uneroded soil and sediment for each experiment as

$$\mathrm{MWD} = \sum_{i=1}^{n} \overline{x_i} \times w_i\,, \qquad (3.1)$$

where $\overline{x_i}$ was the mean diameter (μm) of size fraction $i$, $w_i$ the proportion of the total sample (by weight) of that size fraction $i$, and n the total number of size fractions.

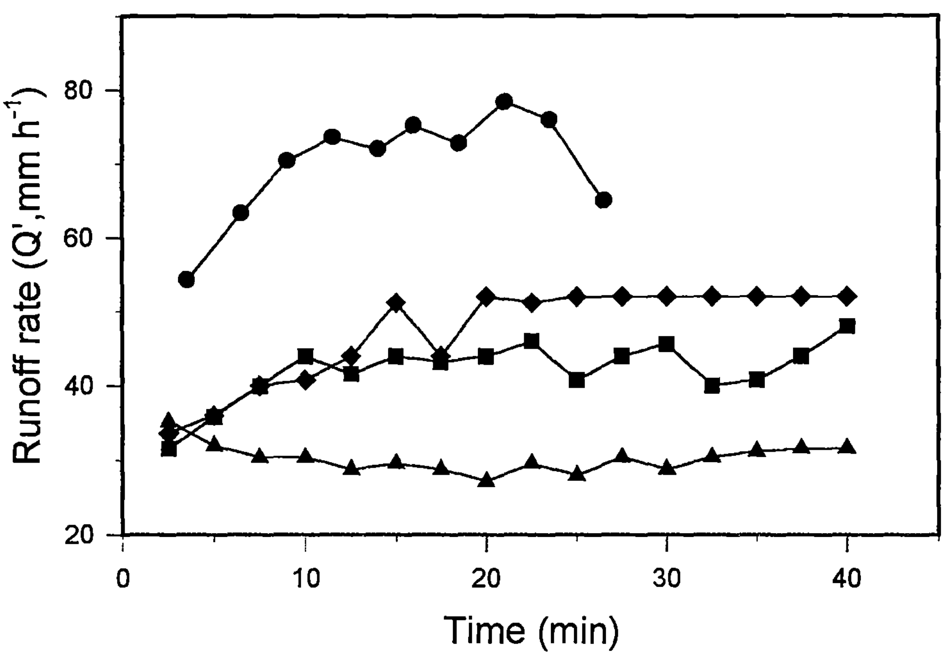


**Fig. 3.1** - Variation of runoff rate (Q') with time from runoff for field erosion experiments. #1 -● , #2 -■ , #3 -▲ , #4 -◆ .

## *3.3 RESULTS AND DISCUSSION*

### 3.3.1 Runoff rate (Q') and soil structure

Runoff rate (Q', mm h$^{-1}$) varied with time within an experiment and also between experiments (Fig. 3.1). Q' was the highest in event #1. In this event, there was a sharp increase in Q' within the first 10 minutes of erosion, Q' remained relatively stable for about 10 minutes before dropping sharply at the end of experiment. In other experiments (#2 to #4) runoff rate was relatively steady after the first 20 minutes of erosion.

Differences in Q' between experiments and temporal variation within an experiment may have been due to the differences in rainfall rate and antecedent soil moisture (ASM). A higher rainfall rate may have contributed towards a higher runoff rate in event #1 compared with other events (Table 3.1, Fig. 3.1). Per cent of rainfall that appeared as runoff varied between experiments (32 % in #3 to 74 % in #4), which suggests that other factors, apart from rainfall rate, contributed to variation in Q'. However, there may have been other losses of rain due to seepage and interflow which contributed to the differences in per cent of rainfall that appeared as runoff. There were also occasional losses of runoff and sediment due to leakage.

ASM is known to affect Q' directly by influencing the capacity of soils to absorb water. Soils at field capacity have lower infiltration rate than drier soils, thus producing higher Q'. ASM also affects Q' indirectly because it influences the rate of soil wetting and aggregate stability. Fast wetting of a dry soil may enhance aggregate breakdown due to slaking and dispersion of clay. These processes reduce the size of soil material such that soil physical condition is modified with a reduction in porosity and a decrease of infiltration rate, leading to an increase in runoff rate (Beare and Bruce, 1993; Le Bissonnais *et al.*, 1993; Truman and Bradford, 1993; Bresson and Moran, 1995).

For a given soil Q' can change from one erosion event to another due to a change in soil structure, even without a change in rainfall rate. This is because prevailing weather conditions between erosion events, particularly the rainfall regime, is known to influence the structure of bare soil (Dexter *et al.*, 1983). Soil structure depends on the size-distribution of primary particles and the forces affecting their arrangement (Marshall *et al.*, 1996). Physical and biological factors contribute to the rearrangement of particles. Cycles of wetting and drying also influence the nature and strength of these factors altering the structure of soil (Makeyeva, 1989). The structure of a soil can be measured by the size-distribution of soil aggregates, as obtained by sieving. An average size for the structural unit of soil is given by the Mean Weight Diameter (MWD). The values of MWD determined for the uneroded soil prior to rainfall simulations were: 2048 ± 327 μm for #1, 1260 ± 142 μm for #3, and 1690 μm for #4 (no SE was available for #4, and no measurements of MWD were available for #2). These values of MWD show that soil structure prior to erosion was different for various simulated erosion events.

### 3.3.2 Sediment concentration

The mean sediment concentrations ($\bar{c}$) and the mean rates of soil loss ($C_L$) for the erosion experiments are presented in Table 3.2. Similar $\bar{c}$ were observed for events #1 and #2, and for events #3 and #4. $C_L$ was different for each event except for events #1 and #3 which were similar.

Table 3.2

Mean sediment concentration ($\bar{c}$, kg, $m^{-3}$) and mean rate of soil loss ($C_L$, kg $m^{-2}$ $min^{-1}$) for the erosion events. In parenthesis are the SE of the means (n = 6 for event #1, and n = 8 for the other events).

| Erosion event | $\bar{c}$ (kg $m^{-3}$) | $C_L$ ( kg $m^{-2}$ $min^{-1}$) |
|---|---|---|
| #1 | 5.0 (0.62) | 0.011 (0.0014) |
| #2 | 4.7 (0.54) | 0.006 (0.0005) |
| #3 | 13.4 (5.30) | 0.013 (0.0057) |
| #4 | 16.4 (1.09) | 0.024 (0.0020) |

Sediment concentration ($c$) varied little over time for all events, which was in agreement with values of $\bar{c}$ (Table 3.2), except during the first five minutes of event #3, when $c$ was higher than at other times (Fig. 3.2). When slope angle and length are relatively constant (as in the field experiments), an increase in runoff rate is associated with an increase in stream power Sediment concentration is expected to increase when there is an increase in stream power (Proffitt *et al.*, 1993; Rose, 1993). However, results of Figs. 3.1 and 3.2 show that a high value of Q' was not associated with a high value of *c*. Event #1 had the highest Q', but not the highest *c*; similarly, event #4 had the highest *c* associated with a medium value of Q'. No rills were observed during erosion for any event

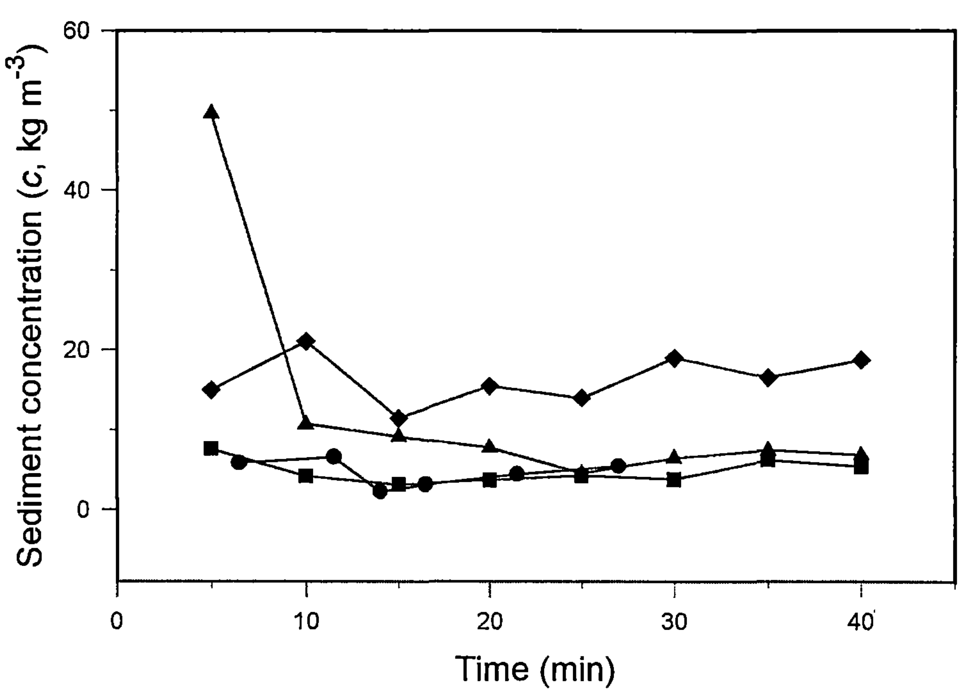


**Fig. 3.2** - Variation of sediment concentration (*c*) with time from runoff in various erosion events (#1 to #4) under simulated rainfall. #1 -● , #2 -■ , #3 -▲ ,#4 -◆ .

A change in sediment concentration at the beginning of an erosion event may depend on the sensitivity of aggregates to wetting and ASM. When subjected to rainfall, some pre-wetted soils can develop a greater resistance to erosion than if they were dry. This may affect changes in *c* in the first few minutes of erosion events (Luk and Hamilton, 1986; Le Bissonnais and Singer, 1992; Govers and Loch, 1993; Ghidey and Alberts, 1994; Zeng *et al.*, 1994; Le Bissonnais *et al.*, 1995). This is due to a short-term increase in aggregate stability of pre-wetted soils compared with dry soils. Such short-term changes tend to disappear some time after erosion, i.e., when the soil becomes saturated (Le Bissonnais and Singer, 1992; Le Bissonnais *et al.*, 1995; Govers and Loch, 1993; and Ghidey and Alberts, 1994).

### 3.3.3 Erosion in relation to soil strength

In addition to runoff rate, soil strength can affect erosion significantly (Cruse and Larson, 1977; Watson and Laflen, 1986; Bradford *et al.*, 1987; Rose *et al.*, 1990; Misra and Rose, 1995, Hanson, 1996). Values of penetrometer resistance ($P_a$) and shear strength ($\tau_a$) after rain were lower than those before rain ($P_b$ and $\tau_b$) (Table 3.3). A reduction in soil strength could be due to an increase in water content (Luk and Hamilton, 1986; Causarano, 1993; Zeng *et al.*, 1994; Hanson, 1996) as well as due to formation of a deposited layer of low strength following erosion (Rose *et al.*, 1990; Hairsine and Rose, 1991, 1992a). However, shear strength decreased significantly with rain but not penetrometer resistance. These differences could be due to the method of measurement and variability related with the method of measurement. While the number of measurements were similar for both measures of strength, the variability of penetrometer resistance was much greater than that of shear strength (Table 3.3). Penetrometer resistance measurements were made over a smaller area but at greater depths than those of shear strength. Therefore, measurements of $P_b$ and $P_a$ were unlikely to reflect any reduction in soil strength near the surface arising from deposition of eroded material or breakdown of aggregates due to rain. Nevertheless, both measurements indicated a decrease in soil strength after erosion which could simply be due to an increase in water content (Causarano, 1993). The order of both Torvane and penetrometer readings for various simulated events were similar before and after rain, with minor exceptions: #1 > #2 > #3 > #4 (Table 3.3).

Table 3.3

Penetrometer resistance (measured with a pocket penetrometer, $P_b$ and $P_a$) and shear strength (measured with a Torvane, $\tau_b$ and $\tau_a$) of soil before and after simulated rainfall events. Mean values with standard errors (in parenthesis) are shown for n measurements.

| | Penetrometer resistance (kPa) | | | Shear strength (kPa) | | |
|---|---|---|---|---|---|---|
| Event | Before, $P_b$ | After, $P_a$ | n | Before, $\tau_b$ | After, $\tau_a$ | n |
| #1 | 2078 (317) | n.m. | 16 | n.m. | n.m. | |
| #2 | 1436 (159) | 1186 (181) | 10 | 9.51 (0.63) | 7.20 (0.34) | 10 |
| #3 | 757 (107) | 546 (188) | 10 | 8.38 (0.08) | 7.67 (0.22) | 10 |
| #4 | 242 (170) | 156 (67) | 10 | 5.32 (0.09) | 2.49 (0.21) | 10 |

n.m. - not measured

Variation in sediment concentration with measurements of soil strength (Fig. 3.3) showed that *c* decreased with increasing $P_b$, $P_a$, $\tau_b$, and $\tau_a$. These results are in agreement with current interpretation of the role of soil strength in soil erosion (Rose *et al*, 1990).

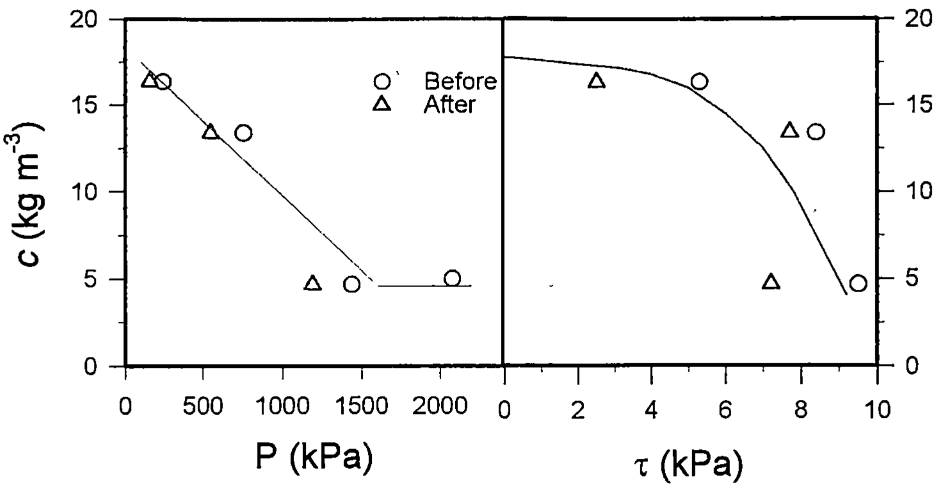


**Fig. 3.3** - Variation in average sediment concentration (*c*) with variation in penetrometer resistance (P) and shear strength ($\tau$) before and after erosion.

Several authors have explored the possibility of relating erosion with soil strength (Cruse and Larson, 1977; Al-Durrah and Bradford, 1982; Watson and Laflen, 1986; Rose *et al.*, 1990; Bradford and Huang, 1995; Misra and Rose, 1995; Hanson, 1996; Le Bissonnais, 1996). The results of this study suggest that soil strength may be used as an index of soil erodibility, in conjunction with other soil physical conditions (namely, ASM). ASM affects erosion by influencing soil strength and the aggregate stability of soils. Therefore ASM should be considered as a factor when considering indicators of soil erodibility (Luk and Hamilton, 1986). Hanson (1996) related soil strength indices and erodibility for varying bulk density and water content, and concluded that soil strength indices alone provide poor indicators for erosion, which strengthens the concept that an index of erodibility based on soil strength should be combined with other soil properties.

### 3.3.4 Size-distribution of soil and sediment

The composition of uneroded soil and sediment for the various events is presented in Fig. 3.4. For each experiment, size distribution of sediment is given as an average for various sampling times to indicate the approximate range of variation. Similar distribution for the uneroded soil is given as an average of three replicates (except for event #4 which had only one replicate). For event #2, the composition of the uneroded soil was assumed to be the same as for event #3 because these two events were only seven days apart.

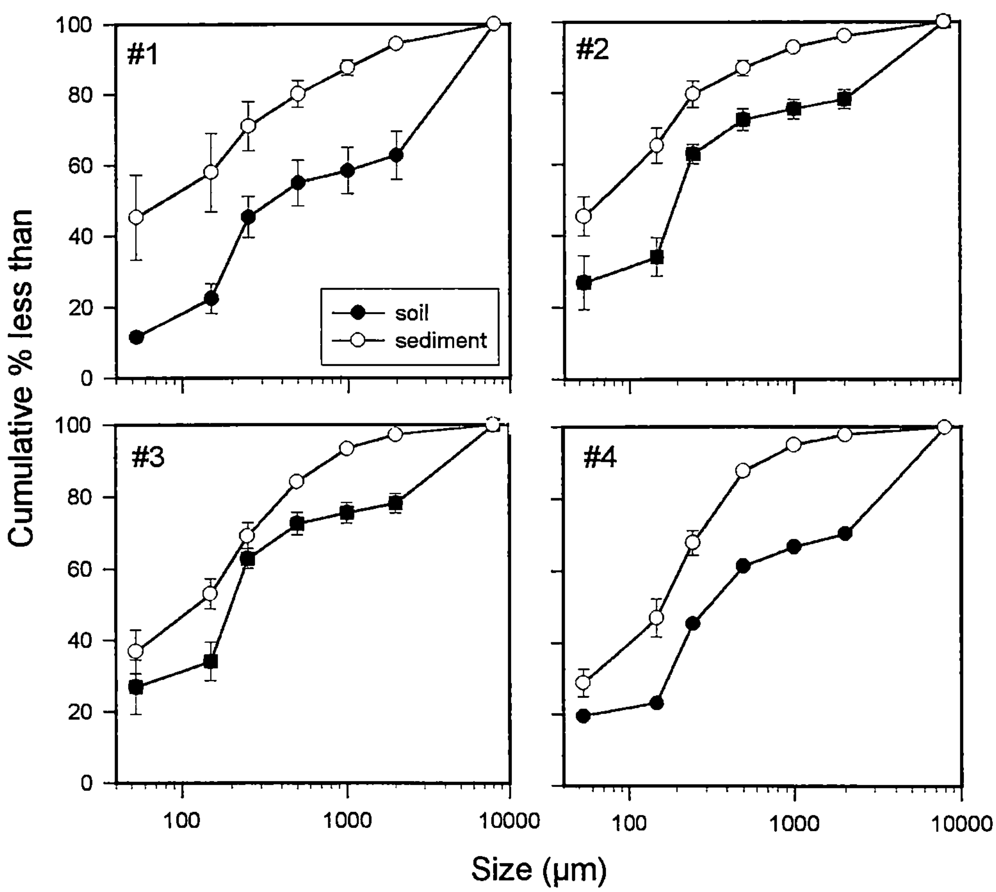


**Fig. 3.4** - Size-distribution of uneroded soil and sediment for simulated erosion events. Size-distribution of the sediment was averaged over 27 minutes of erosion (6 sampling times) for event #1, and 40 minutes of erosion (8 sampling times) for all other events. Vertical bars indicate standard errors.

Mean Weight Diameter (MWD) of uneroded soil and sediment at various sampling times is given in Fig. 3.5. For all experiments, uneroded soil was coarser than sediment (Figs. 3 4 and 3.5). Although the cumulative size-distribution of uneroded soil appeared to be similar for all events (Fig. 3.4), the MWD of uneroded soil (described in Section 3.3 1 and Fig 3 5) varied significantly between events. These changes in MWD of soil may be due to a possible change in soil structure with time, the amount of plant residue, and any spatial variation in soil properties. Despite these differences in MWD of the uneroded soil with time, the sediment composition remained always finer than the uneroded soil both within and between erosion events (Fig. 3.5). The results of Figs. 3.4 and 3.5 at least agree partly with previous work which indicated that eroded sediment was finer than the uneroded soil in the first few minutes of erosion (Walker *et al.*, 1978; Moss *et al.*, 1979; Palis *et al.*, 1990a). However, there was little tendency for the sediment to become similar in composition to the uneroded soil, with increasing time of erosion.

### *3.4 CONCLUDING REMARKS*

Results from this series of *in-situ* erosion events under simulated rainfall indicated that, for a given soil, the differences in sediment concentration between events can not be adequately explained by the differences in rainfall and runoff rates. Other factors such as the duration and intensity of natural rainfall events between experimental erosion events affecting ASM and soil structure (MWD) of the uneroded soil may have some influence on erosion In addition, there was some possibility of spatial variation in soil properties which could have affected the results as the experimental events were never carried out at exactly the same spot. As it was difficult to replicate these events and to control hydrological variables in temporarily set-up runoff plots, these experiments provided impetus for considerable improvements in future experiments (in forthcoming Chapters). There were good correspondences between average sediment concentration and various measures of soil strength, which indicated that erosion may be predicted reasonably well for poorly structured soils using soil strength as an index of soil erodibility These possibilities are further explored in erosion experiments with several forest soils with contrasting structure and outlined in subsequent Chapters

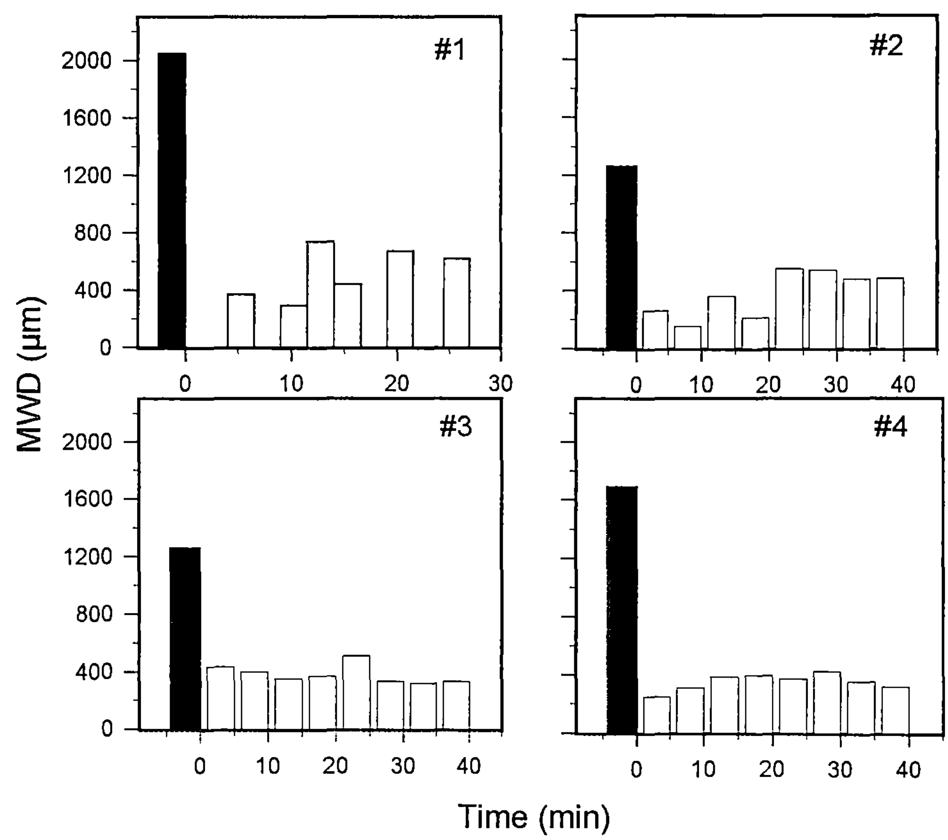


**Fig. 3.5** - Mean Weight Diameter (MWD) of uneroded soil (black) and sediment (white) at various sampling times.

# Chapter 4

## EROSION AND SEDIMENT CHARACTERISTICS

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### *4.1. INTRODUCTION*

Transport of chemicals by erosion is dependent on the size-distribution of sediment and sediment dynamics (Rhoton *et al.*, 1983). This dependency is because first, the distribution of sorbed chemicals (eg. nutrients and pesticides) over the size of sediment is non uniform, second, the fine sediments tend to be richer in soil-sorbed chemicals than coarse sediments, and third, some change in size-distribution of sediment is expected with the duration of erosion (Palis *et al.*, 1990a; Ghadiri and Rose, 1991b). Sediment is known to be richer than the soil from which it originated in the content of clay (Rhoton *et al.*, 1979), proportion of small aggregates (Alberts and Moldenhauer, 1981), and content of phosphorus, nitrogen, and organic carbon (Stoltenberg and White, 1953; Sharpley, 1980; Palis *et al.*, 1990 a,b).

The size-distribution of sediments in an erosion event may depend on how well a soil is aggregated and the characteristics of the erosion event, which can influence the extent to which soil aggregates are broken down. The extent of aggregation in a soil depends on the energy of bonding of aggregates (So *et al.*, 1996) and the characteristics of the erosion event may be quantified in terms of the amount of energy available and the manner in which this energy aids fragmentation of aggregates. The overall effects of the interaction of the available energy for erosion, and the degree of soil aggregation together with deposition may explain the variation in soil loss as well as size-distribution of sediments (Loch and Donnollan, 1982; Truman *et al.*, 1990; and Roth and Eggert, 1994).

Information on erosion from forest soils is limited and, in particular, there is a dearth of information on erosion after managed tree farm operations. A tree farm or a plantation is essentially bare and prone to erosion for a period after clearing and site preparation. Site preparation in plantations usually includes formation of mounds with a typical slope of 16 °. Mounds may contribute to localised erosion, and when such mounds are on a slope and their orientation is up-and-down the slope, the scope for erosion of bare soil is enhanced further.

When site features and erosion events are similar, soil loss and/or sediment characteristics may still differ between consecutive events. Depending on weather conditions between events, a recently cultivated soil may gain strength which may be due to consolidation or hardening of the surface soil. The number and duration of wet and dry cycles may also influence soil aggregation (Hussein and Adey, 1995). Changes in soil strength and aggregation may both influence the size-distribution of uneroded soil and sediment and the amount of soil loss. In this study, erosion and sediment size-distribution of soil are compared for three forest soils and for situations with and without a drying cycle between two consecutive erosion events.

The aim of this study was to quantify erosion and sediment characteristics of cultivated forest soils, and to interpret these in terms of the mechanical stability of soils. Indices of mechanical stability are derived to determine the susceptibility of soils to erosion.

## 4. 2. METHODS

The details of erosion experiments conducted in the laboratory for the cultivated soils D, R, and M, were given in Chapter 2. These soils were collected from the top 0.2 m of recently cultivated eucalypt plantation sites. It should be noted that soil D was a poorly aggregated loamy sand, soil R a strongly aggregated clay, and soil M a clay loam with low aggregate stability when wet. These soils were exposed to simulated rain of constant rate as described in Chapter 2. The details of erosion treatments are given below.

### 4.2.1 Erosion treatments

For each of the three soils, four erosion treatments were applied, and each combination of erosion treatment and soil was replicated three times. The treatments were:

Treatment 1: with low rainfall KE and low slope (2 °),

Treatment 2 with high rainfall KE and low slope, and

Treatment 3 with high rainfall KE and high slope (16 °).

Treatment 3 was sub-divided into treatments 3a and 3b with identical conditions of rainfall KE and slope but with and without a drying cycle between two consecutive simulations.

For treatment 3b, soil was dried under shade for 15 days following treatment 3a. These four treatments permitted comparison of erosion due to varying KE at constant slope, varying slope at constant KE and with or without a drying cycle. It should be noted that the high slope treatment (16 °) corresponded with the slope of a cultivated mound measured for the plantation at the Dover site The downslope length of the erosion bed was 1 m for all experiments. Soil in an erosion tray was used for a maximum of four simulations (during which there was no disturbance of the soil surface), and the erosion treatments were chosen randomly, after which the soil in the erosion tray was replaced. Whenever erosion treatment 3 was applied to a soil, treatment 3a (no drying cycle) was always preceded by 3b (after a drying cycle).

**4.2.2 Data analysis**

Prior to each replicated erosion event, wet uneroded soil was sampled from the buffer area of erosion trays (Section 2.4.4) and its size-distribution was determined without dispersion (subsample 1) or after mechanical and chemical dispersion (subsample 2). Some samples of dry soil were also sieved to determine its size-distribution. The size-distribution of wet sediment samples was determined on selected occasions (details in Section 2.4.4). Mean Weight Diameter (MWD, equation 3.1, Chapter 3) was calculated from the data on size-distribution for soil and sediment samples

*4.2.2.1 Indicators based on MWD*

Taking aggregate size-distribution obtained with dry sieving as the reference, or its corresponding value of MWD, the susceptibility of soils to wetting ($S_W$) may be expressed as

$$S_W = \frac{MWD_d - MWD_w}{MWD_d}, \qquad (4.1)$$

where $MWD_w$ and $MWD_d$ are MWD of the wet soil without dispersion and of the dry sieved soil, respectively. In a similar way, when a soil is dispersed with application of energy in excess of that required to hold aggregates together (i.e. equivalent to bonding energy) in air-dry state, the susceptibility of soil to wetting and dispersion ($S_{WD}$) may be estimated as

$$S_{WD} = \frac{MWD_d - MWD_w'}{MWD_d}, \qquad (4.2)$$

where $MWD_w'$ corresponds to wet soil after dispersion. The difference between $S_{WD}$ and $S_W$ ($S_D$) gives an indication of the susceptibility of soil to dispersion alone. $S_{WD}$ is likely to be >> $S_W$ for any soil.

*4.2.2.2 Energy associated with runoff*

Stream power ($\Omega$, W m$^{-2}$) is the energy of runoff per unit area, some or all of which may be available to remove and transport aggregates from the erosion surface. If there are no rills present on the erosion surface, $\Omega$ for sheet flow of runoff water is

$$\Omega = \rho g S q, \qquad (4.3)$$

where $\rho$ is the density of runoff water (assumed to have a constant value of 1000 kg m$^{-3}$), $g$ the acceleration due to gravity (i.e. a constant of 9.8 m s$^{-2}$), $S$ the sine of the inclination of erosion surface, and $q$ is the volumetric flux of runoff per unit width of erosion surface ($m^3$ $m^{-1}$ $s^{-1}$). Note that $q$ is the product of runoff rate (Q, m $s^{-1}$) and the downslope distance.

## *4.3. RESULTS*

### 4.3.1 Rainfall rate, water content, and bulk density

Analysis of variance for rainfall rate (116 mm $h^{-1}$) indicated that rainfall rate did not differ for various soils and erosion treatments. Mean water content (% by weight ± SE) of prewetted soils before simulations was 34.5 ± 8.3, 55.6 ± 10.0, and 26.2 ± 6.2, for soils D, R and M, respectively. Mean bulk density (± SE) before simulations was 1.06 ± 0.04, 0.63 ± 0.02, and 1.12 ± 0.05 Mg $m^{-3}$ for soils D, R and M, respectively. Water content and bulk density of soil varied significantly with each soil ($p \leq 0.001$). Soil water content also varied with erosion treatment ($p \leq 0.001$).

### 4.3.2 Erosion, runoff, and sediment concentration

Mean values of total soil loss ($E$, kg $m^{-2}$) are given in Table 4.1. Total soil loss was significantly influenced by soil ($p \leq 0.001$), erosion treatment ($p \leq 0.001$) and there was also a significant interaction between soil and erosion treatment ($p \leq 0.001$). As the soil loss for one of the soils (R soil) was much smaller than for the other two soils, the data were not normally distributed. Analysis of variance for comparison of treatments was therefore based on log-transformed data. For all erosion treatments, $E$ was in the order M > D > R. $E$ generally increased with increasing slope and kinetic energy of rain. However, for soil R, erosion treatments had little effect on $E$. This may have been partly due to the lack of sufficient overland flow during erosion compared to interflow (*i.e.* subsurface flow along the slope) for this soil. For soils D and M, erosion after a dry cycle was significantly lower than after a wet cycle. Wet and dry cycles had no significant effect on erosion for soil R.

Table 4.1

Soil loss ($E$, kg m$^{-2}$) measured for simulated rainfall events of 40 min duration using a constant rainfall rate of 116 mm h$^{-1}$, for three soils (Dover = D, Ridgley = R, and Maydena = M). Values of soil loss followed by the same letter(s) are not significantly different at $p \leq 0.05$.

| | Soil loss (kg m$^{-2}$) for soils | | |
|---|---|---|---|
| Erosion treatments | D | R | M |
| 1. low energy, low slope | 0.076 a | 0.003 a | 0.110 ab |
| 2. high energy, low slope | 0.401 b | 0.077 a | 0.972 c |
| 3a. high energy, high slope, wet | 4.906 f | 0.087 a | 7.306 g |
| 3b. high energy, high slope, dry | 2.714 d | 0.170 ab | 3.272 e |

Mean values of runoff rate (Q', mm h$^{-1}$) are given in Table 4.2 (note that runoff rate is denoted as Q when measurement units are in m s$^{-1}$, and as Q' when in mm h$^{-1}$). Q' was significantly affected by soil type (p= 0.001) and erosion treatments (p=0.008). Q' was significantly lower for R soil than for the other soils. Treatment 3b had significantly lower Q' than for the other treatments. Q' was always lower than rainfall rate (P), although the soils were wetted to an apparent saturation, prior to simulations. However, all soil in the erosion tray was not saturated as shown by the data on soil water content before erosion (detailed in Chapter 5). Another reason for Q' being lower than P was the positioning of the rainfall collector for measuring rainfall rate, which was located on top of the erosion trays. This position allowed more rainfall to be collected than the rainfall that actually reached the tray.

Table 4.2

Mean runoff rate (Q', mm $h^{-1}$) for various soils and erosion treatments. Values of runoff followed by same letter(s) are not significantly different at $p \leq 0.05$.

| Soil | Q' (mm $h^{-1}$) | Treatment | Q' (mm $h^{-1}$) |
|---|---|---|---|
| D | 95.2 b | 1 | 98.4 b |
| R | 84.7 a | 2 | 93.9 b |
| M | 95.3 b | 3a | 94.2 b |
| | | 3b | 80.4 a |

Sediment concentration generally remained high in the beginning of erosion events and then gradually reduced until an equilibrium reached in about 20-25 minutes. Sediment concentration for R soil showed little change with time. Details of these are given in Chapter 5.

### 4.3.3 Energy available for erosion

The main sources of energy available for erosion arose from rainfall and/or runoff. Due to the nature of treatments used in this study, the energy arising from rainfall and runoff varied considerably. As previously stated, the KE of rain was greater for treatments 2 and 3 than for treatment 1. Thus in terms of energy available for erosion due to rainfall, treatment 1 had low energy, whereas treatments 2 and 3 had high energy.

Table 4.3

Mean values of stream power (Ω) estimated for various simulated erosion events. Standard errors are shown in parenthesis for 3 replicated events.

| Erosion Treatment | Stream power (Ω, W $m^{-2}$) for | | |
|---|---|---|---|
| | Soil D | Soil R | Soil M |
| 1 | 0.0097 | 0.0089 | 0.0096 |
| | (0.0003) | (0.0002) | (0.0005) |
| 2 | 0.0092 | 0.0091 | 0.0085 |
| | (0.0003) | (0.0001) | (0.0009) |
| 3a | 0.0759 | 0.0634 | 0.0733 |
| | (0.0021) | (0.0032) | (0.0021) |
| 3b | 0.0605 | 0 0496 | 0.0712 |
| | (0.0034) | (0.0068) | (0.0015) |

The Ω for erosion treatments at low slope (treatments 1 and 2) was ~ 0.009 W $m^{-2}$ (Table 4.3), i.e. close to the threshold stream power reported by Proffitt *et al.* (1991). The threshold stream power is the minimum amount of stream power required to entrain sediment from a soil bed. Thus, in the low slope treatments for all soils, rainfall may have been more effective than overland flow in causing erosion. The Ω for high slope (treatment 3) varied from 0.05 to 0 08 W $m^{-2}$, which was well above the threshold stream power value. Thus, for treatment 3 both runoff and rainfall would have contributed to erosion. In terms of energy available for erosion due to overland flow alone, treatments 1 and 2 had low energy and treatment 3 high energy.

In terms of total energy available for erosion due to rainfall and runoff, treatment 1 with low rainfall KE and Ω close to the threshold value was a low energy event; treatment 2 with high rainfall KE, but still low Ω can be regarded as a medium energy erosion event; and treatment 3 with high rainfall KE and $\Omega$ above the threshold value, was a high energy erosion event. The relative magnitude of energy available for various erosion events described here is used as a determinant of the size characteristics of the sediments in the following section

**4.3.4 Relative mechanical stability of soils**

The size-distribution of the wet uneroded soils with and without dispersion, and that due to dry sieving is given in Fig. 4.1. The relative difference in size-distribution arising from these methods of separation may be taken as an indicator of the relative mechanical stability of soils when they are simply wetted or dispersed after wetting. MWD calculated for each of the curves in Fig. 4.1 are given in Table 4.4 with the indices of relative mechanical stability. It should be noted that the calculated value of MWD declines if a given size-distribution is dominated by small aggregates.

Values of $MWD_w'$ (after wetting and dispersion) for soils was in the order D > M = R, while $MWD_w$ (after wetting only) was in the order R > D = M, and $MWD_d$ (after dry sieving) was R > M > D. The indices of relative mechanical stability calculated for these soils showed soil M to be most susceptible to wetting (highest $S_W$) and soil R to dispersion alone (highest $S_D$). Both soils R and M were equally susceptible to wetting and dispersion, and more susceptible than soil D.

Table 4.4

Mean Weight Diameter (MWD, μm) of the uneroded soil with dispersion ($MWD_w$', n = 24 for each soil) and without dispersion ($MWD_w$, n = 3 for soils D and M, and n = 4 for soil R ), and after dry-sieving ($MWD_d$, n = 3 for each soil). Standard errors are given in parentheses for n replicate measurements. $S_W$, $S_{WD}$, and $S_D$ are, respectively, indicators of the susceptibility of soils to wetting, wetting and dispersion, and dispersion.

| Soil | $MWD_w$' | $MWD_w$ | $MWD_d$ | $S_W$ | $S_{WD}$ | $S_D$ |
|---|---|---|---|---|---|---|
| D | 138 | 762 | 1054 | 0 28 | 0.87 | 0.59 |
| | (3) | (45) | (42) | | | |
| R | 39 | 4036 | 4226 | 0 05 | 0.99 | 0.94 |
| | (<1) | (67) | (79) | | | |
| M | 52 | 694 | 3002 | 0.77 | 0.98 | 0.21 |
| | (2) | (51) | (17) | | | |

Table 4.5

Mean Weight Diameter (MWD, μm) of wet sediments without dispersion as influenced by erosion treatments and soil types. Values of MWD followed by same letter(s) are not significantly different at $p \leq 0.05$.

| Erosion treatment | MWD of the sediment from | | |
|---|---|---|---|
| | Soil D | Soil R | Soil M |
| 1 | 273 a | 1918 c | 166 a |
| 2 | 1136 b | 2697 d | 348 a |
| 3a | 303 a | 3711 e | 220 a |
| 3b | 366 a | 3541 e | 314 a |

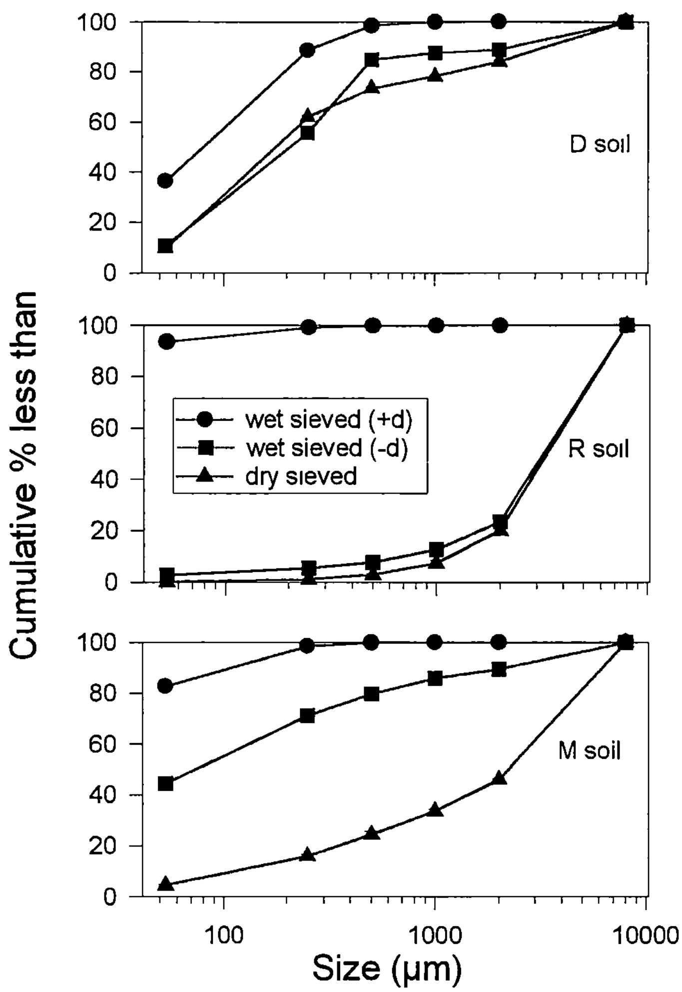


**Fig. 4.1** - Size-distribution of the uneroded soil as influenced by various methods of sieving and dispersion. The notations '+d' and '-d', respectively, denote wet sieving with and without dispersion. Vertical error bars represent SE of mean values; however, these were often smaller than the size of symbols.

### 4.3.5 Size-distribution of sediment

The size-distribution of sediments measured for 0-5, 12.5-17.5 and 35-40 minutes after runoff was used to calculate MWD of sediments. Analysis of variance showed no significant difference in MWD with time of sampling suggesting that size-distribution of sediment was reasonably constant over the entire duration of the erosion events. However, there was a significant difference in MWD for soils and erosion treatments ($p < 0.001$). It is interesting to note that the effect of various erosion treatments on soil loss (Table 4.1) was very different from the effect of the same treatments on MWD of sediments (Table 4.5). Soil M had significant differences in soil loss due to erosion treatments, but these treatments had no effect on MWD of sediment. In contrast, soil loss for R soil was not influenced by erosion treatments, but the MWD of sediment varied considerably with erosion treatments, except for treatments 3a and 3b. Soil loss from D soil was significantly influenced by erosion treatments but the MWD of sediments for this soil did not vary as much with erosion treatments. The wetting and drying cycle did not influence MWD of sediment significantly for any of the soils used in this study. The MWD of sediments in Table 4.5 was generally lower than the MWD of the uneroded soil obtained with dry sieving or that when soils were wet without dispersion (Table 4.4). However, the MWD of sediments for all soils and all erosion treatments was greater than that obtained for wet uneroded soil after dispersion, indicating that sediments were coarser than the uneroded soil after dispersion. MWD of sediment arising from soil D for treatment 2 was unexpectedly higher than the MWD for other erosion treatments or for uneroded soil (Tables 4.4 and 4.5). This would suggest that for this soil and treatment, coarse fractions present in the sediment may not have been properly accounted for during sampling of the uneroded soil.

## *4.4. DISCUSSION AND CONCLUDING REMARKS*

### 4.4.1 Erosion and runoff in relation to size-distribution of uneroded soil

Soil loss was the lowest for R soil (Table 4 1) which had a larger proportion of large, strong aggregates (Fig. 4.1) or a higher $MWD_w$ (Table 4.5) than the other two soils. As erosion is likely to be small for soils with higher $MWD_w$, these results show that lower soil losses due to higher $MWD_w$ are likely to be associated with strongly aggregated soils. This is also supported by previous studies on forest soils which show that erosion is strongly influenced by aggregates of > 2 mm in diameter (Farmer and van Haveren, 1971).

Apart from MWD, strongly aggregated soils (e.g. soil R) may also influence runoff characteristics, partly due to increased surface roughness and the presence of macropores influencing infiltration, compared with poorly aggregated soils Although all soils were pre-wetted in an identical manner before simulated erosion events, runoff rates for R soil were significantly lower than for the other soils (Table 4.2), and there was also a significant reduction when the soils were dried for a fortnight (treatment 3b, Table 4.2). Drying of a saturated soil and rewetting of that soil prior to erosion may have altered total porosity (Horn *et al.*, 1994) leading to incomplete saturation during pre-wetting.

The magnitude of erosion is often attributed to soil texture, particularly the clay content of soil. However, there are conflicting reports when erosion is related to clay content. For example, erosion has been shown to be inversely related with clay content (Bubenzer and Jones, 1971; De Ploey and Poesen, 1985; Schjønning, 1994), whereas Meyer and Harmon (1989) did not find any relationship of erosion with clay content. Thus, clay content of a soil may not be a suitable indicator of susceptibility to erosion when a wide range of soils is considered. In this study, erosion did not relate well with clay content.

There have been also attempts to relate erosion or erodibility of soils with aggregate size-distribution obtained for wet uneroded soils. Once again, there does not appear to be a universal relationship between erosion and aggregate size-distribution (Bryan, 1976; Young and Onstad, 1978; Luk, 1979; Bradford *et al.*, 1987). Aggregate size-distribution can be quantified in several ways. Mean Weight Diameter is one of the widely used indices of size-distribution. Despite the lack of a good relationship between size-distribution and erosion, the value of MWD calculated from published data (Alberts *et al.*, 1983; Mitchell *et al.*, 1983; Luk and Hamilton, 1986) showed good agreement with the resistance of soils to erosion: higher MWD was always related with lower erosion. The values of MWD calculated for the soils used in this study (wet sieved with no dispersion) were 762, 4036 and 694 μm, respectively for soils D, R and M. It can be seen from Table 4.1 that erosion for these soils was M > D > R, indicating a general support for the use of MWD as an indicator of the resistance of soils to erosion. The correspondence between the resistance to erosion and MWD is also supported by theories for modeling erosion processes (Rose, 1993). Such theories acknowledge that for soils which consist of large, stable aggregates, erosion is likely to be reduced as these soils tend to have higher depositability. More recently, Amezketa *et al.* (1996) found good correlations between soil erosion and stability parameters derived for soils using the MWD after fast wetting and MWD of soils stirred after wetting, which gives strength to the use of MWD and related parameters as indices of soil erosion.

### 4.4.2 Size-distribution of sediment in relation to relative mechanical stability of soils

The transportability of sediment depends on its size, as fine sediment produced during erosion can travel a long distance from the point of its origin compared with coarse sediment. Soils which have strong and stable aggregates are likely to have greater resistance to erosion because they may not break down easily under the impact of raindrops as well as shear stress associated with runoff. Aggregates of such soils may also break into few large aggregates during erosion such that sediment will have a lower transportability compared with soils which tend to slake and/or disperse easily.

In most erosion events (both natural and simulated), finer sediment is produced in the first few minutes (Walker *et al.,* 1978; Rose *et al.,* 1989; Palis *et al.,* 1990a). As the duration for the first sample is increased beyond 2 minutes (in this study as well as in studies of Ghadiri and Rose, 1991b), the sediment size may reflect the size of material that is produced by net erosion, *i.e.* dependent on the rate of erosion-deposition per unit area of erosion surface Due to the small length of erosion trays used in this study, there may have been little opportunity for deposition, and the depth of water would have been less than the size of raindrops This may explain the lack of a temporal variation in the size-distribution of sediment in this study. Comparison of the size-distribution of sediments or its MWD (Table 4.5) with that for uneroded soil without dispersion (Table 4.4) indicated sediment to be finer in all cases except for soil D with treatment 2. It would appear that erosion in these simulated events may have been dominated by rainfall rather than rainfall and runoff or runoff alone, to produce sediments finer than the uneroded soil.

Most erosion models aim at quantifying the amount of sediment produced per unit area and time (Nearing *et al.*, 1989; Misra and Rose, 1996). However, there is little emphasis on the characteristic of sediment that is produced during erosion The sediment characteristics, i.e. its MWD or size-distribution, may be influenced by the mechanical stability of uneroded soil aggregates and the amount of energy available during erosion, as described below.

The poorly aggregated soil D (indicated by its lower $MWD_d$ in Table 4.4 and high sand content in Table 2 1) was intermediate in soil loss (Table 4.1). This soil was intermediate in its susceptibility to wetting (see values of $S_W$ in Table 4.4) and to wetting and dispersion (low $S_{WD}$ in Table 4 4). The strongly aggregated soil R which had highest $MWD_d$ and $MWD_w$, was least susceptible to wetting (Table 4.4) and produced lowest soil loss (Table 4.1) of coarse sediments, i.e. high MWD (Table 4.5). Although this soil had the highest clay content (Table 2.1), the energy available during erosion from rainfall and runoff was not sufficient to overcome the bonding energy of aggregates to cause dispersion. As this soil was most resistant to disaggregation arising from wetting alone, the MWD of the sediment remained nearly as high as for the uneroded soil (Tables 4.4 and 4.5). Soil M was most susceptible to wetting (Table 4.4) which enhanced soil loss, irrespective of erosion treatments (Table 4.1) and produced finest sediment (Table 4 5) despite high $MWD_d$ (Table 4.4) Thus, change in relative mechanical stability of soils due to wetting and dispersion can be regarded as a suitable indicator for quantity and quality (i.e size-distribution) of sediment produced during erosion.

It is obvious from this work that the precise amount of energy required for disaggregation of soils is not known and also the amount of energy from rain and runoff used in erosion. The latter (i e. energy from rainfall and runoff) are difficult to measure directly and are available only as theoretical estimates (Proffitt, *et al.*, 1993; Misra and Rose, 1996). However, it is possible to measure energy applied to soil suspension precisely (Raine and So, 1993). A possible extension of this work will be to determine the susceptibility of soils to wetting and dispersion in terms of the quantity of applied energy required to disaggregate soils, so that suitable indices of erosion and sediment quality can be developed without the need of direct measurement of erosion and sediment quality.

# Chapter 5

## THE EFFECTS OF SOIL STRUCTURE AND STRENGTH ON EROSION AND ERODIBILITY OF SOIL

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### *5.1 INTRODUCTION*

The magnitude and temporal variation of sediment concentration ($c$) during an erosion event may be explained by the simultaneous action of five processes (details in Appendix): detachment and re-detachment by rainfall; entrainment and re-entrainment by runoff; and deposition (Misra and Rose, 1989; Hairsine and Rose, 1991; Hairsine and Rose, 1992a, b; Rose *et al.*, 1993; Misra and Rose, 1996). Here, detachment and entrainment respectively refer to the erosion of uneroded soil due to rainfall and runoff. Re-detachment and re-entrainment refer to similar processes when rainfall and runoff are acting on the deposited material. Deposition is the process of settling of sediment within the runoff water. The distribution of settling velocity indicates the proportion of soil or sediment by mass that can settle at a given velocity. The average settling velocity for a given distribution is referred as *depositability* ($\phi$)

The erosion model GUEST (Griffith University Erosion System Template) considers these erosion processes to interpret data on sediment concentration from bare soil in single erosion events and to provide erodibility parameters (Misra and Rose, 1989, 1996) For situations where stream power ($\Omega$) (*i.e.*, proportional to the product of slope, runoff rate and slope length; exact expression in equation 4.3, Chapter 4) arising from runoff does not exceed a threshold stream power ($\Omega_0$), the data on sediment concentration may be interpreted in terms of rainfall-driven erosion processes and deposition, thus yielding erodibility parameters $a$ (rainfall detachability) and $a_d$ (rainfall re-detachability). When $\Omega > \Omega_0$, the data on sediment concentration is interpreted in terms of all the five processes mentioned earlier and yields an additional erodibility parameter ($J$, J kg$^{-1}$) which is the specific energy of entrainment. $J$ is the amount of energy required to entrain a unit mass of uneroded soil and the sediment concentration for a given value of $J$ is referred to be at the source limit ($c_s$). In some situations, $J$ can be $\approx 0$ indicating the sediment concentration to have reached transport limit ($c_t$) Such a situation is likely to occur only for a short period of time, or even not occur at all during an erosion event (Hairsine and Rose, 1992a, b) unless there is a supply of sediment from erosion processes that is not explicitly included in the model. The possibility of sediment concentration reaching the transport limit arises when erosion is dominated by rilling and frequent collapse of rill banks covering the entire active erosion surface with a deposited material that can be readily re-entrained.

When detailed information on $c$ during an erosion event is unavailable, an approximate erodibility parameter, $\beta$, can be determined from the data on average sediment concentration for the entire event (Misra and Rose, 1989). This parameter neglects the effects of rainfall-driven erosion processes on $c$, and it essentially quantifies the extent to which the average sediment concentration is less than the sediment concentration at the transport limit.

There are some indications that the erodibility parameters obtained from GUEST ($a$, $a_d$, $J$ and $\beta$) may be affected by soil strength (Misra and Rose, 1995), although a definitive correspondence between erodibility parameters and any measure of soil strength is yet to be established. These authors noted that soil structure (aggregation) may have some additional influence on erosion and erodibility. There have been also other reports indicating direct or indirect effects of soil strength on erosion and erodibility (Cruse and Larson, 1977; Al-Durrah and Bradford, 1982; Watson and Laflen, 1986; Rose *et al.*, 1990; Bradford and Huang, 1995; Hanson, 1996; Le Bissonnais, 1996).

When a soil is exposed to repeated erosion events, it is subjected to successive cycles of wetting and drying. Wetting and drying cycles may modify soil structure leading to changes in erosion and the erodibility of soils. When a dry soil is wetted, the entry of water into the soil may produce internal and external forces promoting aggregate breakdown. Internal forces in a soil due to entry of water may arise from the release of entrapped air, differential swelling of clay minerals, slaking and dispersion of clay particles (Marshall *et al.*, 1996b). External forces during wetting of a soil by rain include the impact of raindrops and of overland flow which may cause aggregate breakdown Aggregate breakdown can change soil structure by modifying the size and stability of aggregates, and soil porosity Upon drying, soil material may reorganise leading to shrinkage and cracking of clays, and coalescence of neighbouring soil material. Small particles may clog soil pores and a crust can form on the surface of soil which could reduce the infiltration rate and runoff, thus affecting erosion (Le Bissonnais *et al.*, 1989, 1993; Loch, 1994; Bresson and Moran, 1995; Marshall *et al.*, 1996b).

The work reported in this Chapter is aimed at examining the effects of soil physical condition including soil strength and wet-dry cycles on sediment concentration for various soils and erosion treatments. Also examined are the effects of soil strength on the erodibility parameters derived with the steady-state erosion model GUEST.

### *5.2 METHODS*

Three cultivated forest soils (soils D, R, and M) were subjected to simulated rainfall of varying kinetic energy (KE) at varying slopes, and with or without a drying cycle. A description of the soils was given in Section 2.1, and the details of erosion experiments were given in Section 2.4.

Briefly, soils were collected from the top 0.2 m of recently cultivated eucalypt plantation sites and then subjected to simulated rain (116 mm $h^{-1}$) in erosion trays of 0.8 $m^2$. Four erosion treatments were applied to all the three soils, and each combination of erosion treatment and soil was replicated three times. The erosion treatments were. treatment 1 with low rainfall KE and low slope (2 °), treatment 2 with high rainfall KE and low slope, and treatment 3 with high rainfall KE and high slope (16 °). Treatment 3 had two sub-treatments (3a and 3b) that were identical in rainfall KE and slope, but had a drying cycle between them. For treatment 3b, soil was dried under shade for 15 days following treatment 3a. These four erosion treatments permitted the comparison of erosion at varying KE and constant slope, varying slope and constant KE, and with or without a drying cycle.

For each simulated event, several measurements were made and a detailed account of these measurements was given in Section 2.4 For each event, antecedent water content and bulk density were measured on samples of uneroded soil collected from the buffer area of erosion trays. The size-distribution of uneroded soil was obtained by wet-sieving (without mechanical or chemical dispersion) for every event. Immediately before and after simulations, soil strength was measured with a pocket penetrometer (for penetrometer resistance) and a Torvane shear device (for surface shear strength). Volume of runoff water (L) and amount of sediment (kg) were measured every 2.5 minutes during each event of 40 minute duration. Runoff rate and sediment concentration were calculated at each sampling time If there were rills (with an orientation of up-and-down the slope) during an event, these were measured in terms of number of rills per metre width of erosion tray, approximate cross-sectional shape of rill (rectangular or trapezoidal), average depth of rill, and width at the top and bottom of the cross-section of rill. All measurements on rills were made with a mm ruler at the end of each simulated erosion event. Most of the erosion events were continuously recorded with a video camera with time-lag setting, and observations of rills made for each event were validated upon confirmation with video recording. Of all the erosion experiments, there were rills observed for only two replicated events of soil M (treatment 3a). For one of the events, there was one rectangular rill for the 0.8 m wide erosion tray (i.e. 1.25 $m^{-1}$) of average depth 0.005 m and width 0.020 m . For the other event, there were five rectangular rills in the same tray (i.e 6.25 $m^{-1}$) average depth 0.005 m and width 0.045 m.

## *5.3 ESTIMATION OF DEPOSITABILITY AND ERODIBILITY PARAMETERS*

The GUEST model was used for the interpretation of data of sediment concentration as a function of time within an event as well as to interpret average sediment concentration for an event (Misra and Rose, 1989). A number of constants, parameters and input variables is required for the model as listed in Table 5.1. A general description of the model GUEST is presented in the Appendix .

### 5.3.1 Input data for erosion model GUEST

As GUEST is a steady-state model, it uses equilibrium sediment concentration ($c$) as an input variable to estimate erodibility parameters. During an event, equilibrium in sediment concentration is achieved when $c$ varies little with time. For each replicated event of various erosion treatments considered in this study, equilibrium sediment concentration ($c_{eq}$) was taken as the average of $c$ over consecutive sampling times when variation in $c$ was ≤ 10 % with time. Note that $c_{eq}$ chosen in this manner may not indicate the true steady state value of $c$ as the size distribution of sediment could still vary with time. The runoff rate ($Q_{eq}$, $m^3\ m^{-2}\ s^{-1}$) used for analysis with GUEST was calculated using the same sampling times as for $c_{eq}$. Rainfall rate (P, $m^3\ m^{-2}\ s^{-1}$) was taken as the average rate of rainfall measured for the entire simulated event. Values of $c_d$ were the sediment concentration at equilibrium of treatments 1 and 2, where $\Omega \approx \Omega_0$ and therefore, runoff-driven erosion processes were unlikely to have influenced sediment concentration.

For estimation of parameter β, average sediment concentration ($\bar{c}$) and runoff rate $\bar{Q}$ were determined for the entire duration of each erosion event. The value of $\bar{c}$ was calculated as the total amount of sediment (kg) divided by the total volume of runoff water plus sediment ($m^3$), and $\bar{Q}$ was determined as the total volume of runoff water ($m^3$) per unit area divided by the total duration of event.

Table 5 1

Values and explanation of various constants and variables used for the analysis of erosion data with the model GUEST

| Parameters | Value | Source/ Explanation |
|---|---|---|
| CONSTANTS | | |
| Density of water, $\rho$ | 1000 kg $m^{-3}$ | Misra and Rose (1996) |
| Acceleration due to gravity, $g$ | 9 81 m $s^{-2}$ | Misra and Rose (1996) |
| Dimensionless exponent, m | 5/3 | Misra and Rose (1996) |
| Manning's roughness parameter, $n$ | 0.025 $m^{-1/3}$ s | Misra and Rose (1996) |
| Threshold stream power, $\Omega_0$ | 0.01 W $m^{-2}$ | Misra and Rose (1996) |
| Fraction of excess stream power, F | 0.10 | Proffitt *et al.* (1993), Misra and Rose (1996) |
| INPUT DATA | | |
| Wet density of sediment | Equation 5.2 | Loch and Rosewell (1992) |
| Downslope length, L | 1 m | constant for all events |
| Area, A | 0.8 $m^2$ | constant for all events |
| Slope, $S$ | 3.5 and 27.6 % | according to treatments |
| Sediment concentration, $c_{eq}$ or $\bar{c}$ | variable | varied with event |
| Runoff rate, $Q_{eq}$ or $\overline{Q}$ | variable | varied with event |
| Rainfall rate, P | variable | varied with event |
| No. of rills per m width, N | variable | varied with event, rectangular rills only |
| Rill width, W | variable | varied with event |
| [a] Sediment concentration due to net rainfall detachment, $c_d$ | variable | varied with event (see text) |
| Depositability, $\phi$ | variable | varied with soil |

[a] This input variable is required for the calculation of rainfall-driven erodibility parameters and for the interpretation of data on $c_{eq}$

Depositability ($\phi$, m $s^{-1}$) of soil or sediment is defined as:

$$\phi = \sum_{i=1}^{I} v_i / I \ , \tag{5.1}$$

where $v_i$ is the settling velocity (m $s^{-1}$) of class *i* of soil or sediment when divided into I number of arbitrary classes of equal mass (usually I=50). Depositability was calculated from the data on wet-sieving of soil with the program GUDPRO 3.1 (Griffith University Depositability PROgram) (Lisle *et al.*, 1995). This program requires the wet-density of soil ($\sigma$, kg $m^{-3}$) as an input datum, which was calculated from the relationship between wet density ($\sigma$), and % sand ($s$, > 20 $\mu$m) as shown below (Loch and Rosewell, 1992):

$$\sigma = 1462.1 + 48(1.03259)^{s} \tag{5.2}$$

Wet-densities for soils D, R, and M were 2089, 1558, and 1636 kg $m^{-3}$, respectively. Depositabilities for the uneroded soil were 0.053 (± 0.003), 0.152 (± 0.003), and 0.033 (± 0.003) m $s^{-1}$ for soils D, R, and M, respectively.

**5.3.2 Expected output**

Rainfall re-detachability ($a_d$, kg $m^{-3}$) and detachability ($a$, kg $m^{-3}$) were calculated for all events of treatments 1 and 2 with the expressions (Misra and Rose, 1996):

$$a_d = \frac{c_d \phi}{0.9P} \tag{5.3}$$

and $$a = Q_{eq}\left(\frac{P}{c_d} - \frac{\phi}{a_d}\right)^{-1}, \qquad (5.4)$$

where $Q_{eq}$ ($m^3$ $m^{-2}$ $s^{-1}$) was the runoff rate per unit area at equilibrium, P (m $s^{-1}$) the rainfall rate, and $c_d$ (kg $m^{-3}$) the sediment concentration at equilibrium due to rainfall-driven erosion processes

Parameter β was calculated with the GUEST model using average sediment concentration ($\bar{c}$) and from an estimate of sediment concentration at transport limit due to runoff only ($\overline{c_t}$) for treatments 3a and 3b with the expression (Misra and Rose, 1989):

$$\bar{c} = \bar{c}_t^{\beta}, \qquad (5.5)$$

It should be noted that the notation $\overline{c_t}$ is used here to indicate that first, this estimate of $c_t$ is based on runoff rate averaged over the entire duration of an erosion event and second, it does not include any contribution from rainfall-driven processes. The parameter $J$ was estimated numerically with program GUEST (Misra and Rose, 1989, 1996) by obtaining a solution of the differential equations for mass conservation of sediment. Full expressions of these equations were given by Misra and Rose (1996) for erosion events with or without rilling. Parameter $J$ was estimated for events of treatments 3a and 3b only, where $\Omega > \Omega_0$ and thus, erosion due to entrainment was expected for these treatments. Values of $c_d$ required for the analysis of $J$ were taken from treatment 2, because this erosion treatment had the same KE of rain as for treatments 3a and 3b, thus providing appropriate values of sediment concentration due to rainfall-driven processes under similar conditions of rainfall. When rills formed during erosion, rill characteristics were used in the model.

## *5.4 RESULTS*

### 5.4.1 Soil physical condition at the time of erosion

Mean values of antecedent soil moisture (ASM, % by weight) and volumetric water content ($\theta$, % by volume) measured just before erosion are presented in Table 5.2. ASM was significantly different for various soils and erosion treatments ($p < 0.001$) without any significant interaction. The order of ASM for the studied soils was R > D > M. For the erosion treatments, ASM was in the order treatment 3a > 1 = 2 > 3b. Volumetric water content ($\theta$) was not significantly different between soils (mean $\theta$ for the soils studied = 33.8 %), but it was significantly different for various erosion treatments ($p < 0.001$). The order of $\theta$ for the erosion treatments was the same as for ASM.

Bulk density (BD) measured before erosion was significantly different between soils ($p < 0.001$) but not due to erosion treatments. Values of BD (g $cm^{-3}$) were 1.06, 0.63, and 1.12 for soils D, R, and M, respectively, and they were significantly different from each other. BD followed the opposite order as for ASM.

Mean values of soil strength (penetrometer resistance measured before and after erosion, $P_b$ and $P_a$; and surface shear strength before and after erosion, $\tau_b$ and $\tau_a$) are given in Tables 5.3 and 5 4. Penetrometer resistance after erosion ($P_a$) and surface shear strength before and after erosion ($\tau_b$ and $\tau_a$) had a significant difference among soils ($p < 0.001$), but no difference among erosion treatments ($p > 0.05$), except that $\tau_b$

was significantly lower for treatment 3a than for treatment 3b of soils D and M. Penetrometer resistance measured before erosion ($P_b$) was significantly influenced by soil ($p < 0\ 001$) and erosion treatment ($p < 0.001$) and there was also a significant interaction between soil and treatment. As $P_b$ for various soils and treatments varied widely, the data were not normally distributed. Thus, analysis of variance was performed on log-transformed data and comparison among treatments was based on that analysis.

Table 5.2

Antecedent soil moisture (ASM, % by weight) and volumetric water content (θ, % by volume) of soils and erosion treatments. Numbers followed by same letter(s) within a column are not significantly different ($p \leq 0.05$).

| Soil | ASM (%) |
|---|---|
| D | 34.5 b |
| R | 55.5 c |
| M | 26.2 a |

| Treatment | ASM (%) | θ (%) |
|---|---|---|
| 1. low slope, low KE | 43.3 b | 37.5 b |
| 2. low slope, high KE | 41.2 b | 34.9 b |
| 3a. high slope, high KE, wet | 50.0 bc | 45.6 c |
| 3b. high slope, high KE, dry | 20.5 a | 16.9 a |

For various soils studied, $P_b$, $P_a$, $\tau_b$, and $\tau_a$ were generally in the order M > D ≥ R (Tables 5.3 and 5.4). For the erosion treatments, $P_b$ was generally in the order 3b > 1 = 2 > 3a (Table 5.4). For soil R, erosion treatments had no effect on $P_b$. For soils D and M, $P_b$ was mostly affected by the treatments 3a and 3b (due to a wet/ dry cycle).

Among all the measured parameters of soil strength, $P_b$ was the only parameter significantly influenced by erosion treatments.

Table 5.3

Mean penetrometer resistance after erosion ($P_a$) and surface shear strength before ($\tau_b$) and after ($\tau_a$) erosion. Numbers followed by same letter(s) in each column are not significantly different ($p \leq 0.05$).

| Soil | $P_a$ (kPa) | $\tau_b$ (kPa) | $\tau_a$ (kPa) |
|---|---|---|---|
| D | 151 a | 5.8 a | 5.0 b |
| R | 212 a | 4.0 a | 3.5 a |
| M | 821 b | 10.0 b | 6.8 c |

Table 5.4

Mean penetrometer resistance before erosion ($P_b$). Numbers followed by same letter(s) are not significantly different ($p \leq 0.05$).

| | $P_b$ (kPa) | | |
|---|---|---|---|
| Treatments | Soil D | Soil R | Soil M |
| 1 | 271 b | 241 b | 1327 cd |
| 2 | 390 b | 247 b | 1108 c |
| 3a | 83 a | 226 b | 687 bc |
| 3b | 1335 cd | 224 b | 2536 d |

**5.4.2 Sediment concentration**

Variation of sediment concentration ($c$) with time of runoff/ erosion is presented for various soils and erosion treatments in Fig. 5.1. For any soil, the magnitude of $c$ and its variation with time was smaller for treatments with low slope (treatments 1 and 2) than those with high slope (treatments 3a and 3b). For treatments with high slope, the magnitude of $c$ and its variation at a given time of erosion was mostly high in the beginning of erosion. Assuming an equilibrium in $c$ with time of erosion to be achieved in an event when the variation in $c$ with time is $\leq 10$ % (Section 5.3.1), the data in Fig. 5.1 show that equilibrium in $c$ was reached after about 20 - 25 minutes of erosion for all treatments. The statistical variation in values of $c$ with respect to various treatments and soils are described below on the basis of analysis of variance of the data on $\bar{c}$ and $c_{eq}$ for replicated events.

A comparison of the data in Tables 5.5 and 5.6 indicated values of $c_{eq}$ to be smaller than $\bar{c}$ for all soils and erosion treatments. The difference between $c_{eq}$ and $\bar{c}$ was greatest for soil M and least for soil R. However, the overall effects of erosion treatments on $\bar{c}$ or $c_{eq}$ were similar for each soil. For example, a change of KE of rain (treatments 1 vs 2) or a change in slope (treatments 2 vs 3a) did not greatly affect $\bar{c}$ or $c_{eq}$ of soil R, whereas these treatments greatly influenced $\bar{c}$ or $c_{eq}$ of soils D and M.

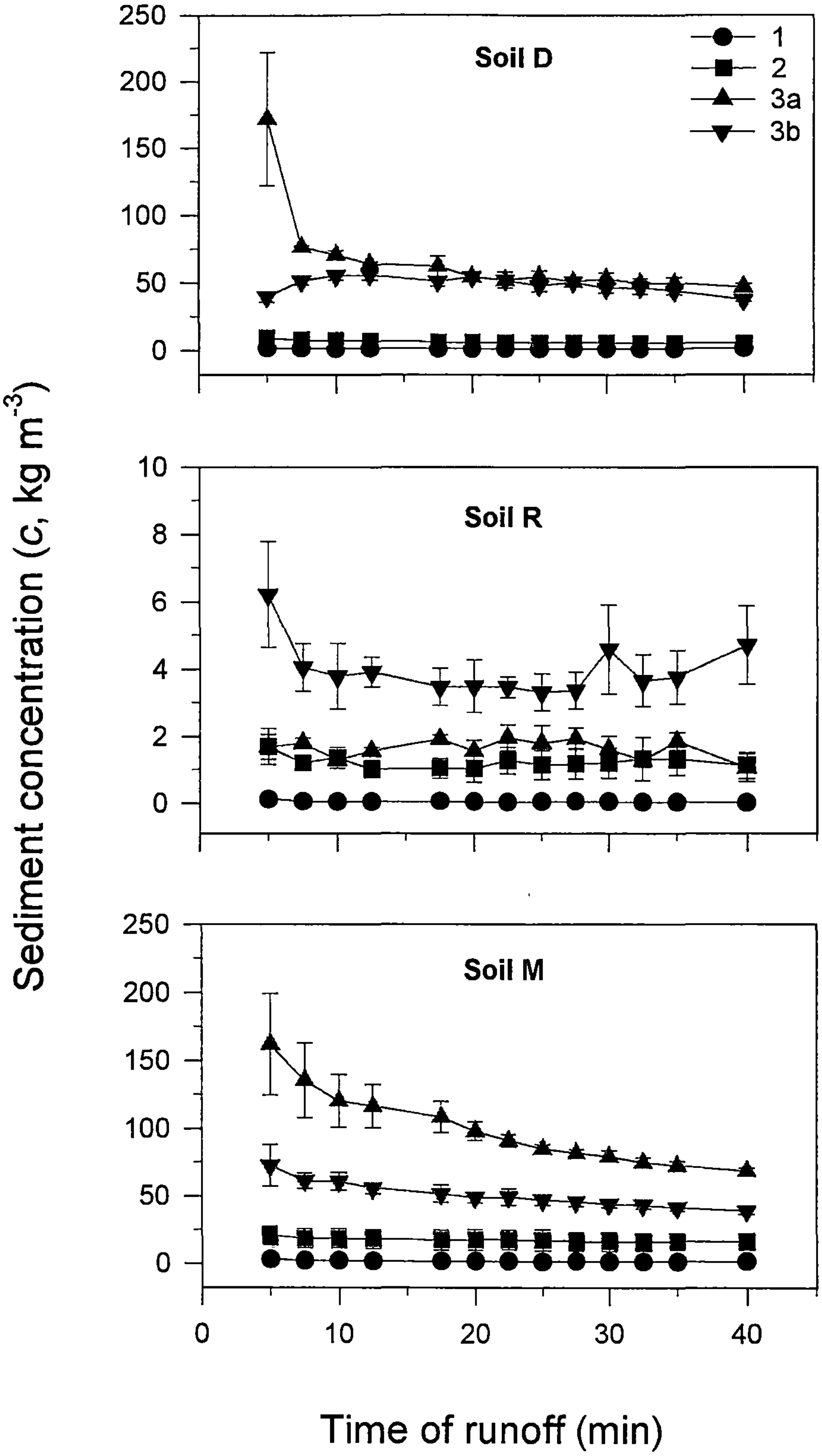


**Fig. 5.1** - Variation of sediment concentration (*c*) with time of runoff during erosion experiments. Vertical bars are SE of mean values (n=3). Note the difference in scale of *c* for soil R compared with the other soils.

Table 5 5

Average sediment concentration ($\bar{c}$, kg $m^{-3}$) for various soils and erosion treatments. Numbers followed by the same letter(s) are not significantly different ($p \leq 0.05$).

| | $\bar{c}$ (kg $m^{-3}$) | | |
|---|---|---|---|
| Treatment | Soil D | Soil R | Soil M |
| 1 | 1.13 b | 0 04 a | 1.63 b |
| 2 | 6.10 c | 1.22 b | 17.42 d |
| 3a | 64 13 ef | 1 63 b | 99.30 f |
| 3b | 48.46 e | 3.98 c | 50.90 e |

Table 5.6

Sediment concentration at equilibrium ($c_{eq}$, kg $m^{-3}$) for various soils and erosion treatments. Numbers followed by the same letter(s) are not significantly different ($p \leq 0$ 05).

| | $c_{eq}$ (kg $m^{-3}$) | | |
|---|---|---|---|
| Treatment | Soil D | Soil R | Soil M |
| 1 | 1.08 a | 0.03 a | 1.34 a |
| 2 | 5.29 a | 1.20 a | 16 35 b |
| 3a | 50.55 c | 1.63 a | 71.86 d |
| 3b | 45.77 c | 3.65 a | 43 98 c |

As $\bar{c}$ for soil R was much smaller than that for the other soils (Fig. 5.1), the data had a skewed distribution. Therefore, analysis of variance for $\bar{c}$ was made on log-transformed data. Skewness did not affect the data of $c_{eq}$. Both $\bar{c}$ and $c_{eq}$ were significantly different for soils and erosion treatments, and there was also a significant interaction between soils and treatments ($p < 0.001$). Mean values of $\bar{c}$ and $c_{eq}$ for various soils and erosion treatments are shown respectively in Tables 5.5 and 5.6. Mean values of $\bar{c}$ were in the order M > D > R. For any soil studied, treatment 1 had the lowest $\bar{c}$ followed by treatment 2 (Table 5.5). There was a significant difference in $\bar{c}$ for treatments 3a and 3b of soil R, but not of soils D and M.

For most erosion treatments, $c_{eq}$ of various soils was in the order M > D > R (Table 5.6). The change in KE of rain (treatments 1 vs. 2) did not affect $c_{eq}$ of soils D and R. A wet/ dry cycle (treatments 3a and 3b) influenced $c_{eq}$ significantly for soil M, but not for soils D and R.

### 5.4.3 Magnitude and variability of erodibility parameters

#### *5.4.3.1 Parameters a and $a_d$*

Estimates of detachability ($a$) and re-detachability ($a_d$) due to rainfall calculated with GUEST (Equations 5.4 and 5.3) are given in Table 5.7. As the terminal velocity of raindrops was not attained in this study, the analysis of $a$ and $a_d$ with equations 5.3 and 5.4 were approximate and are used here mainly for comparison of treatments. Irrespective of soil or erosion treatment, values of $a_d$ were greater than $a$ by two orders of magnitude. With an increase in KE due to rain, values of $a$ and $a_d$ increased for all soils (5 - 36 times). These increases were the highest for soil R and least for soil D. Estimates of $a$ and $a_d$ were lower for soil R than for soils D and M. Between soils D and M, values of $a$ and $a_d$ differed significantly when KE due to rain was high (*i.e.*, treatment 2), but were similar when KE was low (treatment 1).

Table 5 7

Rainfall detachability ($a$, kg m$^{-3}$) and re-detachability ($a_d$, kg m$^{-3}$) of various soils for erosion treatments 1 and 2. Values in parenthesis are standard errors (n=2 for treatment 2 of soil D, and n=3 for other soils and treatments)

| | Treatment 1 | | Treatment 2 | |
|---|---|---|---|---|
| Soil | $a$ (kg m$^{-3}$) | $a_d$ (kg m$^{-3}$) | $a$ (kg m$^{-3}$) | $a_d$ (kg m$^{-3}$) |
| D | 9.9 (3.2) | 2013 (689) | 51.5 (1.6) | 10440 (631) |
| R | 0.29 (0.16) | 191 (104) | 10.5 (2.5) | 6338 (1428) |
| M | 13.7 (1.1) | 1695 (106) | 137.6 (32.3) | 19285 (4946) |

*5.4.3.2 Parameter $\beta$*

In Table 5.8 are presented the estimates of erodibility parameter $\beta$ for various soils and for those erosion treatments where $\Omega > \Omega_0$ (*i.e.*, treatments 3a and 3b). These estimates show that $\beta$ was in the order D > M > R, with very low values of $\beta$ obtained for soil R. The effect of wet/ dry cycle on $\beta$ was significant for soils D and R, but not for soil M

*5.4.3.3 Parameter J*

In Table 5 9 are presented the values of $c_{eq}$ for high slope treatments (*i.e.*, for treatments 3a and 3b where $\Omega > \Omega_0$) and the corresponding estimate of $J$ for soils D and M. Parameter $J$ could not be estimated for soil R because there was no significant difference in $c_{eq}$ between treatments 2 and 3 (Table 5.6) It should be noted that $J$ can be estimated with GUEST only when runoff entrainment is effective. An indication of effective entrainment is when $c_{eq}$ increases with an increase in stream power (from $\Omega \leq \Omega_0$ to $\Omega > \Omega_0$). Values of J were greater for soil M than soil D.

Table 5.8

Mean values of erodibility parameter β for various soils and erosion treatments . Standard error of β (in parentheses) were based on three replicated events for all except soil D of treatment 3a, which had data for two events only.

| | Treatment 3a | | Treatment 3b | |
|---|---|---|---|---|
| Soil | β | | β | |
| D | 0.95 | (0.017) | 0.91 | (0.016) |
| R | 0.13 | (0 022) | 0 39 | (0.055) |
| M | 0.73 | (0.068) | 0.76 | (0.018) |

Table 5.9

Sediment concentration at equilibrium ($c_{eq}$) and erodibility parameter $J$ estimated for treatments 3a and 3b of soils D and M. Values in parenthesis are standard errors of the mean values (n=3).

| Soil - treatment | $c_{eq}$ (kg m$^{-3}$) | $J$ (J kg$^{-1}$) |
|---|---|---|
| D - 3a | 50.6 (3.5) | 2 0 (0 4) |
| D - 3b | 45.8 (2.4) | 1.9 (0.3) |
| M - 3a | 71.9 (2.8) | 9.7 (6.4) |
| M - 3b | 44.0 (3.1) | 33.3 (10.4) [a] |

[a] Mean and SE are based on 2 replicated events. $J$ could not be estimated for the third event.

## *5.5 DISCUSSION*

### 5.5.1 Sediment concentration and soil strength

In this study, soil physical condition at the time of erosion characterised by bulk density (BD), water content (ASM and $\theta$) and strength of surface soil ($P_b$ and $\tau_b$) varied significantly with soil type (Tables 5.2, 5.3 and 5.4). Although these measurements were taken after the soils were wetted to apparent saturation, marked differences in ASM, $\theta$, $P_b$ and $\tau_b$ were observed for erosion treatments with and without a drying cycle (treatments 3b and 3a, respectively). Values of ASM and $\theta$ were lower for treatment 3b than 3a because water taken up by the soil during wetting was less in 3b than 3a. Therefore, in treatment 3b water was insufficient to saturate the entire depth of soil in the erosion trays. This indicates that drying of a soil after erosion may have influenced its surface structure such that the surface soil reached saturation rapidly during wetting. This observation supports the importance of the amount and rate of water uptake during wetting on erosion as indicated in studies of Govers and Loch (1993).

Variation in physical conditions of the surface soil could be also attributed to the structural sensitivity of these soils contributing to surface sealing and/or crusting affecting infiltration and runoff (Le Bissonnais *et al.*, 1993; Bresson and Moran, 1995). Soil strength increased significantly after the drying cycle (for example, $P_b$ in Table 5.4) for soils D and M but not for soil R as a result of the differences in these soils' susceptibility to wetting. The index of susceptibility to wetting ($S_W$) was 0.28, 0.77 and 0.05 on a scale of 0-1 for soils D, M and R, respectively (Table 4.4, Chapter 4).

An increase in slope and KE of rain increased sediment concentration (Tables 5.5 and 5.6). However, at high slope, significant difference in sediment concentration with wet/dry cycle was most pronounced for soil M followed by soil D. For soil D, wet/dry cycle mainly affected the temporal variation of $c$ in the first few minutes of erosion (Fig. 5 1). There was some indication that changes in soil strength arising from wet/dry cycles influenced the sediment concentration (Fig. 5.2). Consideration of the combined data for treatments 3a and 3b indicated average sediment concentration ($\bar{c}$) for each soil to decrease exponentially with an increase in $P_b$ (Fig. 5.2a) and linearly with an increase in $\tau_b$ (Fig. 5.2b). Such effects of soil strength on sediment concentration are consistent with reports from other experimental studies and current theory of the role of soil strength in erosion (Al-Durrah and Bradford, 1982; Bradford *et al.*, 1986; Bradford and Huang, 1995; Sharma *et al.*, 1995, Hanson, 1996; Rose *et al.*, 1990; Misra and Rose, 1995).

**5.5.2 Erodibility parameters and soil strength**

The magnitude for rainfall-driven erodibility parameters ($a$ and $a_d$, estimated for low slope treatments 1 and 2) were similar for soils D and M, but low for soil R (Table 5.7) when compared with values reported for other soils by Proffitt *et al.* (1991) and Misra and Rose (1995). There was little opportunity to examine the effects of soil strength on these erodibility parameters because for erosion treatments 1 and 2, soil strength (e.g. $P_b$) was low and the variation in soil strength was small (Table 5.4). It was not possible to combine data for both treatments because $a$ and $a_d$ were influenced by variation in KE as well as soil strength.

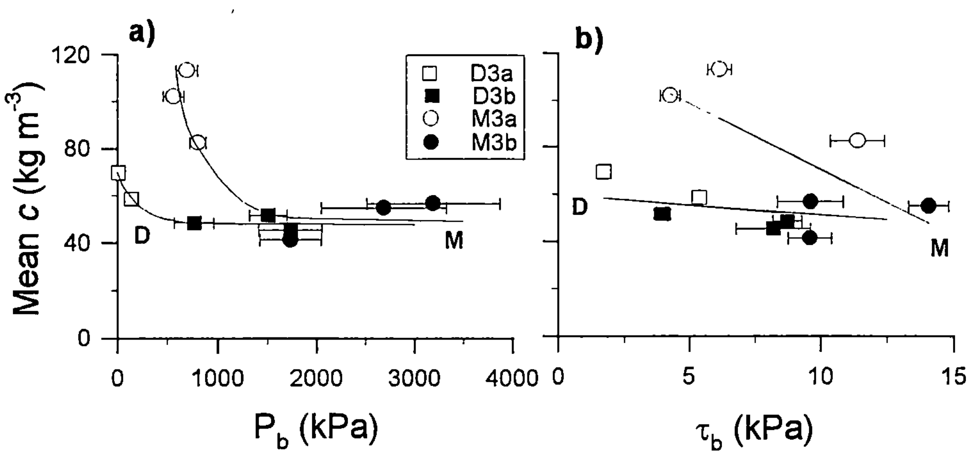


**Fig. 5.2** - Variation of sediment concentration ($c$) with a) penetrometer resistance before erosion (Pb), and b) Torvane shear strength before erosion ($\tau_b$), for treatments 3a and 3b of soils D and M.

The erodibility parameter, $\beta$ was the lowest for soil R, and its value increased considerably after the drying cycle for this soil (Table 5.8) without any significant change in soil strength (Tables 5.3 and 5.4). For soil D, $\beta$ remained close to 0.9 for the treatments 3a and 3b, which indicates that the sediment concentration ($c$) for this soil was closer to the transport limit than the other soils. Soil M, which had shown a large increase in soil strength and a decrease in $c$ with drying (Tables 5.4-5.6), had an intermediate value of $\beta$, that was not affected significantly by drying. The detailed variation in $\beta$ with $P_b$ and $\tau_b$ of soils D and M is shown in Fig. 5.3. For both soils, no definitive dependence of $\beta$ on $P_b$ was evident. The variation in $\beta$ with $\tau_b$ was also different for the two soils. Similar uncertainty in relating erodibility parameters with soil strength could be also seen when the erodibility parameter $J$ was plotted against soil strength for both soils (Fig. 5.4). There was little improvement in the dependence of β or $J$ on soil strength (data not shown) when β and $J$ were plotted against soil strength after erosion ($P_a$ and $\tau_a$)

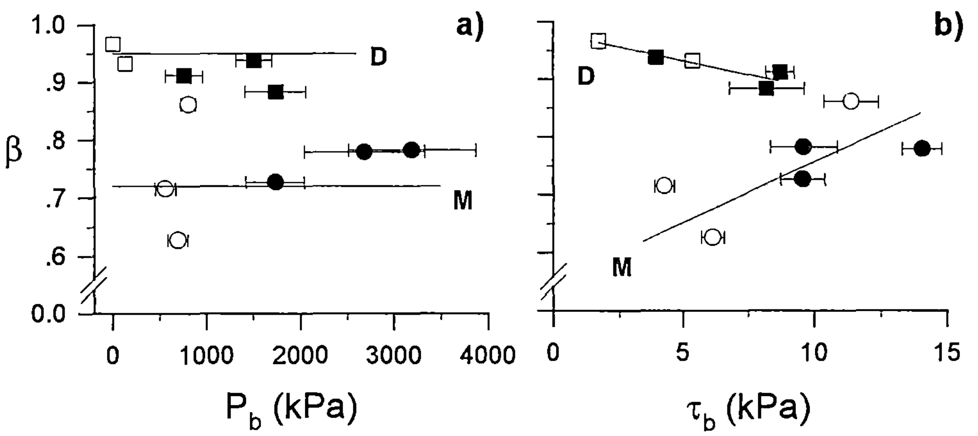


**Fig. 5.3** - Variation of β with a) penetrometer resistance before erosion ($P_b$), and b) Torvane shear strength before erosion ($\tau_b$), for treatments 3a and 3b of soils D and M. Same legend as for Fig. 5.2.

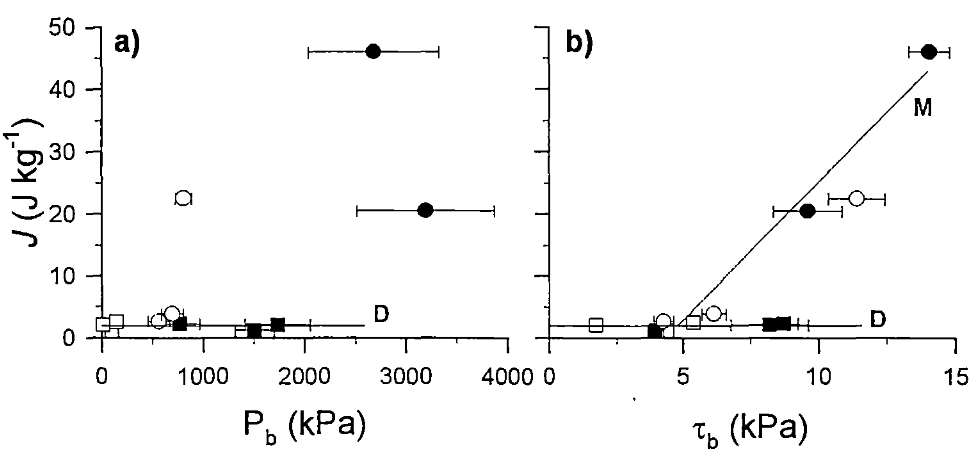


**Fig. 5.4** - Variation of $J$ with a) penetrometer resistance before erosion ($P_b$), and b) Torvane shear strength before erosion ($\tau_b$), for treatments 3a and 3b of soils D and M. Same legend as for Fig. 5.2.

It is expected from the theory used in the erosion model GUEST (Misra and Rose, 1989; Rose et al., 1990; Hairsine and Rose, 1992a, 1992b; Rose, 1993; Misra and Rose, 1995, 1996) that when there is an increase in cohesive strength of soil, erodibility parameter $\beta$ will decrease and $J$ will increase. These expectations were met reasonably well in a previous study (Misra and Rose, 1995), where the variation in these erodibility parameters with variation in soil strength could be explained for a soil with and without compaction. However, these authors pointed out that uncertainty in relating erodibility parameters and soil strength may arise due to practical difficulties in taking into account the effects of soil structure on cohesive strength. In this study, the variation in sediment concentration with variation in soil strength was consistent with the theoretical expectation of the model GUEST and other erosion models (Nearing *et al.*, 1989); but the variation in erodibility parameters $\beta$ and $J$ with soil strength (penetrometer resistance and shear strength) was uncertain. The reasons for these uncertainties are elaborated below in light of the mechanisms which explain the variation in soil strength with time and that due to drying and rewetting of a saturated soil.

### 5.5.3 Possible effects of aging and drying of soil on erodibility and strength

The variation in erosion and erodibility with and without a dry period illustrated in this study may have an impact on the prospect of long term prediction of erosion. In field studies, variation in soil erosion and $\beta$ with time can be explained reasonably well when there is a change in cropping or tillage (Paningbatan *et al.*, 1995; Presbitero *et al.*, 1995, Hashim *et al.*; 1995, Ciesiolka *et al.*, 1995). Other studies have shown that erodibility of a wet soil could change with time because of a change in soil cohesion arising from the development of inter- and intra-particle bonds

(Shainberg *et al.*, 1996). However, these authors found the effects of aging on erodibility of soils to be dependent on soil type. The lack of a unique relationship between erodibility and soil strength for the soils D and M observed in this study (Figs. 5.3 and 5.4) could be because of a difference in mechanisms by which soil strength increased in these soils after drying.

It is well known that soil strength can change with time from disturbance (which occurs due to tillage or passage of vehicles) with or without a change in bulk density and water content (Dexter, 1991). Attempts have been made (Nearing and West, 1988) to combine the concepts of variation in soil strength with time following the disturbance of surface soil due to rain/erosion with that due to drying, to interpret possible variation in soil erodibility with time (Nearing *et al.*, 1988). There are two mechanisms which explain increase in soil strength with time (Dexter *et al.*, 1988): one is due to the rearrangement of clay particles (thixotropy effect) affecting pore-size distribution and matric potential in soil, and the other is due to the formation of cementing bonds between soil particles, even in the absence of organic matter (Utomo and Dexter, 1981; Kemper and Rosenau, 1984). To explore the possible application of these mechanisms, a dimensionless packing-density ($D_p$) was estimated for soils D and M from the measured values of BD, ASM (%), and OM (organic matter in %, calculated as 1.72 $\times$ organic carbon from Table 3.1) to indicate the proportion of volume occupied by the soil minerals. Following the method of Utomo and Dexter (1981):

$$D_p = \frac{BD}{PD}\left[\frac{100 - OM}{100 + ASM}\right]. \tag{5.6}$$

For both soils, the particle density (PD) was taken as 2.7 Mg $m^{-3}$ in the above calculation. Values of $D_p$ in Table 5.10 indicated that the increase in strength ($P_b$ and $\tau_b$) after drying of soil D may have been due to a dense packing following the rearrangement of particles. As there was no change in packing density for the soil M, it is stipulated that the increase in strength in this soil may have been due to the formation of cementing bonds between particles. There was some indication that soil strength ($P_b$ and $\tau_b$) was affected by the degree of saturation in soil (*i.e.*, $S_d$, the per cent of pore space occupied by water, given in Table 5.10) when there was a change in packing density (i.e. for soil D) but not as well when packing density did not vary (as for soil M) (Fig. 5.5).

Table 5.10

Variation in strength-related soil properties with and without allowing the soil to dry following erosion. Data are shown for two soils (D and M) and erosion treatments 3a (without drying) and 3b (after drying). Values in parentheses are standard errors of measurements or estimates for replicated erosion events.

| | Soil D | | Soil M | |
|---|---|---|---|---|
| Soil properties | 3a | 3b | 3a | 3b |
| $P_b$ (kPa) | 83 (39) | 1335 (292) | 687 (70) | 2536 (428) |
| $\tau_b$ (kPa) | 3.84 (1.09) | 6 96 (1.51) | 7.27 (2 14) | 11.09 (1.49) |
| $S_d$ (%) | 86.7 (6.6) | 25.9 (11.6) | 74.0 (11.8) | 23.6 (5.0) |
| $D_p$ | 0.266 (0.007) | 0.332 (0.015) | 0.313 (0.018) | 0.326 (0.009) |

As soil strength can vary due to a change in packing density and/or due to a change in the number of inter-particle bonds, the effects of these changes on soil cohesion (a component of shear strength) is uncertain. Therefore, it may be possible to infer soil erodibility from soil strength reasonably well only when there is some certainty that the measurement of soil strength reflects the cohesive strength of soil.

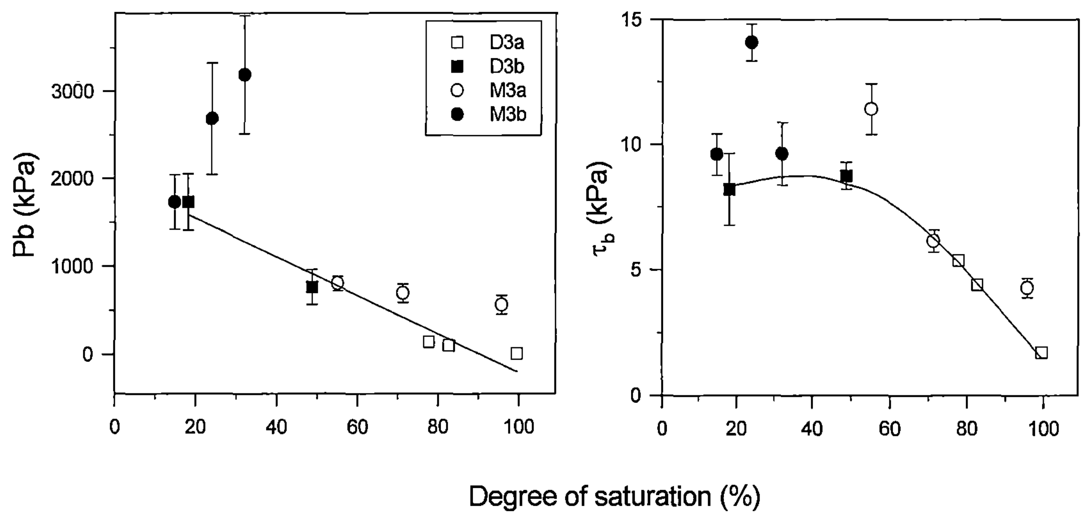


**Fig. 5.5** - Variation of soil strength ($P_b$ and $\tau_b$) before erosion with the degree of saturation ($S_d$) in soil, for soils D and M. Possible relationship between soil strength and $S_d$ is illustrated for soil D with a solid line/ curve.

## *5.6 CONCLUDING REMARKS*

The results of this study illustrate that simple measures of soil strength are able to explain differences in erosion (sediment concentration). However, soil strength alone, without consideration of additional soil properties (e.g. the effects of soil structure on soil strength), may not be sufficient to allow prediction of erodibility parameters for soils. It may be possible to overcome the difficulties in relating erodibility with soil strength if a consistent approach can be developed to explain the variation in soil strength with time and drying and/or with development of techniques for measurement and interpretation of soil cohesion directly.

Reports from recent, *in-situ* erosion studies (Paningbatan *et al.*, 1995, Presbitero *et al.*, 1995; Hashim *et al.*; 1995; Ciesiolka *et al.*, 1995) are able to interpret changes in erosion and erodibility parameter (β) with changes in cultural practices in farms (cropping, rotation and tillage). The variation in erosion and erodibility with and without a dry period outlined here not only illustrates the possible variation in strength and structure of soil between consecutive erosion events, but also indicates the potential difficulties with the long-term prediction of erosion.

# Chapter 6

## THE DYNAMICS OF AGGREGATE BREAKDOWN UNDER DISRUPTIVE FORCES APPLIED BY AN ULTRASONIC PROBE

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### *6.1 INTRODUCTION*

When a structured soil is subjected to external stresses, failure of soil occurs if the applied stresses exceed the strength of the bonds holding particles and aggregates together (Dexter, 1988; So *et al.*, 1996). The strength of the bonds between soil particles and aggregates is one of the fundamental soil properties that affects aggregate stability. To evaluate aggregate stability, some disruptive force is usually applied to soil aggregates. As a result, the aggregates break down into various sizes depending on the hierarchical stability of aggregates. The disruptive forces from natural phenomena and from human interference can also break the bonds within and between soil aggregates. Such forces arise from the impact of raindrops on soil, the impact of wetting and drying (slaking, dispersion, crusting), and forces arising from mechanical action (eg. tillage, cultivation and traffic).

During erosion, the soil surface is subjected to the impact of raindrops and the soil may also be subjected to wetting and drying. As a result, aggregate breakdown may occur, depending on the structural stability of soils and the stresses arising due to erosion. Soil erodibility is influenced by the soil's structural response to erosion. Aggregate stability can provide a measure of soil erodibility (Bryan, 1976; Pauwels *et al.*, 1976; Luk, 1979; Amezketa *et al.*, 1996), because stability of soil aggregates is influenced by soil structure and by the disruptive forces applied during tests designed for the measurement of aggregate stability. A suitable method to measure aggregate stability may provide a simple and inexpensive way of assessing soil erodibility without the requirements of erosion studies.

There is no general agreement on the best method to measure aggregate stability. The more commonly used method is wet-sieving (Yoder, 1936; Nijhawan and Olmstead, 1947, Kemper and Koch, 1966, Le Bissonnais, 1995). An alternative method is 'waterdrop impact' (Al-Durrah and Bradford, 1981; Huang *et al.*, 1982), which allows measurement of the force and the resultant aggregate breakdown during the impact of single waterdrops to soil aggregates. The main differences between these methods is the nature, intensity and distribution of disruptive forces, the pre-treatment of soil samples, and the size of aggregates used. The force applied to soil during wet-sieving is difficult to measure, but this method has the advantage of allowing the use of whole soil samples. In studies of waterdrop-impact, the disruptive forces can be measured, but application of this method is restricted to aggregates of limited size. Fuller and Goh (1992) suggested that the quantification of aggregate stability could be improved by controlling the energy of the disruptive forces applied to soil, and that ultrasonic energy would be suitable for this purpose. The application of ultrasonic energy for the measurement of aggregate stability enables a wide range of aggregate-sizes (whole soil samples) to be used, and also an estimate of the disruptive energy applied.

Ultrasonics is the study and application of sound propagated at frequencies beyond the range audible to human ear (> 18 MHz) (Blitz, 1967; Blitz and Simpson, 1996). The generation and propagation of ultrasonic sound is the same as for audible sound. Sound (and ultrasound) propagates by waves. The propagation of ultrasound is affected by the molecular structure of the medium in which it propagates. In water the wavelength of ultrasound is $1.5 \times 10^{-2}$ to $1.5 \times 10^{-4}$ cm. Liquids subjected to high-frequency ultrasound ($10^7$ - $10^9$ Hz; Blitz, 1967) experience rapidly alternating low- and high-pressure; this is accompanied by a number of effects that can only be described by the laws of nonlinear acoustics (Prokhorov, 1981). Among the nonlinear acoustic effects, cavitation is perhaps the most important phenomenon that occurs. Cavitation takes place in a liquid subjected to ultrasound when gas bubbles (existing or being formed) in the liquid increase in size and pulsate with the ultrasonic frequency. When the gas bubbles collapse (due to differences in pressure inside and outside the bubbles), high local pressures of the order of thousands of atmospheres arise releasing energy at a tremendous rate and producing shock waves (Blitz, 1967; Prokhorov, 1981). These effects have been exploited in the production of emulsions, emulsification of immiscible liquids, to clean surfaces, and disruption of physical and biological structures. In soil-water suspensions, ultrasonic waves push the particles in the direction of wave propagation and the suspended soil aggregates can be broken up by vibration and/ or by cavitation.

Ultrasound has been used as a source of energy to disrupt aggregates in soil-water suspensions for various reasons, particularly for dispersing soil before particle-size analysis (Watson, 1971; Kubota, 1972; Mikhail and Briner, 1978; Busacca *et al*, 1984; Gee and Bauder, 1986). Gee and Bauder (1986) and Gregorich *et al.* (1988)

reported that during dispersion of soil with the use of ultrasound, primary particles of the soil are not destroyed With this method soil dispersion may be achieved without any pre-treatment of soil. Sonification can: enhance dispersion of aggregates (that is usually achieved with addition of dispersing agents), remove binding agents (such as organic matter, iron oxide, and carbonates) and remove flocculating agents (such as soluble salts) without using chemicals. Sonification can also disperse the soil samples without contamination from the ultrasonic probe. Contamination of the soil samples due to abrasion of the probe tip can be avoided by using a cup horn device. This device replaces the ultrasonic probe tip during sonification, and enables the soil sample to be isolated from the ultrasonic source (Busacca *et al.*, 1984). Therefore, soil dispersion by sonification is suitable to study the distribution and nature of organic matter associated with various-size fractions of soil (Shaymukhametov, 1974; Christensen, 1985; Gregorich *et al.*, 1988, 1989; Fuller and Goh,. 1992; Eriksen *et al.*, 1995). The method of ultrasonic dispersion of soil is also suitable for mineralogical analysis (Gee and Bauder, 1986). The application of ultrasonic energy to disperse soil has been extended to measurement of aggregate stability of soils and to infer mechanisms of aggregation (North, 1976, 1979; Moen and Richardson, 1984; Fuller and Goh, 1992; Raine and So, 1993).

When ultrasonic energy is used to measure aggregate stability, only a fraction of the energy is used to disperse the soil aggregates. North (1976) developed a method to estimate the energy component of ultrasound that produced soil dispersion. Fuller and Goh (1992) adapted North's method to estimated the energy applied to soil-water suspensions during sonification, and examined the relationship between applied energy and aggregate stability of soils. Raine and So (1993, 1994) perfected the method of North (1976) to measure the bonding energy of soil aggregates in a soil-water suspension.

The objectives of the present work were: (1) to test the adequacy of ultrasonification method in determining the energy required for aggregate breakdown for soils of contrasting structure, and (2) to study the dynamics of aggregate breakdown in soil-water suspensions exposed to variable duration of ultrasound energy

## *6.2 METHODS AND DATA ANALYSIS*

### 6.2.1 Methods

A full description of the method undertaken to determine the energy used in soil dispersion was presented in Section 2.5. Briefly, soil-water suspensions were prepared for three forest soils (soils D, R, and M, which were previously used in erosion experiments). Each suspension was exposed to different duration of ultrasonic energy by inserting an ultrasonic probe into the suspension. The size-distribution of the soil aggregates in the suspension following ultrasonic disruption was measured with the pipette method of particle-size analysis (Gee and Bauder, 1986) for the size-fractions < 53 μm, and by sieving for the fractions > 53 μm.

### 6.2.2 Analysis of the energy components of the system

The theory used for the estimation of the energy applied by an ultrasonic probe to a soil-water suspension was presented in Section 2.5.2 (Chapter 2, Materials and Methods). The equation for energy balance in a soil-water suspension subjected to ultrasound (equation 2.3) is the accumulated energy ($E_i$, J) for a given duration ($\Delta t$, s), expressed as:

$$\frac{E_i}{\Delta t} = \frac{1}{\Delta t}\left(E_h + E_c + E_t + L_g\right), \tag{6.1}$$

where $E_h/\Delta t$ is the power (in W or J $s^{-1}$) used in heating the suspension, $E_c/\Delta t$ the power lost by conduction, $E_t/\Delta t$ the power lost by transmission of wave energy, and $L_g/\Delta t$ the power used in disaggregation and dispersion of soil aggregates For a suspension containing soil which had been dispersed completely, $L_g = 0$; therefore,

$$\frac{E_i}{\Delta t} = \frac{1}{\Delta t}\left(E_h' + E_c' + E_t'\right), \tag{6.2}$$

where $E_h'$, $E_c'$, and $E_t'$ correspond to a suspension of completely dispersed soil. Note that a fully-dispersed or completely dispersed soil is referred to here as a soil which was sonified for 15 minutes.

The energy applied in the soil-water suspension ($E_i$) cannot be obtained accurately from the readings of the ultrasonic probe (North, 1976; Christensen, 1992). Therefore, each of the energy components of equations (6.1) and (6.2) was obtained experimentally. Energy components $E_t$ and $E_t'$ were assumed to be $\approx 0$ due to the experimental conditions of the study (as stated in Section 2.5.2) When an unknown but fixed amount of energy $E_i$ is applied for a given duration ($\Delta t$) to a soil-water suspension containing completely dispersed soil and an undispersed soil, equations (6.1) and (6.2) are numerically equal. Therefore, estimation of the components $E_h$, $E_c$, $E_h'$, and $E_c'$ allows the component $L_g$ of equation (6.1) to be calculated.

During sonification of a soil-water suspension, most of the energy is used in heating the suspension. The heat energy ($E_h$ and $E_h'$) was calculated from equation (presented with more detail in Chapter 2 as equation 2.5):

$$E_k = k\Delta T_k, \tag{6.3}$$

where $E_k$ was $E_h$ or $E_h'$, and $\Delta T_k$ was $\Delta T_h$ or $\Delta T_h'$, *i.e.*, the change in temperature with time for a given duration of sonification and for various durations of sonification, and $k$ was a constant that depended on the mass and specific heat of soil, water, and the container used to hold the soil-water suspension. Mean values of $k$ were 252.8, 254.3, and 252.4 J °C$^{-1}$ for soils D, R, and M, respectively. The changes in temperature with time, $\Delta T_h$ and $\Delta T_h'$, were measured during sonification of the soil-water suspension with undispersed and fully-dispersed soil, respectively. In Fig. 6.1 is presented the variation in temperature (T) of soil-water suspensions (with fully dispersed soil) for increasing duration of sonification ($\Delta t$). Similar linear relationships between T and $\Delta t$ were also obtained for soil-water suspensions with undispersed soil. In Table 6.1 are presented the parameters of the linear regressions of Fig. 6.1 (using fully-dispersed soil) and for the same relationship using undispersed soil. The parameters presented in Table 6.1 were used to calculate $\Delta T_h$ and $\Delta T_h'$ for any duration of sonification.

It can be seen from Fig. 6.1 and Table 6.1 that the linear relationship between T and $\Delta t$ had a similar slope for all soils, in either dispersed or undispersed condition (average slope of regression was 0.12 °C s$^{-1}$, $r^2 = 0.99$, $p < 0.001$).

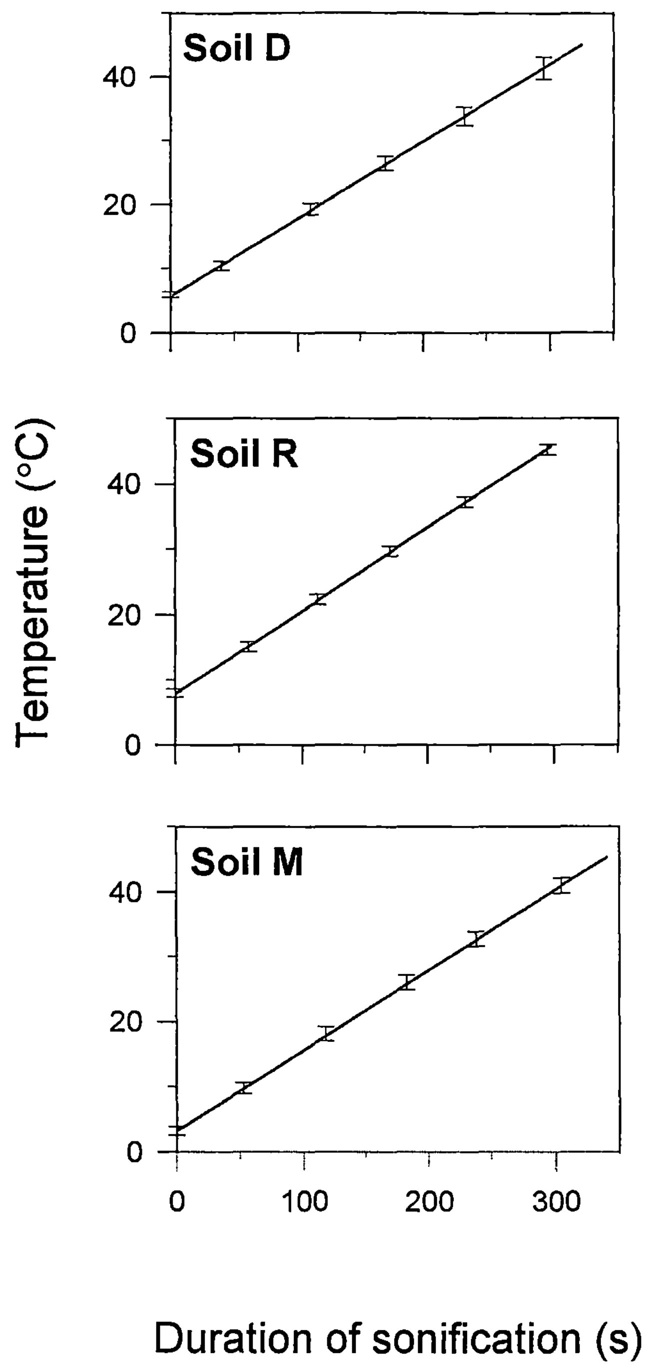


**Fig. 6.1** - Variation of temperature (T) with duration of sonification (Δt) for soils D, R, and M. The lines are regression lines and the bars indicate standard errors of n=4 for soil D, and n=3 for soils R and M. Some data points are shown for clarity, although temperature measurements were made at 1 second intervals. The error bars shown represent full range of the errors.

Table 6.1

Parameters ($a_t$ and $b$) of linear regression $T = a_t\Delta t + b$, for the change in temperature (T, °C) of the soil-water suspension with duration of sonification ($\Delta t$, s) for various undispersed and fully-dispersed soils. Numbers in parentheses are standard errors of the regression parameters. Coefficient of variation ($r^2$) for regression equations is also shown.

| Soil | Undispersed soil | | | Fully dispersed soil | | |
|---|---|---|---|---|---|---|
| | $a_t$ | $b$ | $r^2$ | $a_t$ | $b$ | $r^2$ |
| D | 0.12 ($1.5 \times 10^{-4}$) | 6.0 (0.04) | 0.99 | 0.12 ($1.4 \times 10^{-4}$) | 5.7 (0.03) | 0.99 |
| R | 0.12 ($2.2 \times 10^{-4}$) | 8.4 (0.06) | 0.99 | 0.12 ($1.3 \times 10^{-4}$) | 7.8 (0.02) | 0.99 |
| M | 0.12 ($1.6 \times 10^{-4}$) | 3.6 (0.04) | 0.99 | 0.12 ($1.8 \times 10^{-4}$) | 3.3 (0.04) | 0.99 |

During sonification, some energy is lost from the suspension by conduction. The energy lost by conduction ($E_c$ and $E_c'$) was calculated from equation (6.3), where $E_k$, was $E_c$ or $E_c'$ and $\Delta T_k$ represented the change in temperature with time ($\Delta T_c$ and $\Delta T_c'$) when the suspension was allowed to cool down after sonification. To obtain the values of $\Delta T_c$ and $\Delta T_c'$, the rate of cooling of the suspension was calculated from a Heat Loss Characteristic Curve (HLCC) of the system, which can be described by an equation of exponential decay of temperature with time to a non-zero asymptote:

$$T = T_0 e^{(-r\Delta t)} + T_1 \quad , \qquad (6.4)$$

where T (°C) is the temperature of the soil-water suspension at any duration $\Delta t$ (s), $T_1$ (°C) is the asymptote of the curve, $T_0+T_1$ is the temperature of the suspension at time t=0, and $r$ ($s^{-1}$) is the rate constant of heat loss. For a given duration of sonification

($\Delta t$), $\Delta T_c$ or $\Delta T_c'$ was calculated from the difference of the initial temperature $T_i$ (*i.e.*, $T_0+T_1$ of equation 6 4), and the final temperature ($T_f$) at time t (obtained from equation 6.4). The HLCC was obtained for the soil-water suspension of each soil containing either fully dispersed soil (to calculate $E_c'$) or undispersed soil (for $E_c$).

In Fig. 6 2 is presented the HLCC for fully-dispersed soils. The curves of HLCC for undispersed soils were not significantly different from those of Fig. 6.2, and are not shown. The rate constants (*r*) of heat loss obtained from the HLCC are presented in Table 6.2 for undispersed and fully-dispersed soils. Estimation of all $\Delta T_c$ and $\Delta T_c'$ for any duration of sonification (explained in the preceding paragraph) were based on equation (6.4), with parameter *r* as shown in Table 6 2. The values of *r* (Table 6.2) for a corresponding duration of sonification ($\Delta t$) were very small when compared with the rate of temperature increase for that duration of sonification (slope of the regression, $a_t$, in Table 6.1). The large difference between $a_t$ and *r* suggested that there was a negligible loss of energy from the suspension due to cooling during sonification. Thus, $E_c$ and $E_c'$ were taken to be $\approx 0$ for various duration of sonification used in the experiments.

With $E_t$, $E_t'$, $E_c$, and $E_c'$ being zero, equations (6.1) and (6.2) are reduced to: $E_i/\Delta t = E_h'/\Delta t = \left(E_h + L_g\right)/\Delta t$. Thus, for a given duration of sonification ($\Delta t$), the total energy applied to the suspension ($E_i$) was taken from the estimated value of $E_h'$ (*i.e.*, for a fully-dispersed suspension) for that duration Once $E_i$ was known, $L_g$ was calculated using the estimated value of $E_h$ for an undispersed suspension as $L_g = E_i - E_h$.

Table 6.2

The rate constant of heat loss ($r$, $s^{-1}$) of equation (6.4) for soil-water suspensions containing various undispersed and fully-dispersed soils. Values in parentheses are standard errors of the parameter $r$. Coefficients of variation of the regressions ($r^2$) are given, and they were all significant ($p < 0.001$).

| Soils | Undispersed soil | | Fully dispersed soil | |
|---|---|---|---|---|
| | $r$ | $r^2$ | $r$ | $r^2$ |
| D | $4.6\times10^{-4}$ ($1.4\times10^{-6}$) | 0.74 | $4.6\times10^{-4}$ ($1.6\times10^{-6}$) | 0.76 |
| R | $4.0\times10^{-4}$ ($1.9\times10^{-6}$) | 0.89 | $4.0\times10^{-4}$ ($0.9\times10^{-6}$) | 0.85 |
| M | $3.3\times10^{-4}$ ($1.3\times10^{-6}$) | 0.89 | $3.3\times10^{-4}$ ($1.2\times10^{-6}$) | 0.87 |

Values of accumulated $E_i$ for a given duration ($\Delta t$) were estimated from corresponding values of $E_h'$. From equation (6.3) and values of the slope parameter ($a_t$ in Table 6.1), $E_h'/\Delta t = ka_t$. Thus, the mean power applied to soil-water suspensions was 30.5 W (SE = 0.14, n = 6 for sonifications with fully-dispersed and undispersed soil of three soils). Therefore, the energy ($E_i$) applied during sonifications could be estimated for any duration of application of ultrasonic energy by multiplying the duration of sonification (in seconds) by the power applied.

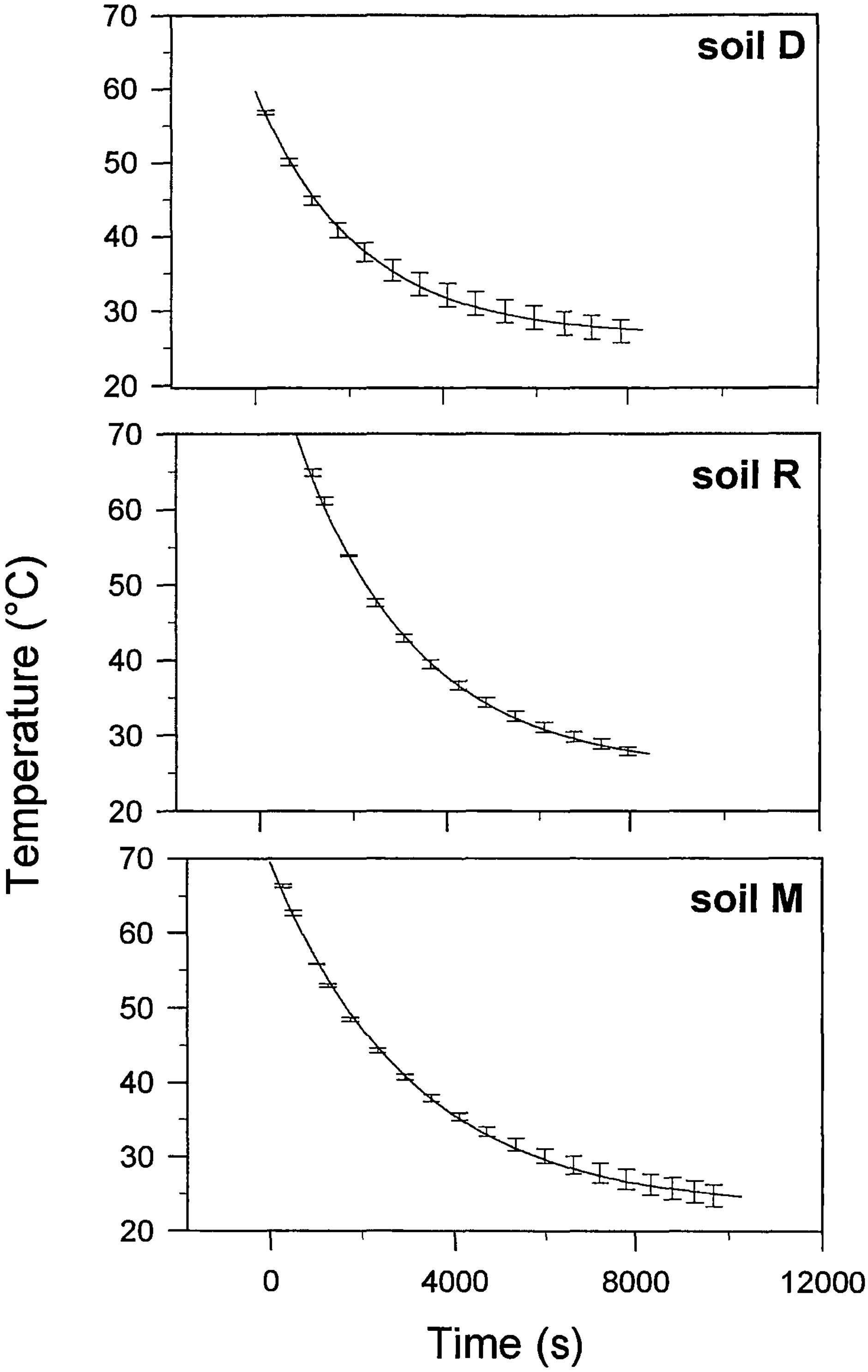


**Fig. 6.2** - Heat loss characteristic curves (HLCC) for fully-dispersed soil. Some data points with error bars are shown for clarity, although measurements were made at 5 second intervals. Error bars are based on n=4 for soil D, and n=3 for soils R and M.

## *6.3 RESULTS AND DISCUSSION*

### 6.3.1 Adequacy of ultrasonic method for estimation of the energy used in breaking down soil aggregates ($L_g$)

The rate of temperature increase with time was not different for undispersed and fully-dispersed soil (*i.e.*, parameter $a_t$ in Table 6.1). This made it impossible to estimate the energy used in breaking down the soil aggregates, $L_g$ (*i.e.*, $L_g = E_i - E_h$). Furthermore, the magnitude of the standard errors (SE) of $E_h$ and $E_i$ for the soils studied in this work was = 4-18, 2-48, and 7-66 J $g^{-1}$, for soils D, R and M, respectively. The values of SE (in J $g^{-1}$) were of similar magnitude as the mean values of $L_g$ (~30 J $g^{-1}$) reported for Vertisols (Raine and So, 1993).

The failure of the ultrasonic method in estimation of $L_g$ may have been due to the high sensitivity of the method to small errors in sampling and weighing of soil and water, possible errors arising from the extent of degassing of the water used to make up suspensions, and error in positioning of the ultrasonic probe in the soil-water suspension affecting the efficiency of sonification (described earlier in Chapter 2). This method could be improved if the power output of the probe is known from an alternative measurement and also by using soil-water suspension of mechanically and chemically dispersed soil as a fully-dispersed soil to estimate $E_h'$ and other energy components of the system.

### 6.3.2 Soil dispersion by immersion-wetting

The distribution of soil into various size-classes as a result of immersion-wetting ($E_t$ = 0 J $g^{-1}$) is given in Table 6.3. The susceptibility of soils to disperse spontaneously into fine material (< 20 μm) when immersion-wetted can be calculated as the proportion of soil dispersed (given in Table 6.3) in relation to the maximum possible dispersion (shown in Table 6.4 as PSA). Dispersion of < 20 μm due to immersion-wetting was 3.5% of the maximum dispersion for soil D, 0% for soil R, and 31% for soil M. Therefore, the susceptibility of soils to disperse due to immersion-wetting was in the order M > D > R.

Table 6.3

Size-distribution of soil after rapid immersion in deionised water. Standard errors for n=3 are shown in parentheses.

| | Soil material (%, by weight) | | |
|---|---|---|---|
| Size-class (μm) | Soil D | Soil R | Soil M |
| < 2 | 0.1 (0.1) | 0 | 2.2 (1.1) |
| 2-20 | 0.6 (0.4) | 0 | 16.5 (0.8) |
| 20-53 | 8.3 (0.4) | 5.9 (0.3) | 12.0 (1.4) |
| 53-250 | 58.7 (1.8) | 14.7 (1.1) | 36.4 (1.9) |
| 250-500 | 13.3 (0.3) | 13.0 (0.2) | 14 7 (0.5) |
| 500-1000 | 6.7 (0.5) | 16.3 (0.5) | 8.2 (0.2) |
| 1000-2000 | 5.7 (0.8) | 17.3 (1.9) | 4.7 (0.3) |
| 2000-8000 | 6.5 (0.7) | 32.9 (1.7) | 5.2 (1.1) |

**6.3.3 Extent of aggregate breakdown and dispersion during sonification**

Dispersion by ultrasonic energy is essentially a process where the bonds responsible for soil aggregation are broken (Edwards and Bremner, 1967; Gregorich *et al.*, 1988). Most of the previous work on the breakdown of soil aggregates with ultrasound have focused on the < 20 μm size-fractions of soil. In this study, the dynamics of aggregate breakdown were analysed for a range of aggregate-sizes exceeding 20 μm (Fig. 6.3). For all soils, the proportion of silt- (2-20 μm) and clay-size (< 2 μm) material increased initially with increasing duration of sonification (or energy applied) and became stable afterwards. A similar trend has been reported for other soils with ultrasonic dispersion (North, 1976; Berezin and Voronin, 1981, Moen and Richardson, 1984, Gregorich *et al.*, 1988; Fuller and Goh, 1992; Raine and So, 1993, 1994, 1997). The duration of sonification after which the proportion of < 2 and 2-20 μm material became stable was, approximately, 260, 850, and 170 s for soils D, R, and M, respectively This was equivalent to 950, 3100, and 640 J $g^{-1}$ for soils D, R, and M, respectively.

The maximum amount of silt and clay obtained after sonification was partly dependent on the proportion of silt and clay of each soil before sonification. Therefore, direct comparison between soils for the amount of silt and clay following sonification has little merit. The comparison between soils for the amount of clay was done from:

$$\%\boldsymbol{D}_c = \frac{\mathrm{MAX_c}}{\mathrm{PSA_c}} \times 100, \qquad (6.5)$$

where $\%D_c$ is the dispersion achieved by sonification, and $\mathrm{PSA_c}$ and $\mathrm{MAX_c}$ are the per cent of clay due to particle-size analysis and to 15 minutes of sonification, respectively. Values of PSA and MAX are presented in Table 6.4 Results of $\%D_c$ were 48% for soil D, 24% for soil R, and 94% for soil M. These results show that susceptibility of soils to clay dispersion after 15 minutes of sonification was in the order M > D > R.

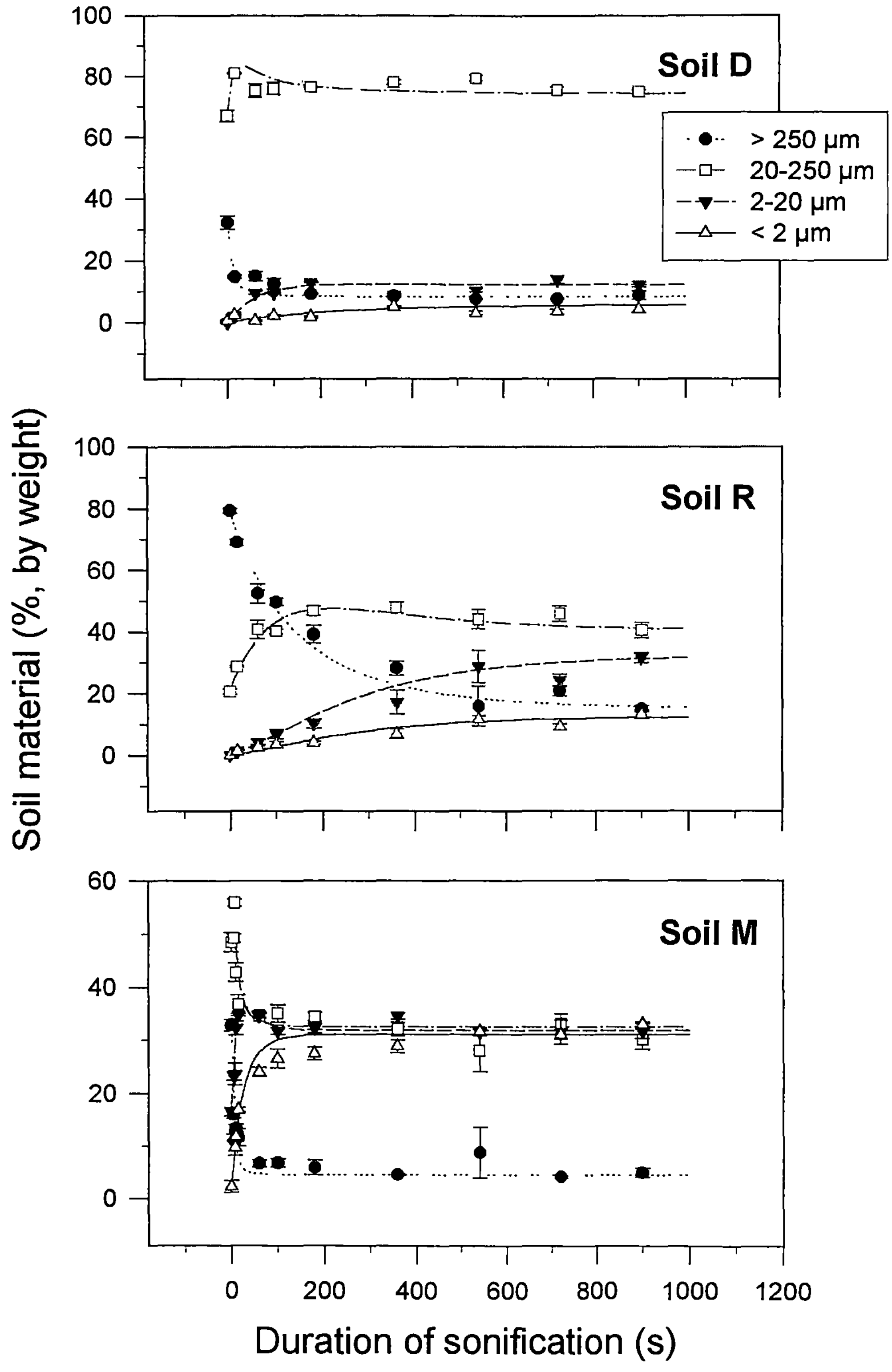


**Fig. 6.3** - Variation in composition of soil material in various size-classes with time of sonification. Symbols represent the measured values, and lines represent the output of a model of aggregate breakdown. Bars on symbols illustrate standard error (n=3) for the measured data.

The change in proportion of aggregates of > 20 μm with increasing duration of applied energy is also given in Fig 6.3. Soil material > 20 μm was grouped in two size-classes, namely microaggregates (20 - 250 μm) and macroaggregates (> 250 μm). Generally, the proportion of macroaggregates for soils D and M declined sharply in the first few seconds of sonification and became constant afterwards. However, for soil R, such a trend took a long duration of sonification. The proportion of microaggregates for soils D and R increased rapidly after sonification started and then became constant. In contrast, for soil M, the temporal variation in microaggregates was similar to that of macroaggregates.

The general pattern of variation for each of the size-fractions shown in Fig. 6.3 follows the breakdown of aggregates of sizes > 250 μm into smaller sizes (< 20 μm) via temporary size-classes of intermediate dimensions. The stable quantity of > 250 μm for a soil may be considered as the proportion in that fraction which is unlikely to be disaggregated further by increasing the duration of sonification, as these materials could be sand. Similar stable values for clay- and silt-sized material for each soil may be taken as the maximum value for these fractions that could be obtained after maximum dispersion with the amount of ultrasonic energy applied. However, if higher energy had been used (*i.e.*, greater duration or greater power of sonification), then a size-distribution different than that in Fig. 6.3 might be obtained. This is based on the findings by Gregorich *et al.* (1988), which indicated that aggregates of 1 - 2 mm, exposed to ultrasonic energy (at 120 W of power), could break down after 300 - 500 J $mL^{-1}$ of applied energy (*i.e.*, equivalent to about 1500 - 2500 J $g^{-1}$, where the soil-water suspensions had a ratio of 1.5, with 15 g of soil in 75 mL of water). They also found that it would be necessary to apply energy in excess of 1500 J $mL^{-1}$ (*i.e.*, equivalent to about 7500 J $g^{-1}$ in that study) to obtain soil dispersion similar to that achieved when hydrogen-peroxide treatment is used followed by 16 h of shaking. The maximum amount of ultrasonic energy applied in this work was about 3300 J $g^{-1}$ that was largely below the 7500 J $g^{-1}$ that may be required to disperse the soil completely. However, Fig. 6.3 shows that the dispersion of soil did not increase with increasing duration of sonification, after it reached a maximum. These results suggest that complete dispersion of soil would have been achieved if the power applied had been increased.

### 6.3.4 A comparison of sonification with particle-size analysis

The size-distribution of aggregates obtained with particle-size analysis of Gee and Bauder (1986) (referred to as PSA, Table 6 4) was taken as the ultimate breakdown of aggregates that could be obtained for a soil without destroying the primary particles. The size-distribution of aggregates achieved with the maximum duration of sonification (MAX) is presented in Table 6.4 for comparison.

The differences in size-distribution of soil for MAX and PSA (Table 6 4) suggested that sonification was not as efficient in breaking down aggregates as was the mechanical and chemical method of dispersion. Due to the difference in the amount of soil used in the two methods there was a possibility of the size-distribution of soil to be somewhat different (30 g of soil for PSA, and 8.3 g of soil for MAX). Although there is still no absolute standard of complete breakdown of aggregates (Christensen, 1992), it seems reasonable to assume that, if more energy had been applied to the soil-water suspensions, dispersion of clay would eventually equal the dispersion obtained with PSA

Table 6.4

Size-distribution of soil (%, by weight) after particle-size analysis (PSA) and after 15 minutes of sonification (MAX). Standard errors are in parentheses (n=3 for most cases, with the exceptions of n=4 for PSA of soil D, and n=5 for MAX of soil R).

| | | Size-class (μm) | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | | < 2 | | 2 - 20 | | 20 - 250 | | > 250 | |
| Soil D | PSA | 8.7 | (0.73) | 11.1 | (0.38) | 68.9 | (0.01) | 11.3 | (0.01) |
| | MAX | 4.2 | (1.18) | 12.1 | (0.81) | 74.9 | (0.70) | 8 8 | (1.43) |
| Soil R | PSA | 53.8 | (5.06) | 23.8 | (1.06) | 19.7 | (0.20) | 2.7 | (0.11) |
| | MAX | 13.1 | (1.42) | 31.6 | (1.51) | 40.4 | (2.24) | 14.9 | (0.89) |
| Soil M | PSA | 35.3 | (1.26) | 24.2 | (2.80) | 38.3 | (0.28) | 2.2 | (0.31) |
| | MAX | 33.1 | (0.41) | 31.7 | (1.39) | 30.2 | (1.94) | 5.0 | (0.87) |

There is ample evidence in the literature that soils are commonly exposed to disruptive forces (due to weathering or cultivation) which have less equivalent energy than the maximum energy used in this study (approximately 3300 J $g^{-1}$). Watts *et al.* (1996) estimated that the energy applied to soils during cultivation (mouldboard plough, shallow plough, chisel plough or rotary digger) was about 0.3 J $g^{-1}$. Russell (1973) reported that the energy dissipated by mouldboard ploughing (to a depth of 20 cm) on a heavy soil was about 0.1 J $g^{-1}$. A rainstorm of 75 mm $h^{-1}$ during one hour dissipates 12 J $g^{-1}$ of energy into the soil surface (North, 1976). Bradford *et al.* (1986) reported that a waterdrop of 5 7-mm diameter falling from a height of 13.9 m at terminal velocity had a kinetic energy of 0.004 J. Wischmeier and Smith (1978) estimated that the energy of rain would be < 0.05 J $g^{-1}$ in most natural rainstorms. Truman *et al.* (1990) calculated the KE of simulated rainfall for intensities of 64 and 127 mm $h^{-1}$ during a 30-minute event as 0 13 J $cm^{-2}$. Therefore, the disruptive forces experienced by soils during PSA and sonification (except in the first few seconds of sonification) are too extreme when compared with what occurs in nature or during tillage. It appears that the amount of energy applied to soil is not meaningful unless duration (or time) is considered, together with the efficiency with which work is done for a given amount of energy.

## 6.4 MODELING THE DYNAMICS OF AGGREGATION/ DISAGGREGATION

### 6.4.1 Description of the model

The variation of soil material in each size-class with increasing time of sonification is referred to as the dynamics of aggregation/disaggregation. It is a complex dynamics due to the co-existence of soil material of several size-classes in aggregated and non-aggregated form, with varying inter-particle and inter-aggregate bonds of different strengths.

The conceptual model presented here represents each size-class of a soil as a compartment defined by the amount of soil material in that size-class The compartments are linked to each other by flows that represent the transfer of soil material from one size-class to another due to aggregation or disaggregation of material in another compartment. For simplification, the soil was divided into four size-classes: < 2 (clay), 2 - 20 (silt), 20 - 250 (microaggregate), and > 250 µm (macroaggregate). Fig. 6.4 shows a schematic diagram of the model. the compartments, $X_i$, represent the amount of soil (in g) of size-class i (with i = 1,...,4);

and the arrows, $F_{ij}$ (with i, j = 1, ..,4), represent the fluxes of soil material from compartment $X_i$ to compartment $X_j$. The model assumes that soil material can be broken down into smaller sizes and that aggregates may form from combination of aggregates of smaller sizes. A notional interpretation of these fluxes is given later (Section 6.4.3).

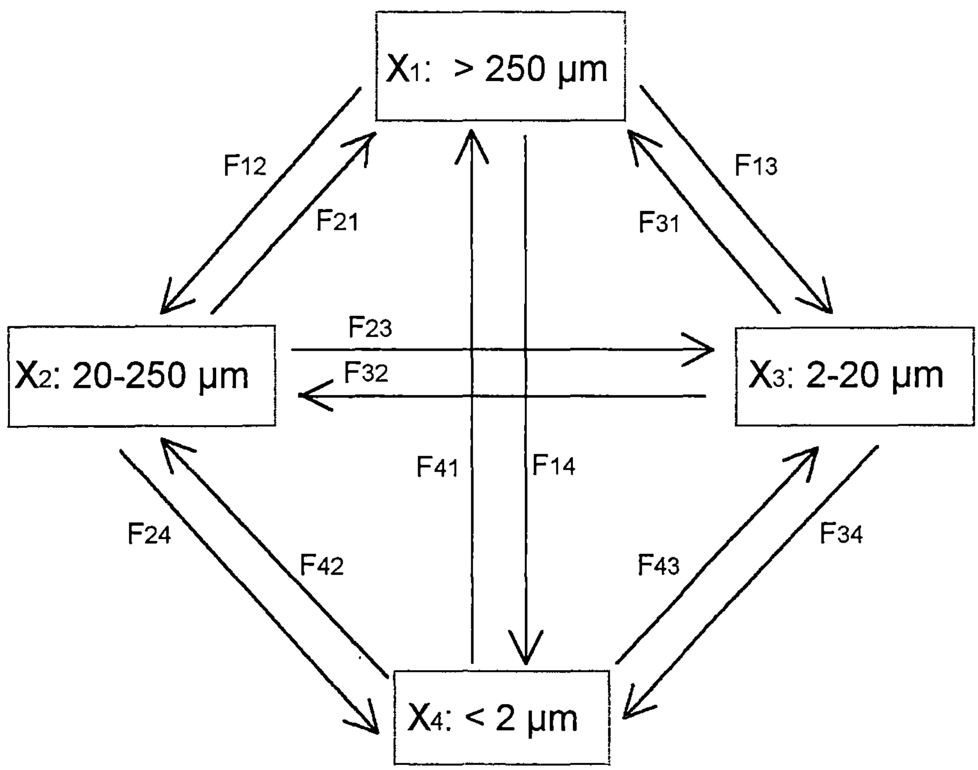


**Fig. 6.4** - Diagram of the model of the dynamics of aggregate breakdown. Each compartment ($X_i$) represents the amount of soil in a distinct size-class. The arrows ($F_{ij}$) represent fluxes of soil from compartment $X_i$ to $X_j$.

The model of aggregation/disaggregation is based on the assumptions that: (1) the flux, $F_{ij}$ (g $s^{-1}$), is proportional to the amount of soil material present within the compartment from where the flux originates ($X_i$), (2) the rate constant of flux, $f_{ij}$ ($s^{-1}$) remains constant for a given duration of sonification, and (3) for a given soil, rate constants of each compartment are independent Therefore,

$$F_{ij} = f_{ij} X_i . \tag{6.6}$$

$F_{ij}$ represents the *rate of aggregation/disaggregation.* The rate of change of soil in each compartment with a change in duration of sonification ($dX_i/dt$) is given for a constant i and j≠i by:

$$\frac{dX_i}{dt} = \sum_{j=1}^{4} F_{ji} - \sum_{j=1}^{4} F_{ij} , \tag{6.7}$$

where $F_{ji}$ represents the rate of influx of soil from the compartment $X_j$ to compartment $X_i$ For example, for i=2, equation (6.7) may be expressed as

$$\frac{dX_2}{dt} = F_{12} + F_{32} + F_{42} - F_{21} - F_{23} - F_{24} . \tag{6.8}$$

The model was restricted to a condition that the net change of amount of soil with time was zero:

$$\sum_{i=1}^{4} \frac{dX_i}{dt} = 0 . \tag{6.9}$$

The rate constants ($f_{ij}$) were estimated by optimisation, so that the model reproduces the observed behaviour of aggregation/disaggregation. The initial values of soil material in each size-class corresponding with the distribution of soil after immersion-wetting (mean values and SE, as shown in Table 6.3) were used as the input data for the model.

The software ModelMaker, version 3 (Walker, 1997) was used to provide numerical methods of optimisation (the Marquardt and simplex methods) to obtain the parameters $f_{ij}$ such that the output from the model was in close agreement with the experimental data in Fig. 6.3. The parameters of the model were adjusted iteratively to minimise the deviation between the predicted and observed $X_i$. The optimisation assumed the errors on the experimental data to be normally distributed. The comparison of the model output with observed data was made to determine the goodness-of-fit with the calculated $\chi^2$-value. Essentially, the method calculated $\chi^2$ based on the deviations of predicted values from observed values, taking into account the standard errors (SE) of observed $X_i$ as weighting factors (weighted by 1/SE), and then made small changes in the parameters until $\chi^2$ reached a minimum.

### 6.4.2 Outputs of the model

The dynamics of aggregate breakdown were modelled for varying duration of sonification at a constant power. Thus, total applied energy was proportional to the duration of sonification. The rate constants ($f_{ij}$) for various combinations of the fluxes ($F_{ij}$) and compartments ($X_i$) are given in Table 6.5. These rate constants may be interpreted in a way similar to that of biochemical reactions, for which the inverse of a rate constant is proportional to the time required to achieve a specific degree of disaggregation or aggregation of soil in the related compartment. For all soils, when the fluxes $F_{13}$, $F_{31}$, $F_{14}$, $F_{41}$, $F_{34}$ and $F_{43}$ were introduced in the model, they resulted in large standard errors between measured and estimated values, and in significant correlation between groups of $f_{ij}$. These results imply that there was little direct breakdown of macroaggregates ($X_1$) into silt ($X_3$) and clay ($X_4$) sizes without an intermediate phase ($X_2$). Therefore, these fluxes mentioned above were considered negligible and taken to be zero, which simplified the diagram of the model shown in Fig. 6.4.

Table 6.5

Mean values and standard errors (in parentheses) of the rate constants as estimated by the model of aggregate breakdown for soils D, R, and M.

| Rate constants ($s^{-1}$) | Soil D | | Soil R | | Soil M | |
|---|---|---|---|---|---|---|
| $f_{12}$ | 0.0795 | (0.0051) | 0.0066 | (0.0003) | 0.2364 | (0.0122) |
| $f_{21}$ | 0.0087 | (0.0006) | 0 0024 | (0.0002) | 0 0324 | (0.0032) |
| $f_{23}$ | 0.0024 | (0.0002) | 0.0022 | (0 0004) | 0.0561 | (0 003) |
| $f_{32}$ | 0.0151 | (0.0017) | 0.0028 | (0.0007) | 0.0572 | (0.0038) |
| $f_{24}$ | 0 0003 | (0.0001) | 0.0009 | (0.0002) | 0.0229 | (0.0013) |
| $f_{42}$ | 0.0041 | (0.0012) | 0.0030 | (0.0010) | 0.0239 | (0.0016) |

The model-predicted amount of soil material in each size-class for various duration of sonification is shown as lines, together with the experimental data in Fig. 6.3. The predicted data agreed well with the experimental observations, with $r^2$ ($p < 0.001$) of 0.94, 0.98, and 0.82 for soils D, R, and M, respectively

The rates of aggregation/disaggregation ($F_{ij}$) for various soils and compartments are presented in Fig. 6.5 Values of $F_{ij}$ and the corresponding rate constant in Table 6.5 were mostly in the order M > D > R. As a result, the time required to obtain stable values of $F_{ij}$ was the longest for soil R in comparison with soils D and M. For all soils studied, the rates of aggregation/disaggregation and corresponding rate constants (in Table 6 5) were generally in the order macroaggregates > microaggregates > silt. Such trends agree well with the present knowledge of the process of aggregation and the hierarchical order of aggregation (Edwards and Bremner, 1967; Moen and Richardson, 1984, Gregorich *et al.*, 1988). These trends also agree with the results obtained on the mechanical stability of these soils in Chapter 4.

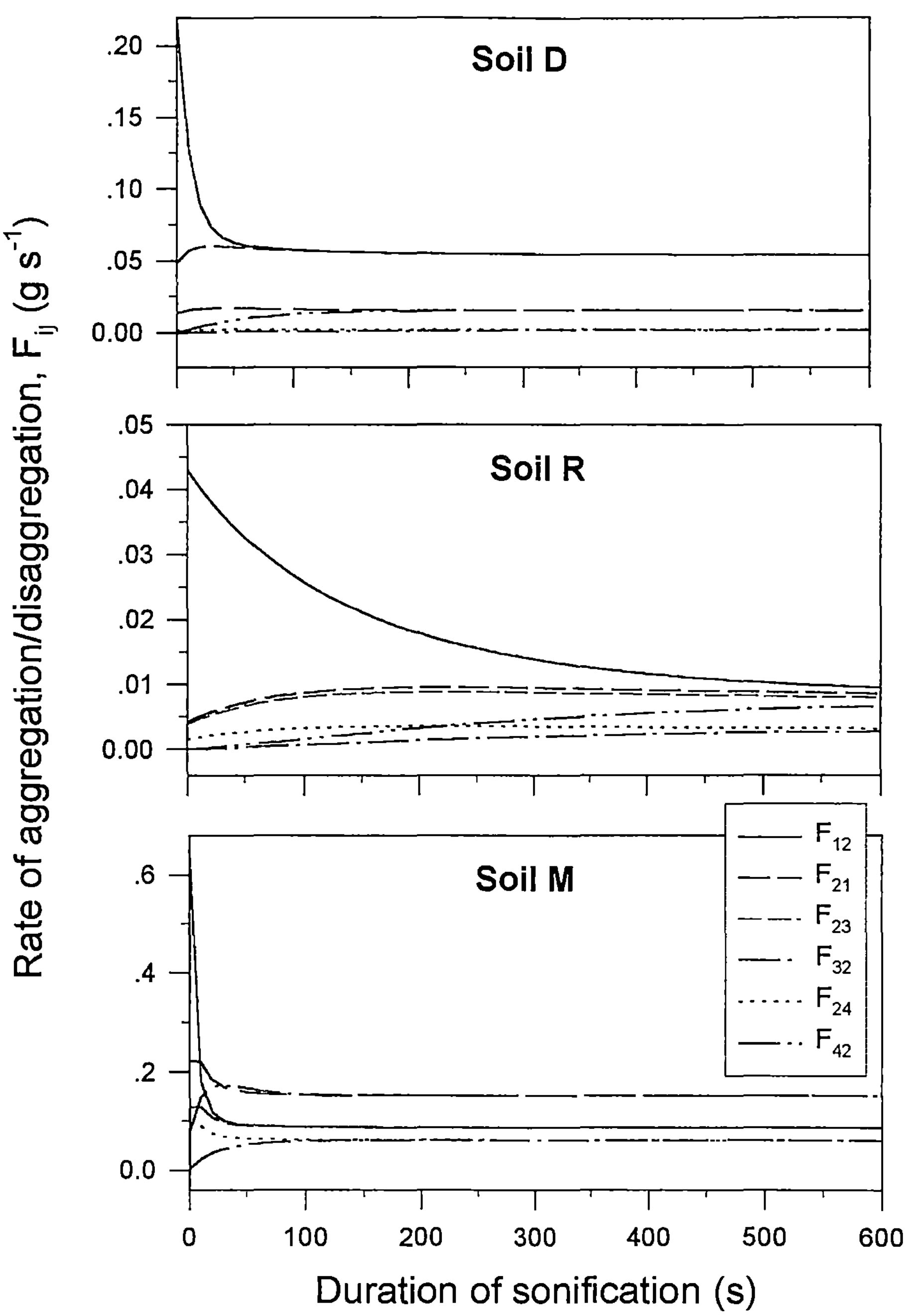


**Fig. 6.5** - Variation in the rate of aggregation/disaggregation ($F_{ij}$) of various soils with duration of sonification. The legend, $F_{ij}$, represents the fluxes of soil from compartment $X_i$ to compartment $X_j$. A maximum duration of 600 s is presented here because after that duration there was little variation in $F_{ij}$. Note the differences in scales of the y-axis for various soils.

**6.4.3 The conceptual basis of the fluxes ($F_{ij}$)**

The approach used in the modelling of two-way fluxes (Fig. 6.4) is based on the notion of biochemical reactions. The model uses a mass balance of aggregates in various compartments for varying durations of sonification. The model does not require a detailed knowledge of underlying processes. Therefore, a qualitative understanding of the processes has been utilised in building the model by identifying the variables (*i.e.*, compartments $X_i$) that influence the processes and the magnitude of their effects in increasing or reducing the fluxes (*i.e.*, the rate constants, $f_{ij}$).

The fluxes in the direction of decreasing size of aggregates describe disaggregation, which include partial and complete collapse of aggregates as evident in studies of slaking and dispersion. The process of disaggregation is enhanced by an increase in applied energy (or increased duration of sonification).

The fluxes of soil material from smaller to larger sizes represent flocculation or re-aggregation of soil material. Flocculation occurs due to coalescence of soil colloids (of diameter < 10 μm) in suspension (Sposito, 1994). Although flocculated soil has a lower bonding strength than that of naturally aggregated soil, flocculation may assist formation of aggregates. Factors, such as clay mineralogy, type and concentration of electrolytes, pH, and the composition of exchangeable cations influence flocculation. It has been shown by Hinds and Lowe (1980) that flocculation of soil following sonification of Gleysolic soils was largely due to the release and dissolution of Fe, Al, Si, and C from soil.

Flocculation during sonification may also occur due to the release of metals from soil during sonification, which may combine with organic matter to form and then precipitate as organometallic complexes (Eriksen *et al.*, 1995), or due to denaturation of organic matter as a result of high temperature close to the ultrasonic probe-tip (Eriksen *et al.*, 1995). More experiments are needed to verify the possibility of flocculation of soils during and/ or after sonification.

It is also possible that increased application of energy with sonification may affect the forces between clay particles and aid flocculation (Sposito, 1994). In a floccule, the bond between particles are weak; therefore, it is unlikely that flocculation could occur during sonification. However, during cooling of the soil-water suspension following sonification, flocculation may have occurred (Busacca *et al.*, 1984; Christensen, 1992).

### 6.6 CONCLUDING REMARKS

This study showed that energy used in aggregate breakdown ($L_g$) is difficult to obtain with ultrasonification due to the high sensitivity of the method to energy-related components, and due to a lack of simplicity with the measurement and estimation procedure.

A model of aggregation/disaggregation of soils was developed to describe the dynamics of aggregate breakdown during sonification. Although it is difficult to relate the model-parameters with the mechanisms of aggregation and disaggregation, it may serve as a tool to obtain quantitative information to describe breakdown of aggregates under disruptive forces for various soil types.

The rates of aggregation/disaggregation for the soils studied were in the order M > D > R, which agreed with the results obtained on the mechanical stability of these soils (Chapter 4), and with the erodibility parameters given in Chapter 5.

Further refinement of the method to study aggregation/disaggregation is required, which may consider maintenance of a constant temperature of the suspension during ultrasonification with a probe of known power, or a source of energy other than that obtained with an ultrasound probe.

# Chapter 7

## NITROGEN LOSS DUE TO EROSION: MEASUREMENT AND PREDICTION

### *7.1 INTRODUCTION*

Loss of nutrients due to erosion has often been attributed to the selective nature of erosion and deposition processes (Stoltenberg and White, 1953; Rose and Dalal, 1988) leading to sediment that is rich in organic matter and clay (Bedell *et al.*, 1946; Stoltenberg and White, 1953; Barrows and Kilmer, 1963; De Bano and Conrad, 1976; Menzel, 1980; Alberts and Moldenhauer, 1981; Flanagan and Foster, 1989). Organic residues are among the first of the constituents to be removed through erosion because of their tendency to concentrate in the surface soil and having a density lower than the other constituents (De Bano and Conrad, 1976). Fine soil material also tends to erode easily (Barrows and Kilmer, 1963; Flanagan and Foster, 1989), presumably due to its lower ability to undergo deposition compared to coarse soil material.

Most soils tend to lose a greater percentage of organic matter and fine material when sediment concentration is low than when it is high. This selective loss of soil nutrients is interpreted to arise from the type of erosion processes involved. Erosion processes due to rainfall can be described in terms of the impact of raindrops on soil causing localised shear stresses and subsequent breakdown of aggregates. In contrast, shear stress arising from runoff water is distributed over a large area and hence, sediment concentration in a runoff-driven erosion event can be much greater than that during a rainfall-driven erosion event (Misra and Rose, 1995), but may not allow as much breakdown of aggregates. The size and composition of sediment is ultimately influenced by the extent of aggregate breakdown caused by rainfall and runoff (Loch and Donnollan, 1982; Proffitt *et al.*, 1993), and by deposition on the erosion surface before the sediment reaches the point of measurement (at the downslope end of an erosion plot). The extent of aggregate breakdown during erosion depends mainly on the structure of soil, the energy associated with raindrop impact, and the depth of water on the soil surface. Without a significant depth of water protecting the soil surface, the outer layer of some soil aggregates may be peeled off by raindrop impact producing fine soil material (a mechanism described as *raindrop stripping* by Ghadiri and Rose 1991a). The outer layer of aggregates may be richer in nutrients than the inner layer, thus peeling off aggregates can produce fine material richer in nutrients than the material of the same size in the uneroded soil (Ghadiri and Rose, 1991a, b)

More than 90% of nutrients (particularly N) are lost from agricultural land by erosion, other than by processes such as leaching or volatilisation. Losses due to erosion are estimated to be in the order of 1 to 100 kg N $ha^{-1}$ $year^{-1}$ (White, 1986, Rose and Dalal, 1988), implicating erosion as one of the major causes of long-term decline in fertility of agricultural soils. Similar information on nutrient loss is currently limited for forest soils used for tree farms. In tree farms, the nutrient concentration of surface soil can be high due to the abundance of woody residue that is left during harvest of the previous tree crop. Although large and heavy residues may act as contact covers in minimising erosion, observations made during the erosion experiments of the present study indicate that the woody residues which break down into fine and partially decomposed material, may be readily removed by runoff water.

Nitrogen loss in an erosion event is the amount of N in sediment ($S_N$, kg ha$^{-1}$), i.e.,

$$S_N = S_L \cdot N_S, \tag{7.1}$$

where $N_S$ is the N-concentration of sediment (kg kg$^{-1}$), and $S_L$ is sediment lost per unit area (kg ha$^{-1}$). In the erosion literature, the quality of sediment as a potential pollutant of waterways is often described in terms of enrichment ratio ($E_R$). In the context of this work, $E_R$ refers to the ratio of nitrogen in the sediment compared to that in the uneroded soil, i.e.,

$$E_R = N_S / N_U, \tag{7.2}$$

where $N_U$ is the N-concentration of the uneroded soil (kg kg$^{-1}$). An $E_R > 1$ for a particular nutrient denotes the sediment to be richer than the uneroded soil, and an $E_R < 1$ denotes impoverishment of the sediment in nutrient compared with the uneroded soil. As the sediment loss ($S_L$) can be predicted with erosion models for single events (e.g. WEPP, Lane and Nearing, 1989; GUEST, Misra and Rose, 1996) or for multiple events on an annual basis (e.g. USLE, Wischmeier and Smith, 1978; RUSLE, Renard *et al.*, 1991), N-loss may be predicted from the measurement of $N_S$ using equation (7.1) or from the knowledge of $E_R$ and $N_U$ using equation (7.2).

Menzel (1980) developed a relationship between $E_R$ and sediment loss by simplifying the equations proposed by Massey and Jackson (1952):

$$\ln(E_R') = u + m\ln(S_L), \tag{7.3}$$

where $E_R'$ was the predicted value of $E_R$, and $u$ and $m$ were regression parameters.

Regression equations are useful for predictive purposes, although these are empirical and therefore do not provide a theoretical basis of the underlying processes relating to erosion and nutrient loss. Palis *et al.* (1990a) tested the theoretical framework of Rose and Dalal (1988) for contrasting experimental situations in which erosion was due to rainfall only or due to the combined effect of rainfall and runoff. This framework assumed that N-concentration of any size-class of sediment was similar for the N-concentration of the same size-class of uneroded soil. This assumption worked well for situations where there was a substantial depth of water on the surface of the soil (Palis *et al*, 1990a). A substantial depth of water on the soil surface reduces the effect of raindrop impact on the soil, thus reducing aggregate breakdown. Although the depth of water was not measured in the erosion experiments of the present study, observations made during the rainfall simulations indicated the depth of water on the soil surface to be lower than the size of raindrops (Chapter 2) on most occasions. Therefore, there was opportunity for substantial raindrop impact on soil and possibility of aggregate breakdown to occur. Aggregate breakdown may allow sediment of various size-classes to have different N-concentration than the same size-classes of the uneroded soil (Palis *et al*, 1997). Consequently, the framework of Palis *et al.* (1990a) could not be used to estimate $E_R$ of N in this study.

The objectives of the present study were: (1) to report N-loss for erosion events with simulated rain on bare forest soils of varying structure; (2) to investigate possible factors contributing to the enrichment of sediment in N, and (3) to evaluate Menzel's method of predicting N-loss due to erosion for the forest soils.

## *7.2 METHODS*

Three forest soils, without any surface cover, were subjected to simulated rainfall of 116 mm $h^{-1}$ in erosion trays of 0.8 $m^2$. Four erosion treatments were applied to all three forest soils, and each combination of soil and erosion treatment was replicated three times. Thus, there were altogether 36 simulated erosion events. The details of the experimental design and methods were given in Chapter 2. Briefly, the erosion treatments were: treatment 1 with low rainfall kinetic energy (KE) and low slope (2 °); treatment 2 with high rainfall KE and low slope; and treatment 3 with high rainfall KE and high slope (16 °). Treatment 3 was sub-divided into treatments 3a and 3b, which were identical in all respects except that 3b experienced a drying cycle of 14 days prior to erosion.

The soils used were Dover (D), Ridgley (R), and Maydena (M). A description of these soils was given in Chapter 2. Soil D was a poorly aggregated loamy sand with 3.63 % of organic carbon (OC); soil R was a strongly aggregated clay with 9.25 % OC; and soil M was a clay loam with low aggregate stability when wet and 2.16 % of OC.

Uneroded soil (sampled prior to erosion), and sediment (sampled at 0-5, 12.5-17.5, and 35-40 min after runoff started) were wet-sieved into six size-classes (> 2000, 2000-1000, 1000-500, 500-250, 250-53, and < 53 μm). Size-distribution of the uneroded soil was obtained for two samples collected before each erosion event started (n=24 for each soil) Size-distribution of sediment was obtained after each sampling of sediment within an erosion event (n=12 for each soil). Sediment sampled in the first and last 5 minutes of erosion (0-5 and 35-40 minutes) was subjected to N analysis. These sampling times were selected to examine temporal variation of N-concentration with duration of erosion.

The uneroded soil and sediment retained in each sieve were analysed for total nitrogen (total-N) as described in Chapter 2. Total-N was analysed for each size-class of uneroded soil that was replicated three times for a soil type. Total-N of each size-class of sediment at 0-5 minutes of erosion for each soil and for the erosion treatments 2 and 3a was analysed for three replicates. Total-N of the other sediment samples was based on one replicate. Three replicates of the whole samples of uneroded soil (without sieving), for each combination of soil and erosion treatment, were also analysed for total-N. The total-N of the whole uneroded soil obtained by two methods (*i.e.*, direct measurement of the total-N of the whole unsieved soil, and estimate of total-N from the distribution of N-concentration over size and the size-distribution of soil) was not significantly different. This indicated that sampling and measurement errors were similar for both methods. Therefore, N-concentration of the whole sediment samples was calculated on the basis of total-N of the sediment in each size-class and the size-distribution of sediment.

## *7.3 RESULTS*

### 7.3.1 Nitrogen distribution over size of uneroded soil

Statistical analysis of N-concentration for the bulk uneroded soil ($N_U$) indicated N-concentration to be significantly higher ($p < 0.01$) for soil R (0.004 kg $kg^{-1}$, SE = 0.00010), than for soils D (0 0014 kg $kg^{-1}$, SE = 0.00003) and M (0.0014 kg $kg^{-1}$, SE = 0.00008).

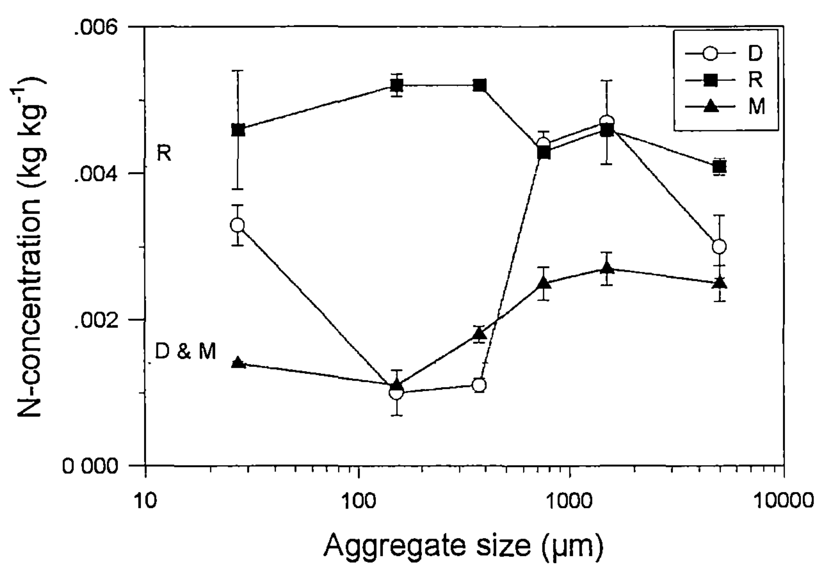


**Fig. 7.1** - Distribution of N-concentration over mean size of aggregates for the uneroded soils D, R, and M. Horizontal dotted lines represent the mean N-concentration for the whole soil samples. Error bars indicate SE of mean values (n = 3); but some error bars were smaller than the size of symbols.

The variation of N-concentration over the size of soil aggregates (Fig. 7.1) was the highest for soil D (coefficient of variation, C.V. = 57%), followed by soil M (C.V. = 33%) and soil R (C.V. = 10%). Overall, soil R had the highest N-concentration for all size-classes (Fig. 7.1), and soil M the lowest N-concentration for all sizes except 250-500 μm. Soils D and R had similar N-concentration in size-classes of 500-1000 and 1000-2000 μm. N-concentration higher than the whole soil was found in aggregates of < 53 μm and > 500 μm of soil D; for all but the fraction > 2000 μm of soil R; and for all except aggregates < 250 μm of soil M (Fig 7.1).

### 7.3.2 Sediment loss and N-concentration in sediment

Sediment losses due to erosion ($S_L$, Mg $ha^{-1}$) were significantly influenced by soil type ($p \leq 0.001$) and by erosion treatment ($p \leq 0.001$), and there was a significant interaction between soil type and erosion treatment ($p \leq 0.001$). Although $S_L$ at 0-5 minutes of erosion was significantly greater than that at 35-40 minutes ($p = 0.001$), interaction between time of sampling and soil and/ or erosion treatment was not significant. Values of $S_L$ in 5 minutes of erosion (averaged over 0-5 and 35-40 minutes of erosion) for all soils and erosion treatments are presented in Fig. 7.2. For all erosion treatments, sediment loss was in the order M > D > R. There was little effect of erosion treatments on sediment loss for soil R. For the other two soils, $S_L$ increased with increasing slope and rainfall KE.

The size-distribution of sediment in the beginning (0-5 minutes), middle (12.5-17.5 minutes), and end of erosion (35-40 minutes) for each soil and erosion treatment is presented in Figs. 7.3-7.5 together with the size-distribution for the uneroded soil. For all soils of treatments 1 and 2 (low slope) there was always a greater proportion of soil material in the finest fraction (< 53 μm) of the sediment than in the uneroded soil. In addition to this, the sediment of soil D contained a higher proportion of material in the size-class of 500-2000 μm compared with the uneroded soil. There was also a greater proportion of material in the size-class of 1000-2000 for the sediment than the uneroded soil of soil R in erosion treatment 1 (*i.e.*, low kinetic energy of rain and low slope).

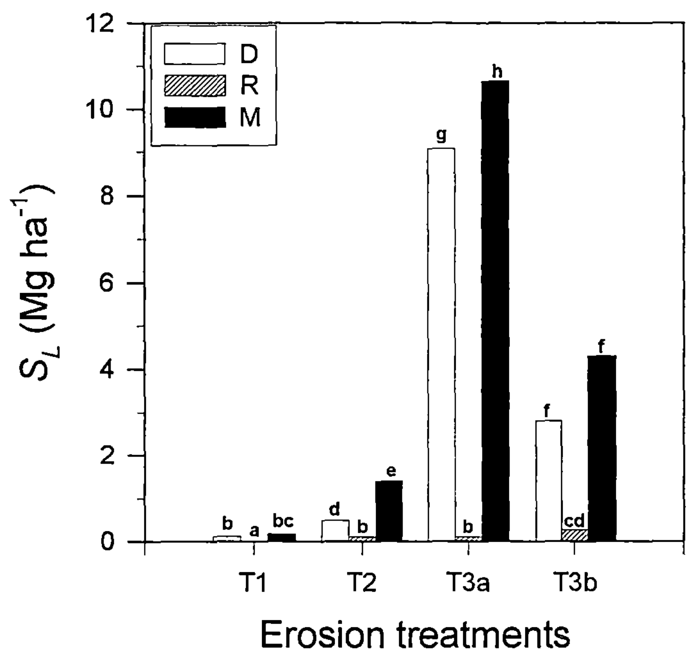


**Fig. 7.2** - Mean sediment loss ($S_L$) for various soils and erosion treatments. Bars with different letter(s) denote significantly different values of $S_L$ ($p < 0.05$).

With high-slope erosion treatments (treatments 3a and 3b), the material of < 53 μm remained higher in sediment than the uneroded soil for the soils R and M. For the soil D, the proportion of material in 250-500 μm size was greater in sediment than the soil. There was some indication of a greater proportion of 1000-2000 μm size in sediment than the uneroded soil at longer duration of erosion for the treatment 3b of soil R.

The reasons for the dominance of a certain size in sediment compared to the uneroded soil for various soils and erosion treatments will not be dealt further in this section because similar details are already given in Chapter 4, on the basis of Mean Weight Diameter (MWD) of sediment and soils.

The N-concentration in each size-class of sediment (sampled at the beginning and end of erosion) and of the uneroded soil, for various soils and erosion treatments are presented in Figs. 7.6-7.8. In most cases, N-concentration of a size-class of sediment was greater than the concentration for the same size-class of the uneroded soil. For most size-classes of all soils, the difference in N-concentration between sediment and soil was greater for erosion treatments 1 and 2 (low slope) than for treatments 3a and 3b (high slope). Overall, there was no clear trend in the variation of N-concentration of sediment with time of sampling.

### 7.3.3 N-loss and enrichment ratio ($E_R$)

N-loss ($S_N$, kg ha$^{-1}$) during 5 minutes of erosion (taken as the mean N-loss for the 0-5 and 35-40 minutes of erosion) for each soil and erosion treatment is given in Fig. 7.9. For the erosion treatments of low slope (treatments 1 and 2), values of $S_N$ were in the order D > M > R, and for the erosion treatments of high slope, in the order M > D > R, despite N-concentration of sediment ($N_S$) being generally in the order R > D ≥ M (Figs. 7.6-7.8). For all soils, N-loss was greater at high KE (treatment 2) than at low KE (treatment 1) and at high slope than at low slope. The variation of $S_N$ (Fig 7.9) for various soils and erosion treatments was similar to that of $S_L$ (Fig. 7.2).

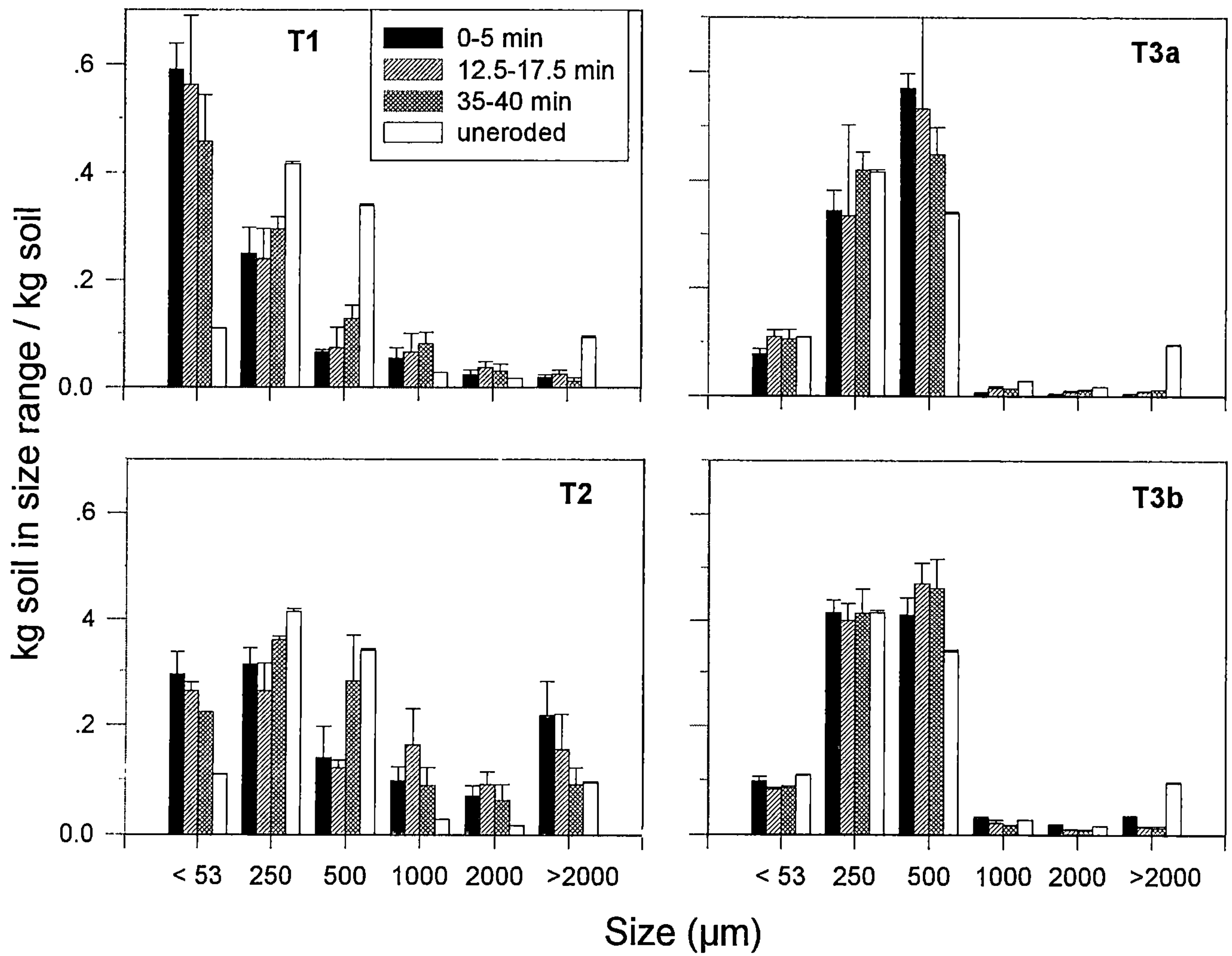


**Fig. 7.3** - Soil D: size-distribution of sediment at three times of sampling and of the uneroded soil for various erosion treatments (1, 2, 3a, and 3b). Note that here and in other figures of this type, a size of 250 µm denotes 53-250 µm; 500 µm denotes 250-500 µm; 1000 µm denotes 500-1000 µm; 2000 µm denotes 1000-2000 µm.

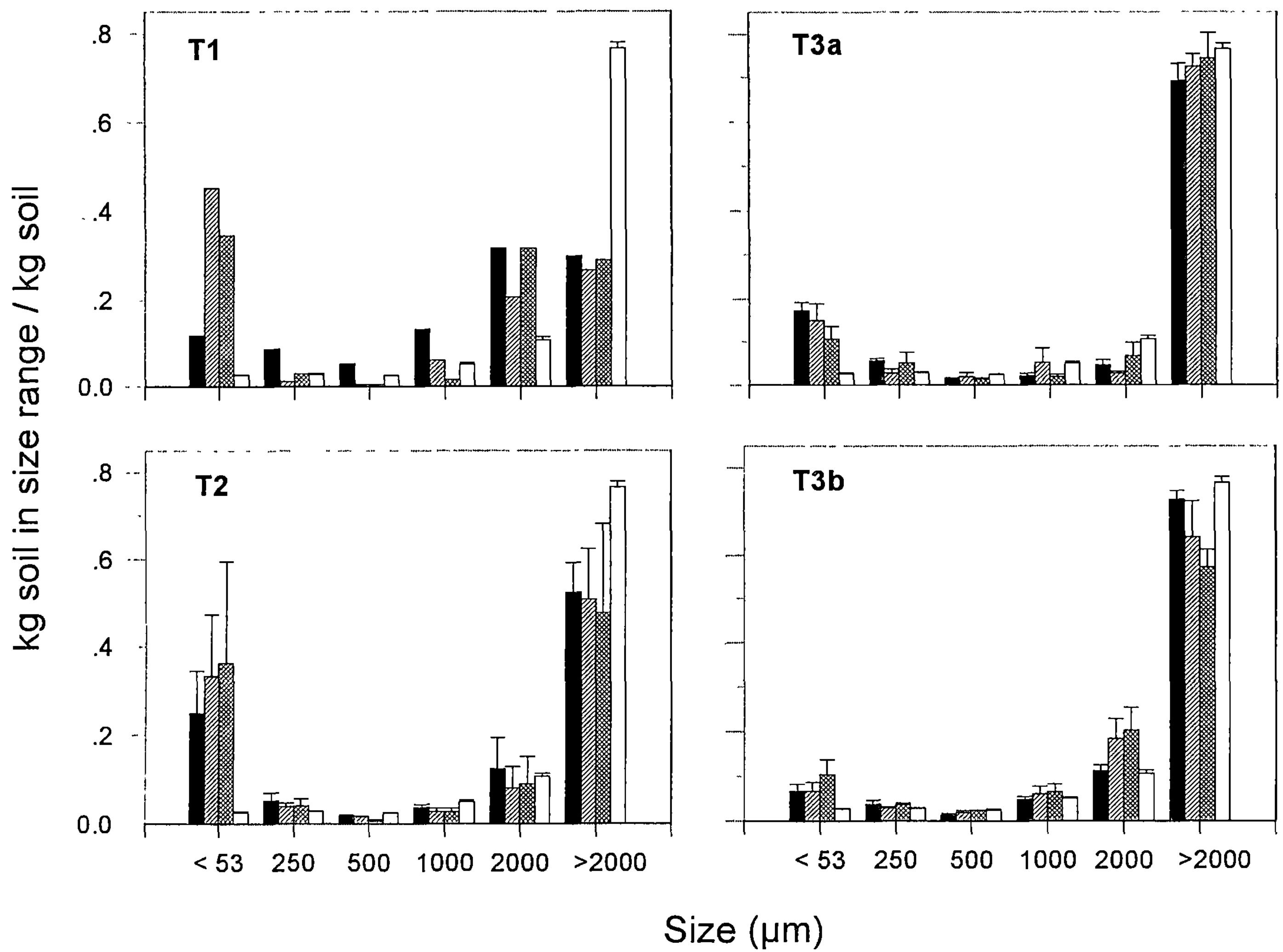


**Fig. 7.4** - Soil R: size-distribution of sediment at three times of sampling and of the uneroded soil for various erosion treatments (1, 2, 3a, and 3b). Same legend as for Fig. 7.3.

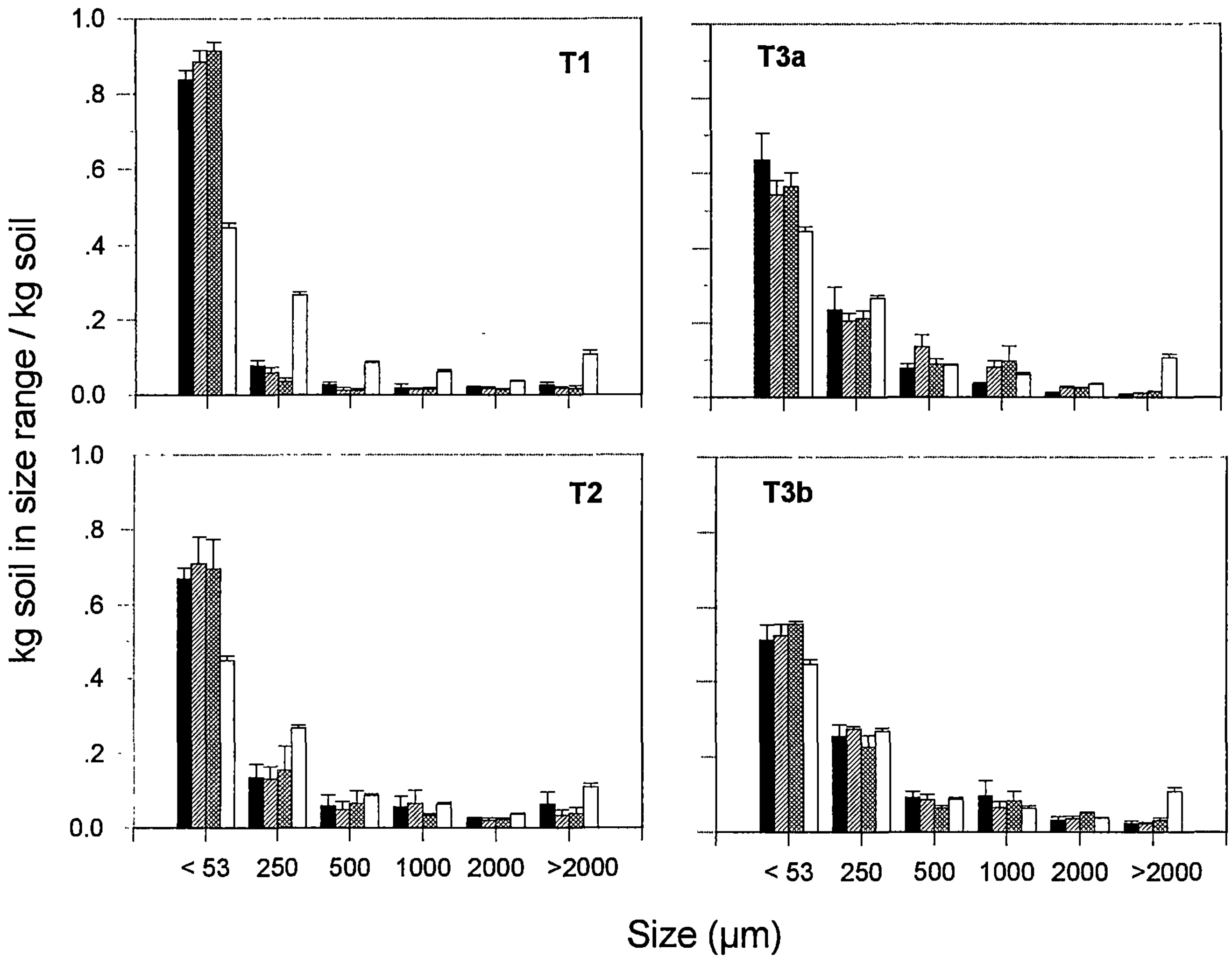


**Fig. 7.5** - Soil M: size-distribution of sediment at three times of sampling and of the uneroded soil for various erosion treatments (1, 2, 3a, and 3b). Same legend as for Fig. 7.3.

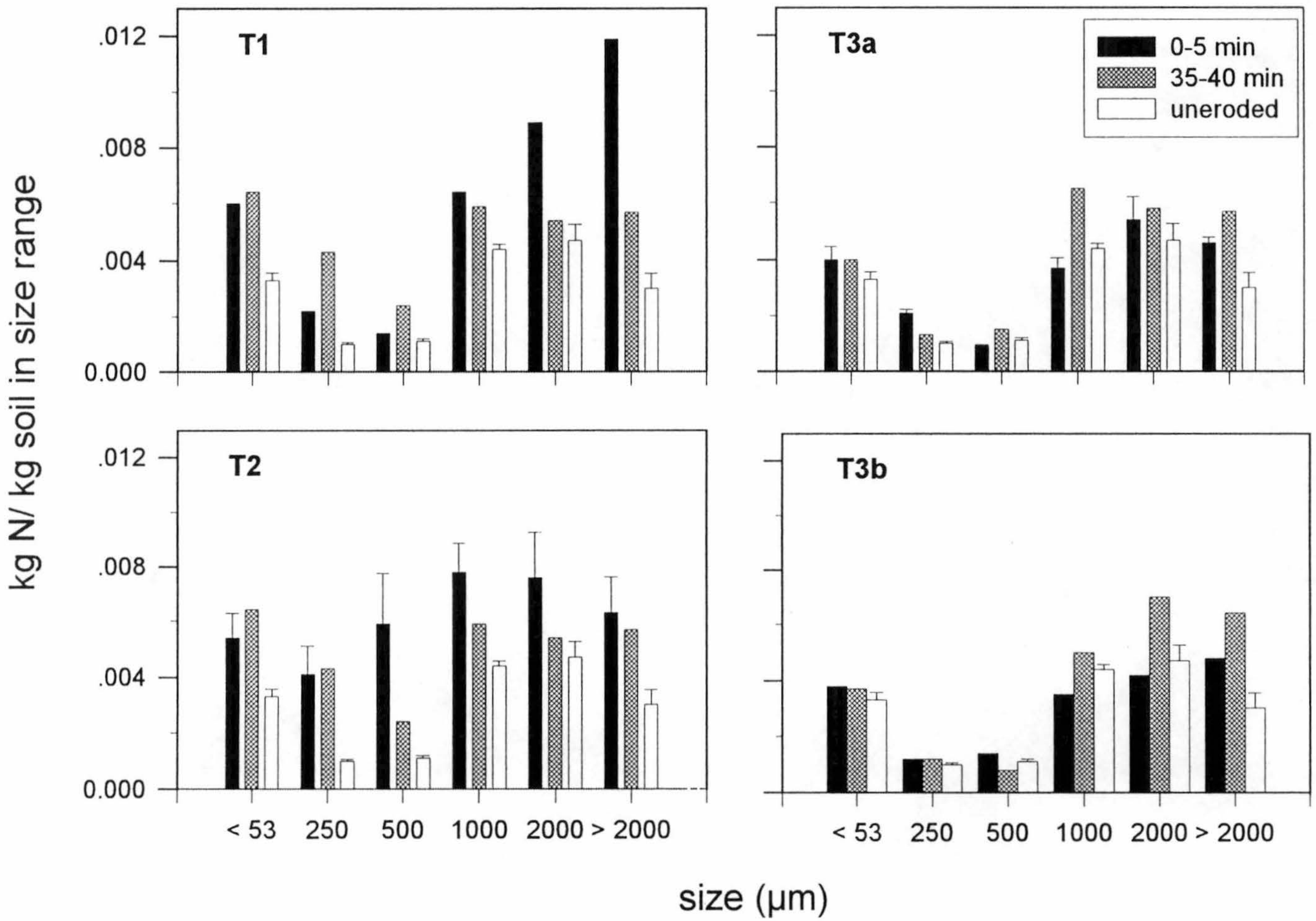


**Fig. 7.6** - Soil D: distribution of N-concentration over size for sediment at two sampling times and for the uneroded soil exposed to various erosion treatments (1, 2, 3a, and 3b). Size refers to a size-class as indicated in Fig. 7.3.

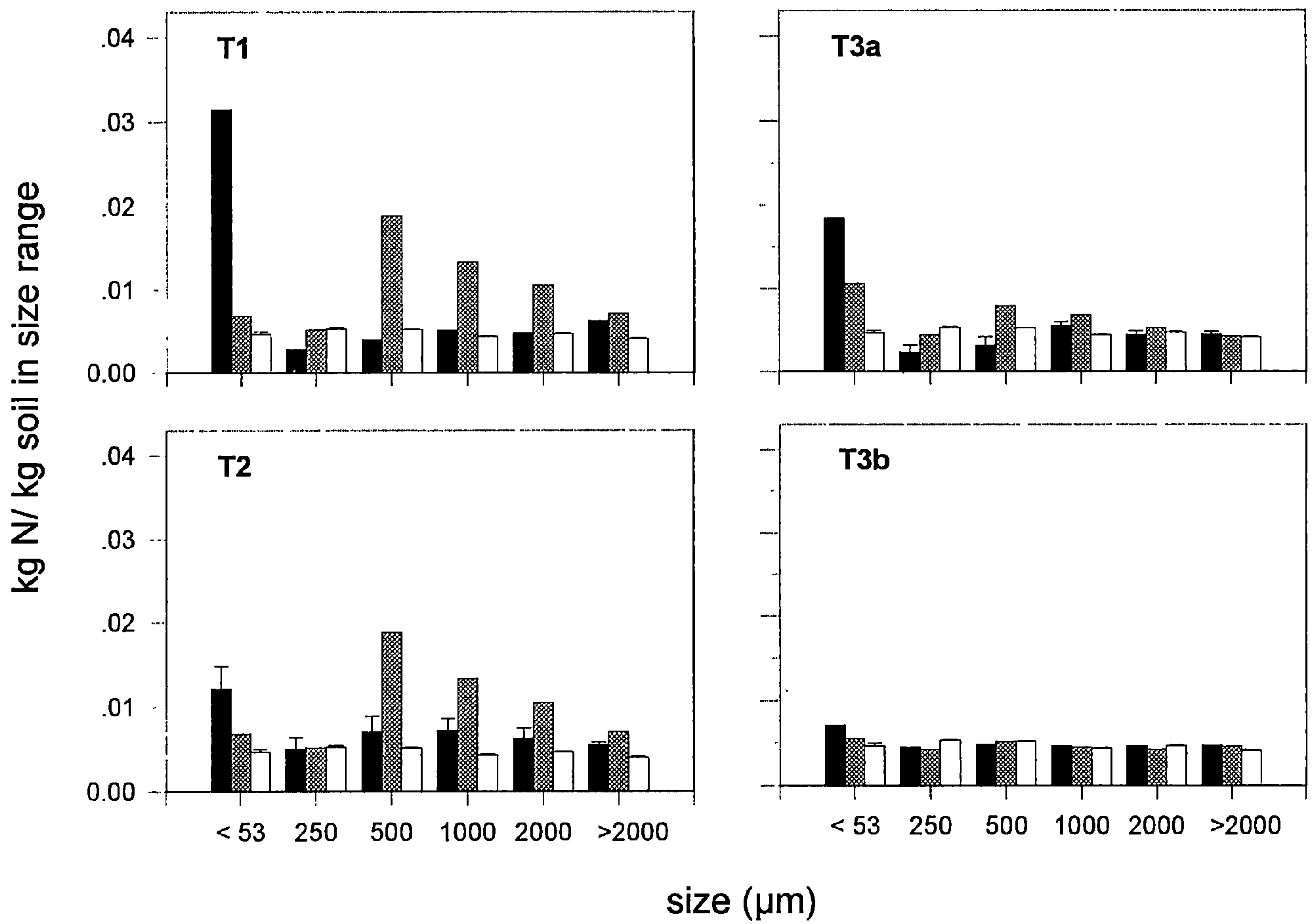


**Fig. 7.7** - Soil R: distribution of N-concentration over size for sediment at two sampling times and for the uneroded soil exposed to various erosion treatments (1, 2, 3a, and 3b). Size refers to a size-class as indicated in Fig. 7.3. Note the difference in scale of y-axis for this figure compared with the previous figure.

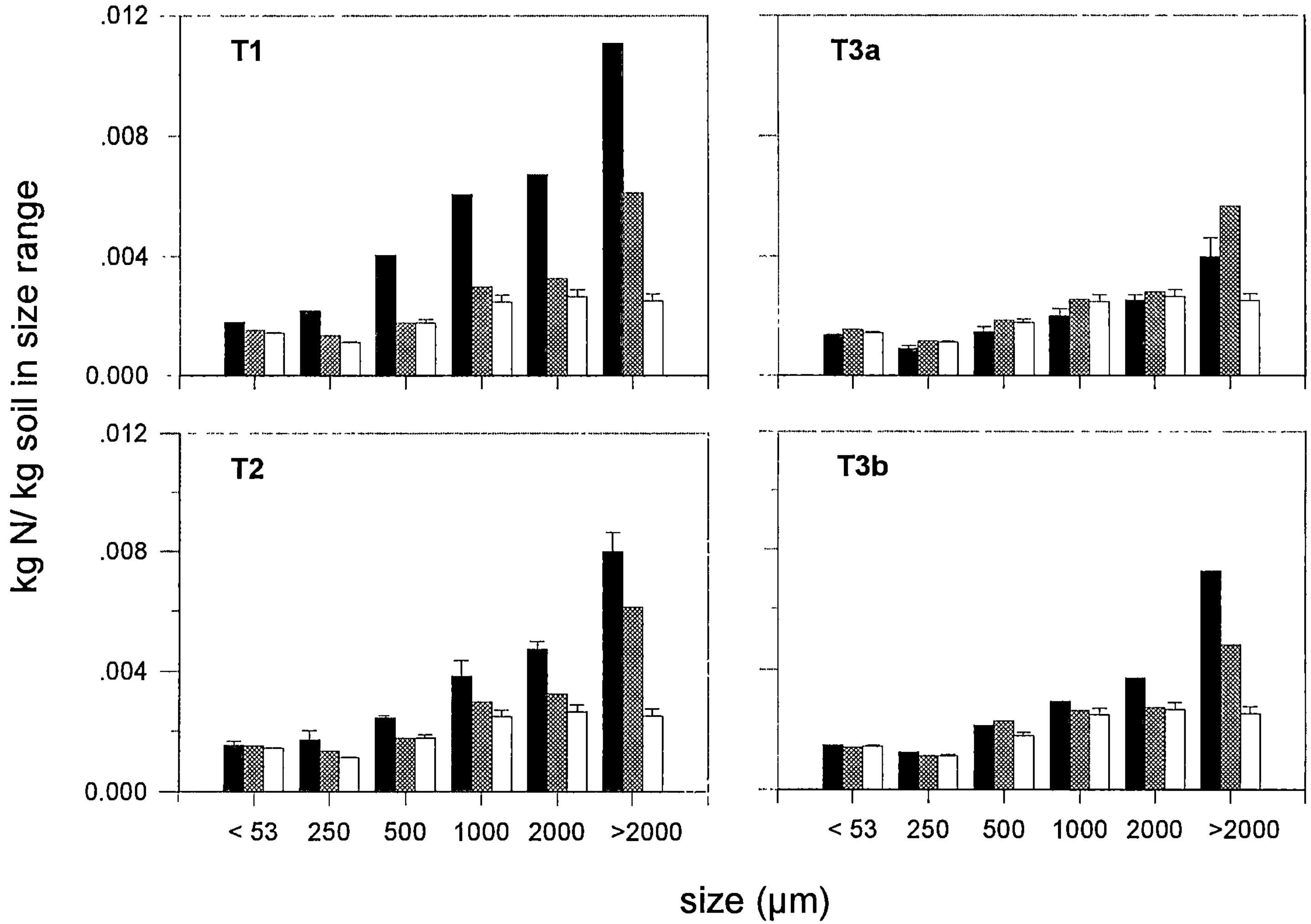


**Fig. 7.8** - Soil M: distribution of N-concentration over size for sediment at two sampling times and for the uneroded soil exposed to various erosion treatments (1, 2, 3a, and 3b). Size refers to a size-class as indicated in Fig. 7.3.

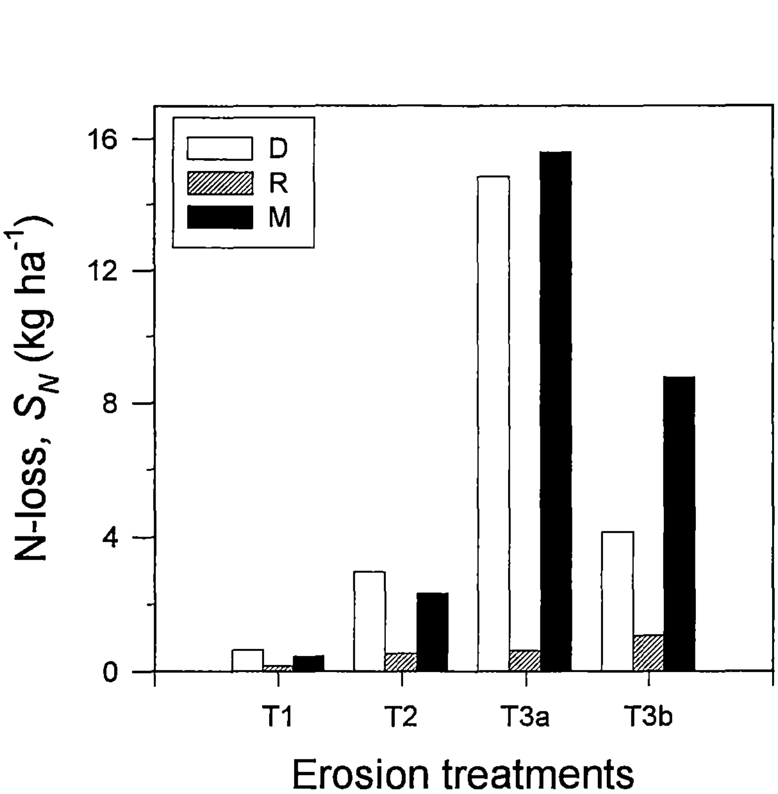


**Fig. 7.9** - N-loss for various soils and erosion treatments, during 5 minutes of erosion.

The overall concentration of N in sediment was greater than that in the uneroded soil, as indicated by values of $E_R > 1$ (Fig. 7.10). For low slope treatments (1 and 2) $E_R$ was in the order D > R > M, and for high slope treatments (3a and 3b) $E_R$ was in the order R > D > M.

Massey and Jackson (1952), Sharpley (1985), Rose and Dalal (1988), Palis *et al.* (1990a), McIsaac *et al.* (1991), Catt *et al.* (1994), Hansen and Nielsen (1995), and Sombatpanit *et al.* (1995) measured $E_R$ of N for a variety of agricultural soils under different crops, cultivation and weather conditions. A mean $E_R$ of 1.86 (SE = 0.139, n = 88) was calculated from these data cited above. The mean $E_R$ calculated over all erosion treatments and the soils D, R, and M was similar to that for the agricultural soils ($E_R$ = 1.97; SE = 0.199, n=24 from 3 soils × 4 erosion treatments × 2 sampling times).

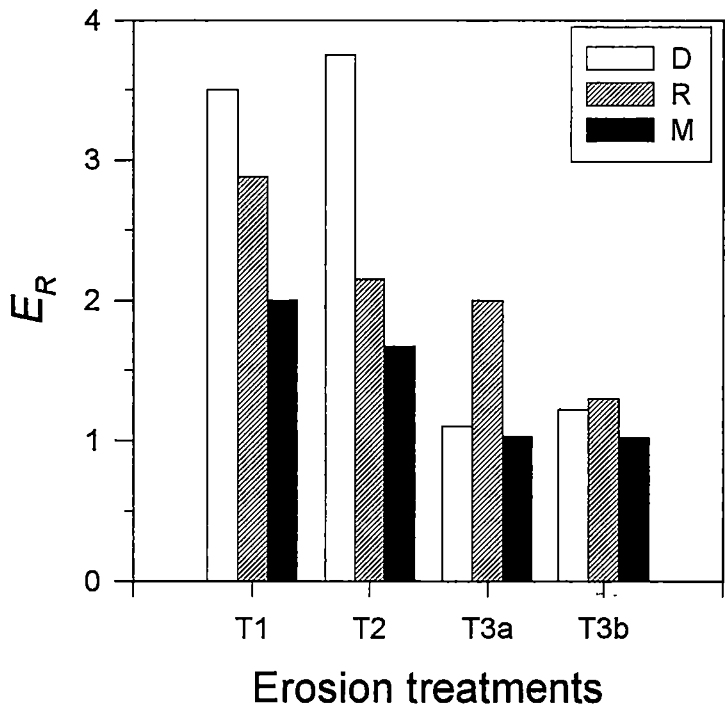


**Fig. 7.10** - Enrichment ratio of N ($E_R$) for various soils and erosion treatments.

### 7.3.4 Prediction of enrichment ratio and N-loss

It is often acknowledged in the erosion literature that if $E_R$ can be predicted then N-loss can be estimated without analysis of sediment for N. Here, measured values of $E_R$ were compared with predicted values of $E_R$ (denoted as $E_R'$). Values of $E_R'$ were obtained with the method of Menzel (equation 7.3) with parameters $u$ = 2 and $m$ = -0.2 (as suggested by Menzel, 1980). Results show that $E_R'$ using Menzel's method generally overestimated $E_R$ for all soils (Fig. 7.11). The values of $E_R'$ shown in Fig. 7.11 were further used to predict N-loss ($S_N'$) with equation (7.1) and with measured values of sediment loss ($S_L$, Mg ha$^{-1}$). A comparison of $S_N'$ with measured $S_N$

indicated a significant linear relationship (Fig. 7.12) ($r^2 = 0.98$, $p < 0.001$). However, the regression line of $S_N'$ versus $S_N$ had a slope of 1.22 ($\pm$ 0.031), that was significantly different ($p > 0.05$) from the 1:1 line. These results (Figs. 7.11 and 7.12) suggest that inadequacy in the estimation of $E_R$ had a small influence on the predicted N-loss, if sediment loss was measured.

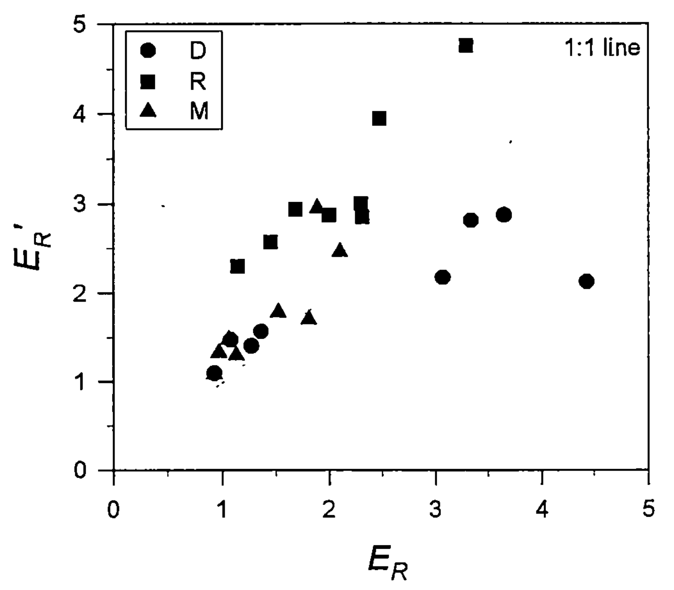


**Fig. 7.11** - $E_R$ predicted with Menzel's equation ($E_R'$) versus measured $E_R$ for various soils, erosion treatments and two sampling times.

Although estimation of $E_R'$ with Menzel's equation allows rapid estimation of N-loss, N-concentration of the uneroded soil ($N_U$) would still be required. To avoid determination of $N_U$, it may be useful to obtain a regression between $S_N$ (in kg ha$^{-1}$) and $S_L$ ( in kg ha$^{-1}$) that is independent of $E_R'$. Thus,

$$S_N' = \alpha_1 + \alpha_2 S_L \tag{7.4}$$

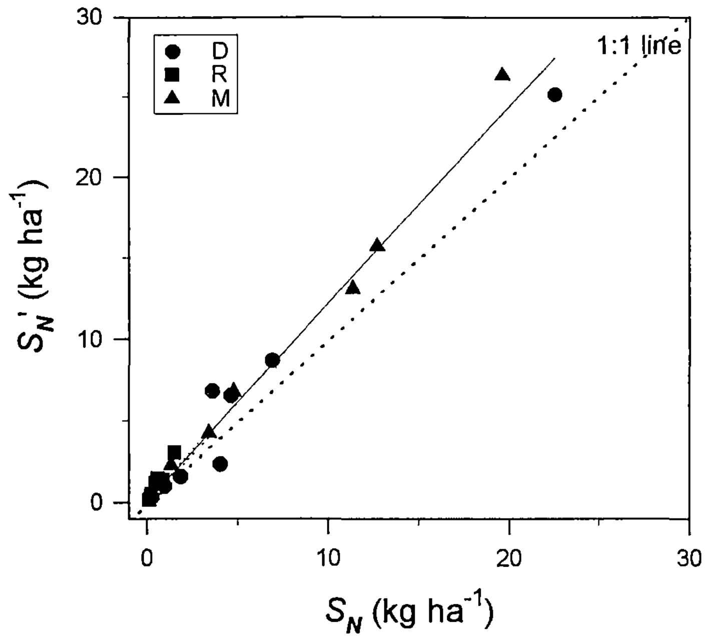


**Fig. 7.12** - N-loss predicted with Menzel's equation ($S_N'$) as a function of measured N-loss ($S_N$) for various soils, erosion treatments and sampling times. The solid line indicates linear regression. The 95% confidence intervals for the regression are shown as dashed lines.

The parameters $\alpha_1$ and $\alpha_2$ of equation (7.4) were obtained by fitting a linear regression to the data on $S_N$ and $S_L$ available in the literature (Table 7.1). These data on erosion and N-loss were obtained from sites used for a range of agricultural crops (alfalfa, wheat, potatoes, barley, corn, sorghum), natural vegetation (rangeland) and bare soil (fallow), and a variety of slopes and climates, and from experimental plots of area ranging from 1 m$^2$ to 4.8 ha. The regression parameters of equation (7.4) based on the published data indicated in Table 7.1 (n = 151) were $\alpha_1 \approx 0$, and $\alpha_2 = 0.00161$ (SE = $4\times10^{-5}$), with $r^2 = 0.92$ ($p < 0.001$) (Fig 7.13).

Table 7.1

Characteristics of the erosion experiments from which the data on $S_L$ and $S_N$ were used to obtain regression parameters of equation (7.4). The number of data available (n) is given. Natural and simulated rain are indicated by N and S, respectively

| Location | Plot area ($m^2$) | Plot slope (%) | Soil Texture | Soil cover | n | Type of rain | Source |
|---|---|---|---|---|---|---|---|
| California (USA) | 36 | 1-3 | loam | rangeland | 4 | N | De Bano & Conrad (1976) |
| Mississippi (USA) | 100 | 5 | silt loam | agr. crops | 5 | N | McDowell & McGregor (1980) |
| Illinois (USA) | 33 | 3-5 | silt loam | agr. crops | 56 | S | McIsaac *et al.* (1989, 1991) |
| UK | 864 | 9 | sand | agr. crops | 12 | N | Catt *et al.* (1994) |
| Denmark | 66 | 10 | loamy sand, sandy loam | agr. crops, fallow | 36 | N | Hansen & Nielsen (1995) |
| Texas (USA) | 4.0-4.8 ha | 1-3 | loam | agr. crops | 6 | N | Richardson & King (1995) |
| Thailand | 46 | 4 | loam | agr. crops, fallow | 10 | N | Sombatpanit *et al.* (1995) |
| Texas (USA) | 1 | n.m. | clay | agr. crops | 4 | S | Torbert *et al.* (1996) |
| Canada | 100 | 9-10 | clay, sandy loam, sandy clay loam, clay loam | agr. crops, fallow | 18 | N | Hargrave & Shaykewich (1997) |

n.m. - not mentioned

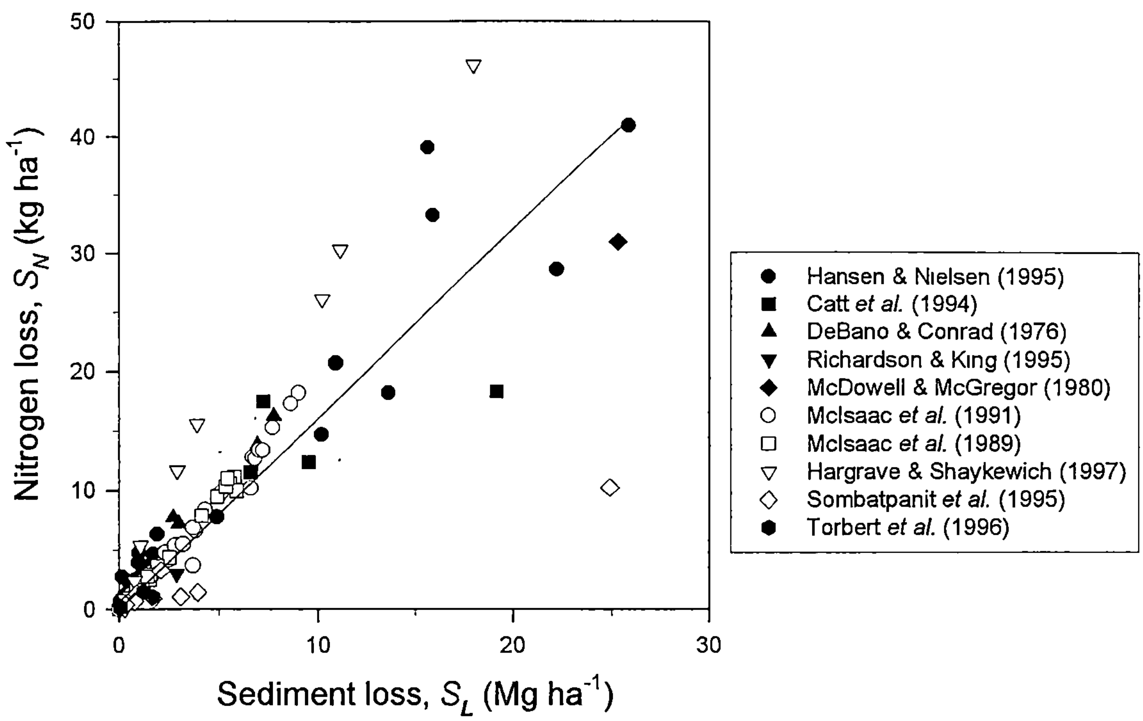


**Fig. 7.13** - Variation of N-loss ($S_N$) with changes in sediment loss ($S_L$) for a variety of erosion conditions. Dotted lines correspond to 95 % confidence intervals of the regression line. Note the difference in units of $S_L$ in this figure and equation (7.4).

## *7.4 DISCUSSION*

### 7.4.1 N-concentration in sediments, N-loss, and enrichment ratio of N

A comparison of N-loss for various soils and erosion treatments (Fig. 7.9) during 5 minutes of erosion indicated the lowest N-loss (0.2 kg ha$^{-1}$ for soil R), for the soil with the lowest sediment loss (Fig. 7.2). N-loss was the highest for soil M (> 15 kg ha$^{-1}$), which also had the highest sediment loss for all erosion treatments (Fig 7.2). Thus, the soil with the highest fertility (as indicated by N-concentration in Fig. 7.1), soil R, was not only less susceptible to erosion but also to loss of fertility, when compared with soils D and M. The losses of N due to erosion reported here were averaged for the first and last 5 minutes of an erosion event.

Irrespective of the size-class of sediments, the N-concentration was greater than that of the uneroded soil (Figs. 7.6-7.8). In order to examine the reasons for enrichment of N in sediment, *concentration ratio* ($C_R$) of N for each size-class of sediment and soil is presented in Table 7.2. The term concentration ratio is referred to as the enrichment ratio of N for a single size-fraction of the sediment. As a contrast, a similar ratio for the whole sample of sediment is the *enrichment ratio* ($E_R$).

The data in Table 7.2 showed that for all soils, values of $C_R$ were mostly higher in treatments 1 and 2 (low slope) than in treatments 3a and 3b (high slope). The overall effects of the variation in $C_R$ with a change in slope carried through to produce a similar variation in $E_R$ with slope (Fig. 7.10). These findings are in agreement with observations that losses of nutrients and organic matter are greater for situations where rainfall dominates over runoff and erosion is low, than when runoff predominates and erosion is high (Stoltenberg and White, 1953; Rose and Dalal, 1988). Due to the absence of a definitive trend in the variation of $C_R$ with size, it is possible that either the distribution of N was uneven in various parts of single aggregates, or that raindrop stripping did not occur. During fragmentation of aggregates, there may be a possibility that the aggregates break down into various sizes. Therefore, the fragments richer in nutrient than the bulk soil can be as available for erosion and deposition as the fragments which have a similar or lower concentration compared with the bulk soil. These rich fragments may include partly or fully decomposed plant residue and particulate organic matter. If such nutrient-rich mineral and organic material is readily available for removal during erosion, then $C_R$ would remain > 1 and would not decline to < 1, even if erosion continues for 40 minutes (Table 7.2).

### 7.4.2 Prediction of enrichment ratio and N-loss

Equation (7.3) is used in the CREAMS model (a field scale model for Chemicals, Runoff, and Erosion from Agricultural Management Systems; Knisel, 1980) to predict the enrichment ratio of both nitrogen and phosphorus due to erosion. The regression parameters of equation (7.3), calculated for the data available in this study, were $u = 2.6 \pm 0.55$ and $m = -0.13 \pm 0\ 09$ ($p < 0.05$), which were within the range suggested by Menzel (1980) ($u = 2 \pm 1$, and $m = -0.2 \pm 0.1$). However, it can be seen from Fig. 7.11 that prediction of $E_R'$ with the parameters $u$ and $m$ of Menzel, did not agree well with the observed $E_R$. These results indicated that small differences in the parameters of the equation (7.3) may produce large differences between estimated and observed $E_R$. The parameters of equation (7.3) calculated from the data of other authors (Table 7.3) differed significantly from the values suggested by Menzel (1980), although Sharpley (1980) and Ghadiri and Rose (1991b) reported these parameters to be in reasonable agreement with Menzel (1980). These conflicting findings suggest that Menzel's equation may not apply to all erosion situations.

Table 7.2

Concentration ratio ($C_R$) of N for various size-classes of sediment arising from various soils and erosion treatments at two sampling times (0-5 and 35-40 minutes).

| | Soil D | | Soil R | | Soil M | |
|---|---|---|---|---|---|---|
| TREATMENT 1 | | | | | | |
| Size-class (µm) | 0-5 min | 35-40 min | 0-5 min | 35-40 min | 0-5 min | 35-40 min |
| > 2000 | 4.01 | 1.66 | 1.54 | 1.16 | 4.41 | 4.27 |
| 2000-1000 | 1.89 | 1 11 | 1.01 | 0.95 | 2.52 | 3.97 |
| 1000-500 | 1.46 | 1.19 | 1.18 | 1 18 | 2.43 | 3.81 |
| 500-250 | 1.29 | 2.00 | 0.76 | 1.79 | 2.27 | 2.61 |
| 250-53 | 2.17 | 2.26 | 0.53 | 4.02 | 1.96 | 2.25 |
| < 53 | 1.79 | 1.66 | 6.79 | 4.98 | 1.25 | 1.27 |
| TREATMENT 2 | | | | | | |
| Size-class (µm) | 0-5 min | 35-40 min | 0-5 min | 35-40 min | 0-5 min | 35-40 min |
| > 2000 | 2.12 | 1.93 | 1.36 | 1.74 | 3.18 | 2.43 |
| 2000-1000 | 1.61 | 1.13 | 1.36 | 2.26 | 1.79 | 1.23 |
| 1000-500 | 1.76 | 1.34 | 1.65 | 3.07 | 1.54 | 1.20 |
| 500-250 | 4.71 | 2.17 | 1.36 | 3.64 | 1.38 | 1.00 |
| 250-53 | 4.47 | 4.25 | 0.94 | 0.99 | 1.52 | 1.19 |
| < 53 | 1.70 | 1.92 | 2.61 | 1.46 | 1.07 | 1.06 |
| TREATMENT 3a | | | | | | |
| Size-class (µm) | 0-5 min | 35-40 min | 0-5 min | 35-40 min | 0-5 min | 35-40 min |
| > 2000 | 1.55 | 1.91 | 1.08 | 1.03 | 1.58 | 2.25 |
| 2000-1000 | 1.14 | 1.23 | 0.93 | 1.12 | 0.96 | 1.06 |
| 1000-500 | 0.83 | 1.47 | 1.26 | 1.57 | 0.81 | 1.03 |
| 500-250 | 0.85 | 1.39 | 0.60 | 1.52 | 0.83 | 1.05 |
| 250-53 | 2.13 | 1.34 | 0.43 | 0.82 | 0.79 | 1.04 |
| < 53 | 1.13 | 1.22 | 3.98 | 2.29 | 0.95 | 1.08 |
| TREATMENT 3b | | | | | | |
| size-class (µm) | 0-5 min | 35-40 min | 0-5 min | 35-40 min | 0-5 min | 35-40 min |
| > 2000 | 1 62 | 2.17 | 1.15 | 1.12 | 2.88 | 1.91 |
| 2000-1000 | 0.89 | 1.49 | 0.99 | 0.91 | 1.40 | 1.02 |
| 1000-500 | 0.80 | 1.13 | 1.07 | 1.02 | 1.18 | 1.05 |
| 500-250 | 1.24 | 0.76 | 0.93 | 0.98 | 1.20 | 1.27 |
| 250-53 | 1.17 | 1.21 | 0.86 | 0.80 | 1.10 | 1.00 |
| < 53 | 1.14 | 1.10 | 1 53 | 1.18 | 1 02 | 0.97 |

Table 7.3

Parameters $u$ and $m$ of equation (7 3) calculated from published data (with standard errors in parentheses) for the sets of data (n) used to derive the parameters of the regression The statistical significance of the regression equation is indicated with $r^2$.

| Source | $u$ | $m$ | n | $r^2$ |
|---|---|---|---|---|
| McIsaac *et al.* (1991) | 0 26 (0 091) | -0.01 (0.012) | 34 | 0.05 (NS) |
| Catt *et al.* (1994 | -0.77 (0.289) | 0.20 (0.044) | 12 | 0.68 (p=0.001) |
| Sombatpanit *et al.* (1995) | 3.69 (0.403) | -0 29 (0.051) | 10 | 0.81 (p<0.001) |

Although the prediction of $E_R$ with the method of Menzel (1980) was poor for the soils and erosion experiments described here (Fig. 7.11), predicted values of N-loss ($S_N'$) using inaccurate estimates of $E_R$ ($E_R'$) agreed reasonably with the observed N-loss ($S_N$) (Fig. 7.12). Thus, $E_R$ need not be estimated accurately to obtain a good prediction of N-loss. This is because the variation in sediment loss ($S_L$) was much greater than the variation in the N-concentration of sediment ($N_S$) used for the calculation of N-loss. As $S_L$ has a stronger influence than $N_S$ in predicting N-loss, the variation of N-loss ($S_N$) for various soils and erosion treatments in Fig. 7.9 was similar to the variation of $S_L$ (Fig. 7.2).

The linear relationship between N-loss and sediment loss (equation 7.4) based on the data from the literature for various erosion experiments (Table 7.1) had a slope $\alpha_2 = 0\ 00161$ kg kg$^{-1}$ ± 0 00004 (Fig. 7 13). This value of $\alpha_2$ implied that a N-concentration ($N_S$) of 0.0016 kg kg$^{-1}$ (or about 0.16 % N) may be used for a wide range of soils and erosion conditions to obtain a good estimate of N-loss The magnitude of $\alpha_2$ and the type of regression in equation (7.4) was similar to those reported by McIsaac *et al.* (1991), Gachene *et al.* (1997), and Hargrave and Shaykewich (1997). Thus, loss of N due to erosion can be estimated from the knowledge of sediment loss alone.

The value of 0.0016 kg $kg^{-1}$ of N obtained for $\alpha_2$ in equation (7.4) was within the nominal range of N-concentration reported for the cultivated layer of most soils (0.1 to 0.2 %, Foth and Ellis, 1988; or 0.08 to 0.4 %, Bremner, 1965). Due to the relatively little variation in N-concentration in the surface layer of most soils, it is reasonable to use a constant value of N-concentration for the sediment, with measured or predicted sediment loss to estimate N-loss due to erosion.

Predicting N-loss ($S_N'$) for the experimental data of this study with the linear regression in equation (7.4) provided the best estimate, because the slope of the line (1.04 ± 0.04, with $r^2$=0.97, $p < 0.001$) did not differ significantly from the 1:1 line (Fig. 7.14). Compared with the prediction of $S_N$ with Menzel's method (Fig. 7.12), using constant N-concentration for sediment ($\alpha_2$) and equation (7.4), the prediction of $S_N'$ was closer to the observed $S_N$ (Fig. 7.14), which suggested that this equation may have a more general application in erosion.

## *7.5 CONCLUDING REMARKS*

Among the soils studied and erosion treatments imposed, nitrogen loss due to erosion was high for those soils (D and M) and erosion treatments (3a and 3b) where the sediment loss was high.

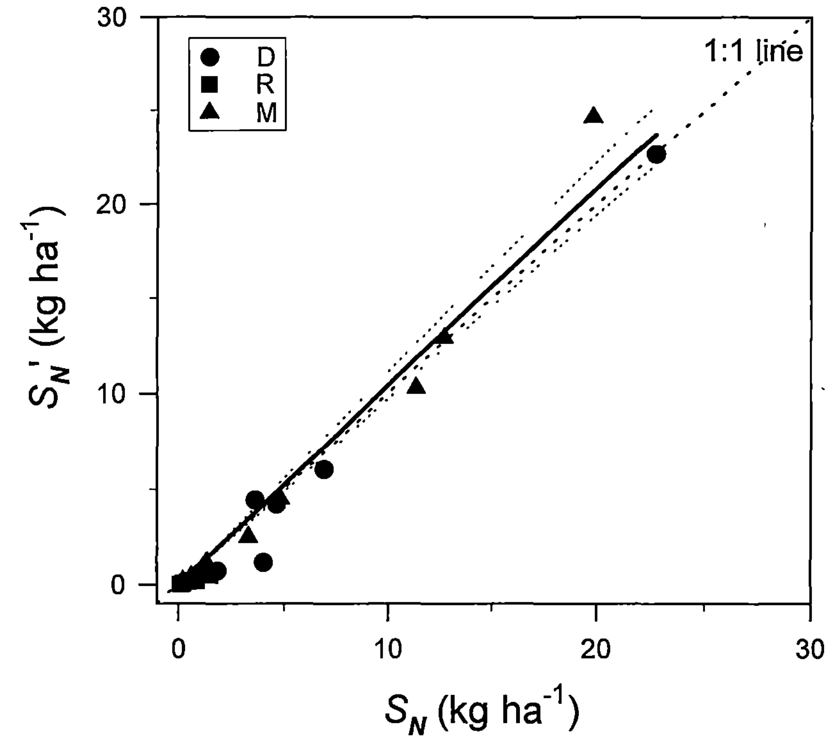


**Fig. 7.14** - N-loss predicted with equation (7.4) ($S_N'$) as a function of measured N-loss ($S_N$) for various soils, erosion treatments and sampling times. The solid line indicates linear regression. The 95% confidence intervals for the regression are shown as dashed lines.

There was no direct evidence of raindrop stripping for the soils studied and erosion treatments used due to a lack of a relationship between concentration ratio ($C_R$) and size of sediment. Nevertheless, both $E_R$ and $C_R$ were generally > 1 indicating the ease with which erosion removes the nutrient-rich component of the soil.

Although predictions of $E_R$ with the method of Menzel (1980) did not agree with the measured $E_R$, predicted N-loss using $E_R$ with Menzel's method agreed reasonably with measured N-loss. The similarity in the variation between sediment loss and N-loss of various soils and erosion treatments suggested that the amount of sediment loss could be used to estimate N-loss.

A linear regression between N-loss and sediment loss obtained with the data from the literature predicted N-loss well for the soils and erosion treatments used in this study. As the method of predicting N-loss proposed in this study is empirical, it needs to be tested further.

# Chapter 8

## GENERAL DISCUSSION AND CONCLUSIONS

Studies of soil erosion in forestry are uncommon when compared with those in agriculture. With recent global expansion of plantations following clearing of native forests, erosion is a pertinent issue in the management of forest soils. A recent review in forestry by Lacey (1993) indicated that the data on erosion and nutrient loss from Australian forests and plantations are almost nonexistent. The research described in this thesis was undertaken to extend the knowledge of erosion studies to plantation soils.

The major objective of this thesis was to determine soil properties that affect erosion and erodibility, so that soil erodibility could be obtained as an independent measure of soil erosion. Measured soil properties were, then, considered to infer erosion directly or via erodibility to avoid the labour and cost of managing erosion plots in the field. It was hypothesised that soil structure (reflecting the mechanical stability of soils) and strength (which is also influenced by structure) were the soil properties which could influence soil erosion. Therefore, quantification of soil structure and strength were considered as measurable quantities in this study. The relationship between aggregate breakdown and soil structure was considered as an important factor for both soil erosion and the characteristics (quality) of sediment. Finally, it was hypothesised that nitrogen loss due to erosion may also be related to soil structure (and indirectly by soil strength) through their influence on the characteristics of sediment and the amount of soil loss.

Experiments were undertaken to meet the above objectives. Results of the experiments on soil erosion in the field were described in Chapter 3 for one soil type. Similar experiments in the laboratory were described in Chapters 4 and 5 to include three forest soils and four erosion treatments. The mechanical stability of soils were characterised to interpret the variation in erosion and sediment characteristics of various soils (Chapter 4). The influence of soil strength and structure on erosion and erodibility of soils was examined in Chapters 3 (field studies) and 5 (laboratory studies). In Chapter 5, erodibility parameters were estimated with the model GUEST and the effects of soil structure and strength on soil erodibility were evaluated. The variation in aggregate breakdown and dispersion with varying duration (amounts) of ultrasonic energy was examined in Chapter 6. Finally, in Chapter 7 variation in nitrogen loss for the laboratory erosion experiments with variation in soil structure was examined to investigate factors contributing to nitrogen enrichment of sediment and to evaluate methods for the prediction of nitrogen loss.

This Chapter provides a synthesis of the main findings of those chapters indicating overall outcomes (conclusions) and the direction for future research. The variability of various measured quantities in the erosion studies are considered first as an important issue to indicate the inherent variability of the system measured, the quality of measurements and the confidence needs to be placed if these quantities are to be predicted in future

### *8.1 VARIABILITY IN MEASUREMENTS OF EROSION*

In studies of erosion with simulated rain of constant rate on a plot of constant length, the runoff rate (Q) and sediment concentration (*c*) are the most important variables which can be measured In laboratory erosion studies (Chapter 4), continuous measurement of these variables were made at a fixed time interval (mostly 2.5 minutes) for events of 40 minutes duration In field erosion studies with the Dover soil (Chapter 3), these variables were measured at slightly different time interval and duration, compared with the laboratory erosion studies Therefore, temporal variability in Q and *c* reported in these studies are partly dependent on the time-scale used for observations. In addition to erosion, the quality of sediment (expressed as Mean Weight Diameter or MWD of sediment) was also another important variable, because the sediment quality was influenced by soil structure (Chapter 4) and influenced nitrogen loss (Chapter 7).

Estimates of variability in field studies of erosion (expressed as coefficient of variation, CV (%) in Table 8.1) showed higher variability in *c* compared with Q and MWD. Although there were no rills in any of the events, the range of CV in *c* for four events indicate that soil condition at the time of erosion (ASM, BD, structure and microtopography variation) may have contributed to such high variation. This variation was dominated by one of the four events (event #3) (Fig. 3.2). As these events were not repeated at the same site, there was an opportunity for the measured variables to be influenced by both the spatial and temporal variation in soil properties.

Table 8.1

Magnitude of variation (CV, %) in sediment concentration (*c*, kg $m^{-3}$), runoff rate (Q, $m^3$ $m^{-2}$ $s^{-1}$), and MWD of sediment (μm) for four field erosion events on soil D. Variation within- and between-events are indicated.

| | CV (%) for | | |
|---|---|---|---|
| Measured variables | Within events | | Between events |
| | Range | Average | Average |
| *c* | 19 - 112 | 49 | 60 |
| Q | 6 - 14 | 10 | 35 |
| MWD | 16 - 41 | 27 | 18 |

Table 8.2

Magnitude of within-event variation (CV, %) in sediment concentration (*c*, kg $m^{-3}$), runoff rate (Q, $m^3$ $m^{-2}$ $s^{-1}$), and MWD of sediment (μm) for 36 simulated erosion events in laboratory studies for three soils. The data include variation over all four erosion treatments.

| | CV (%) for | | | | | |
|---|---|---|---|---|---|---|
| | Soil D | | Soil R | | Soil M | |
| Measured variables | Range | Average | Range | Average | Range | Average |
| *c* | 10 - 41 | 28 | 19 - 118 | 37 | 6 - 41 | 22 |
| Q | 3 - 14 | 8 | 3 - 16 | 8 | 1 - 14 | 5 |
| MWD | 3 - 38 | 22 | 4 - 63 | 14 | 9 - 67 | 24 |

Table 8.3

Magnitude of variation (CV, %) in event-average values of sediment concentration (*c*, kg $m^{-3}$), runoff rate (Q, $m^3\ m^{-2}\ s^{-1}$), and MWD of sediment (μm) for 36 simulated erosion events in laboratory studies for three soils.

| | CV (%) for | | |
|---|---|---|---|
| Soil | *c* | Q | MWD |
| D | 32 | 6 | 30 |
| R | 52 | 10 | 18 |
| M | 29 | 9 | 35 |

The variation in measured variables for laboratory erosion experiments is shown in Table 8.2 as within-event variation. For all soil, the variation in runoff rate (Q) was the lowest and similar to that in the field. Although drainage was not permitted in erosion trays for laboratory experiments (detailed in Chapter 2), the consistency in the variation of Q for field and laboratory experiments indicate that these measurements were most reliable. Note that this may be partly due to the use of simulated rainfall of constant rate in the laboratory and field erosion experiments.

The variation in sediment concentration (*c*) was greater than that for MWD. For soil D, the variability within events in the measurements of MWD were similar for field and laboratory studies (Tables 8.1 and 8.2). However, there was a greater variation in *c* in field studies compared with the laboratory studies; for the reasons indicated earlier. High variation in *c* is expected because of the non-steady nature of erosion processes (particularly in the beginning of an erosion event) which was essentially captured due to the effective sampling strategy used in this study. The variation in *c* for soil R was the highest among the soils studied, which indicated that the variability in *c* could be higher for the soils which have low soil loss (Table 4.1). Such high uncertainty with the measurements of erosion could be due to the *inherent variability of soil loss* that is difficult or impossible to eliminate

(Bryan and Luk, 1981). Although current erosion models (e.g GUEST) allow interpretation of the variation in *c* within an erosion event, it is less likely that such models will interpret the variation in *c* for the period when erosion is essentially non-steady in nature (*i.e.*, in the first few minutes of erosion)

Representation of sediment concentration, runoff rate and MWD of sediment as event-average values indicated greater variation than that observed as within-event variation (Tables 8.1 to 8.3). Since this variation includes variation across the erosion treatments, these data indicate that the choice of erosion treatments was adequate and offered considerable scope for the range of studies in this thesis.

### *8.2 EROSION AND SEDIMENT QUALITY IN RELATION TO AGGREGATE STABILITY*

Aggregate stability is an indicator of soil structure, which has been used in various studies of soil erosion for the interpretation of erodibility (for example, Yoder, 1936; Wischmeier and Mannering, 1969; Bryan, 1976; Beare and Bruce, 1993; Le Bissonnais and Arrouays, 1997) Measurements of aggregate stability are also used to infer structural stability of soil as a result of changes in cropping, tillage and traffic. Most measurements of aggregate stability are based on wet-sieving with one or more sieves, when the soil has been pretreated with various methods of wetting and/ or dispersion. The method of soil pre-treatment is known to influence the size-distribution and stability of soil aggregates and therefore, the selection of a particular pre-treatment depends on the purposes of measurement (Beare and Bruce, 1993). When a set of sieves is used to fractionate soil, the MWD refers to the mean size of aggregates taking the size-distribution of soil into account. In order to relate aggregate stability with erosion, various methods of pre-treatment of soil have been used: fast and slow wetting of soil using water or alcohol, stirring, shaking, and other types of soil dispersion (e.g. Pauwels *et al.*, 1976; Glanville and Smith, 1988; Amezketa *et al.*, 1996; or Le Bissonnais and Arrouays, 1997).

The mechanical stability of aggregates presented in Chapter 4 was based on measurements of the size-distribution of soil (similar to the measurements of aggregate stability) following various pre-treatments. However, the interpretation of these measurements contrast markedly with the studies referred to above. Most of the previous studies have relied largely on the final size-distribution of soil following a particular pre-treatment. In this study, the emphasis is placed on the change in size-distribution (MWD) with a given pre-treatment of soil from the size-distribution (MWD) in a reference state (*i.e.*, when the soil is air-dry). Consideration of these changes in MWD allowed development of the indices of the susceptibility of soils to wetting ($S_W$) and dispersion ($S_D$) as quantitative measures (Table 4.4 in Chapter 4). It is expected that these and similar indices will be of practical use in ranking soils according to their degree of susceptibility to wetting from rain and irrigation, provided that the method of wetting remains unchanged. In a similar way, the susceptibility of soils to dispersion can be ranked if the method of dispersion does not change between soils. To illustrate if these indices will change with a change in the method of wetting or method of dispersion, the data on size-distribution for the three soils from Chapter 4 (Fig. 4.1) are shown together with the data for other types of wetting and dispersion method employed in ultrasonic studies of Chapter 6.

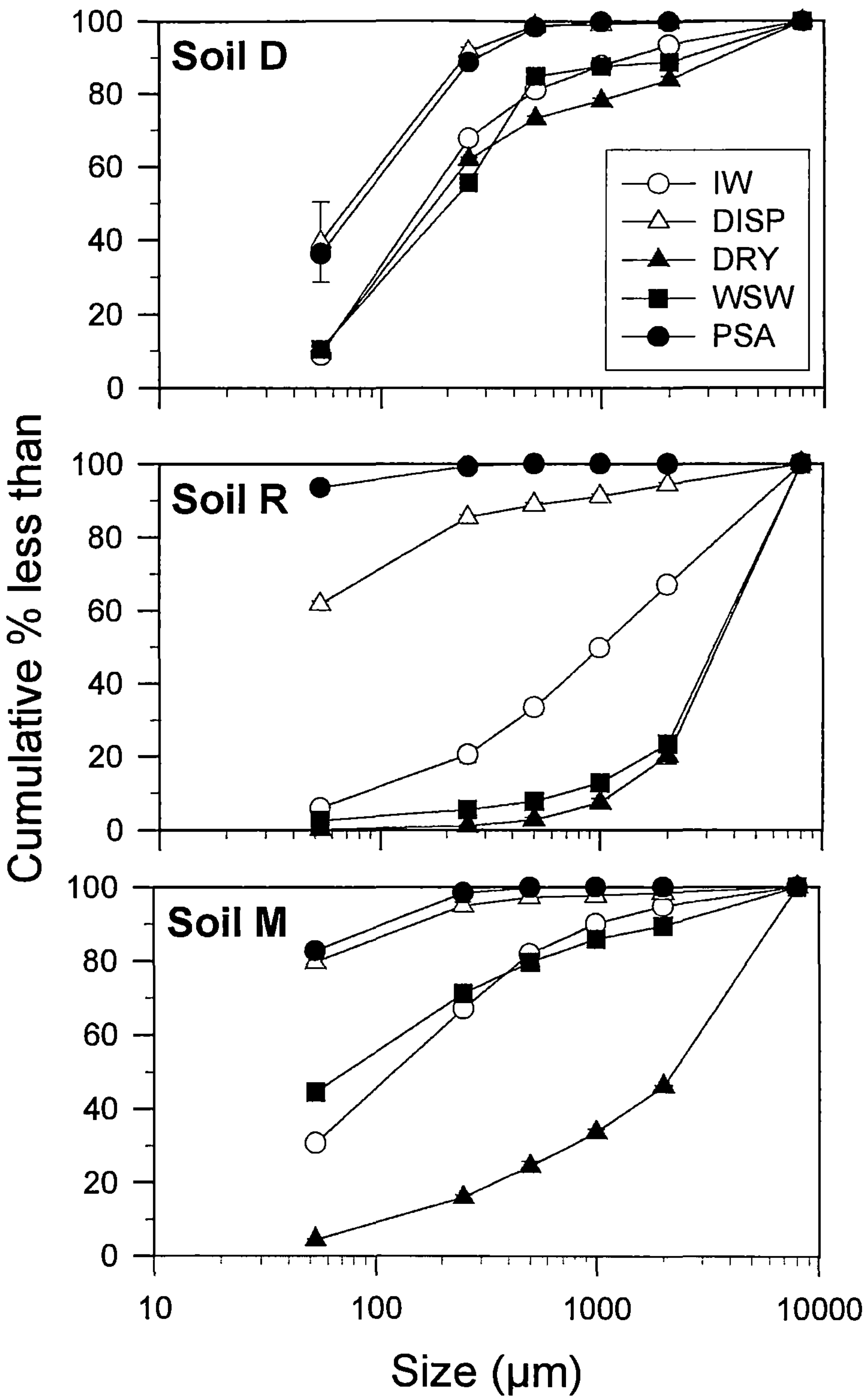


**Fig. 8.1** - Cumulative size-distribution of soils D, R, and M obtained after various methods of pre-treatment of soils (explanation in text): IW, DISP, DRY, WSW, and PSA. Standard errors of mean values are shown as vertical bars on data points.

In Fig. 8.1 the size-distribution of soil for the types of pre-treatments of soil were: (1) air-dry (DRY); (2) wet-sieving of < 53 μm fraction after immersion-wetting of whole soil sample, as used in ultrasonic dispersion study (IW); (3) wet-sieving of soil after wetting the soil with a gentle spray of water (WSW); (4) wet-sieving similar to IW, but after ultrasonic dispersion (DISP); and (5) wet-sieving similar to IW, but after mechanical and chemical dispersion of the soil, as for particle-size analysis (PSA). It should be noted that the < 53 μm fraction for IW and DISP were made up from the fractions < 2, 2-20 and 20-53 μm as reported in Chapter 7. In addition, the size-distribution of DISP was obtained after 15 minutes of sonification. The DRY, WSW, and PSA respectively refer to the dry-sieved, wet-sieved (-d) and wet-sieved (+d) treatments of soil presented previously in Fig. 4.1.

The data in Fig. 8.1 show that, for soils which are susceptible to wetting (Table 4.4) moderately (soil D) or highly (soil M), a small change in the method of wetting or dispersion did not change MWD appreciably. The soil with the most stable structure (soil R) was found to be influenced by the method of wetting (IW and WSW) and dispersion (DISP and PSA). Despite the difference between DISP and PSA treatments in Fig. 8.1 (due to inadequate dispersion of soil R in ultrasonic study, Chapter 6), it is not likely that the ranking of the three soils in terms of their susceptibility to dispersion ($S_D$ in Table 4.4) will change. It should be noted that IW is not an ideal method of wet-sieving or pre-treatment of soil as the wet aggregates of > 53 μm were subjected to a drying treatment before sieving. Therefore, IW may be considered similar to the variation in MWD due to a wetting and drying cycle. Considering the change in the size-distribution of soil after one wetting and drying cycle from its reference state (DRY), it can be seen from Fig. 8.1 that the relative mechanical stability to wetting and drying of the soils will be in the order R > D = M. Such a variation in susceptibility to wetting for soil R was consistent with the influence of wetting and drying cycle on erosion (Section 5.4.2) and erodibility (Table 5.8).

The mechanical stability of soils assessed from the rate of aggregation/ disaggregation of soils ($F_{ij}$, in Chapter 6) with the application of ultrasonic energy to soil-water suspensions was in the order of $F_{ij}$ M > D > R, *i.e.*, similar to the order of their susceptibility to wetting ($S_W$). These results imply that the breakdown of soil aggregates due to erosion may follow a similar order. The order of erodibility parameters for various soils (detachability, *a*, in Table 5.7, and specific energy of entrainment, *J*, in Table 5.9) indicated a good correspondence between $S_W$ ($F_{ij}$) and the erodibility parameters.

The variation in MWD of sediments (as an indicator of sediment quality) and of sediment concentration ( as an indicator of soil erosion) with the susceptibility of each of the three soils to wetting ($S_W$) is shown in Fig. 8.2. The data presented in Fig. 8.2 have been averaged over treatments 1 and 2 for low slope, and 3a and 3b for high slope. These data were given in a slightly different form in Tables 4.5 and 5.5. Irrespective of the slope treatments, MWD of sediment decreased and erosion increased with an increase in $S_W$. As the data in Fig. 8.2 apply to three soils only (despite that it covers a good range of $S_W$), the exact nature of the relationship is uncertain. However, it appears that beyond a certain threshold value of $S_W$ (~ 0.5), further increase in $S_W$ may not change sediment quality much. The general trend in erosion with susceptibility to wetting in Fig. 8 2 was consistent with the current interpretation of the susceptibility of soils to erosion with variation in wet-aggregate stability (as reviewed in Chapter 4). It should be noted that a reduction in MWD of sediment also reduces depositability, hence increases the opportunity for acceleration of erosion.

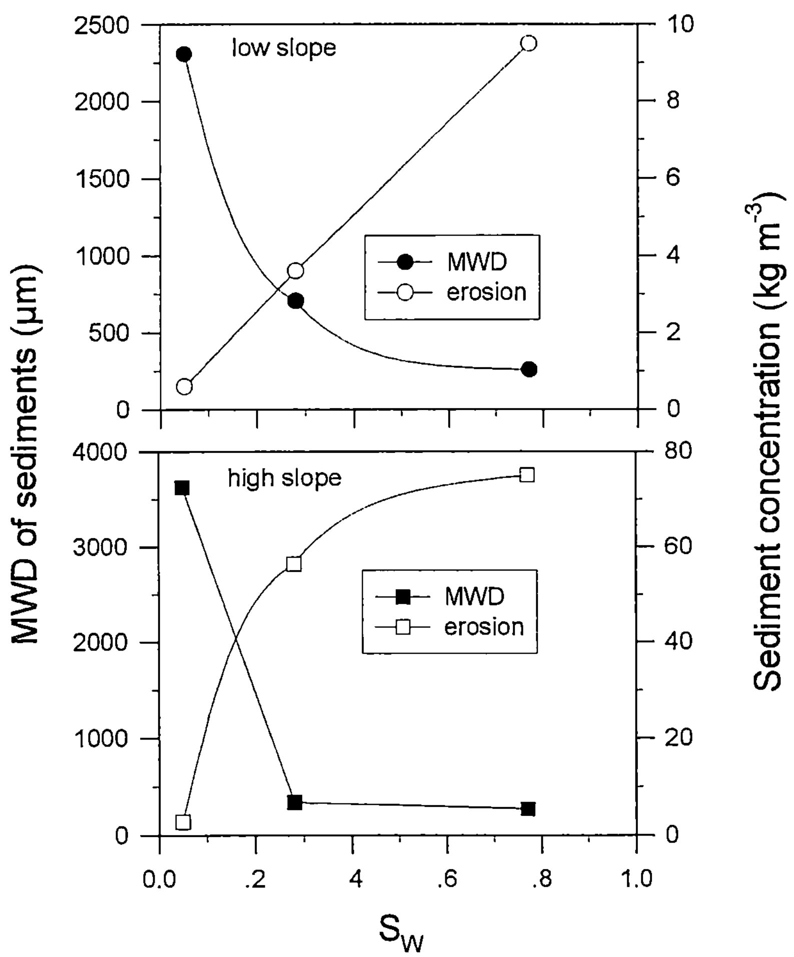


**Fig. 8.2** - Variation in MWD of sediment and sediment concentration with variation in the susceptibility of each of the three soils to wetting ($S_W$). Each data point represents an average value for a soil for low and high slope erosion treatments.

### *8.3 EROSION AND ERODIBILITY IN RELATION TO SOIL STRENGTH*

Soil strength is the ability of a soil to resist deformation under a given applied stress. The strength of the surface soil is an important factor contributing to the resistance of soils to erosion. The measurements of strength relevant to an estimate erodibility considered in this study were shear strength and penetrometer resistance, because such measurements can be made for surface soil rapidly and are inexpensive. Similar strategy have been used in various studies to relate erosion or erodibility with soil strength (Cruse and Larson, 1977; Al-Durrah and Bradford, 1982; Bradford *et al.*, 1986; Watson and Laflen, 1986; Rose *et al.*, 1990; Bradford and Huang, 1995; Misra and Rose, 1995; Sharma *et al.*, 1995; Hanson, 1996; Le Bissonnais, 1996). Although there is a consensus in the literature that erosion decreases with an increase in soil strength, a useable relationship is yet to be found to assess erodibility from the measurements of soil strength.

The relationships between sediment concentration (as an indicator of erosion) and soil strength obtained for field studies (Chapter 3) and in laboratory erosion studies (Chapter 5), agreed well (Figs. 3.3 and 5.2) with the commonly accepted inverse relationship between erosion and soil strength. However, for soil D, the sediment concentration in field studies was lower than that obtained in laboratory studies, indicating lower values for the erodibility parameter $\beta$ in the field than in the laboratory. Soil condition in the laboratory studies represented a recently tilled soil without large stones or large organic debris. Therefore, the condition of the surface soil in laboratory studies may have differed markedly from the soil in the field due to differences in structure arising from the number of wetting and drying cycles following tillage, the amount of organic and mineral debris acting as cover and the depositability of soil. The depositability of soil estimated for the data in Fig. 3.4 indicated significantly higher value of depositability for the soil in the field compared to that in the laboratory (Section 5.3.1).

The relationship between erodibility parameters (*J* and β obtained using the erosion model GUEST) and strength was uncertain for two soils (Figs. 5.3 and 5.4). The underlying theory of the model GUEST (Misra and Rose, 1989; Rose *et al.*, 1990; Hairsine and Rose, 1992a,b; Misra and Rose, 1996) suggests that with an increase in soil strength, *J* is likely to increase and β would decrease. Such a trend between *J* and soil strength was evident for one of the three soils used (soil M), when soil strength was measured with a Torvane shear device (Figs. 5.3 and 5.4). It should be noted that shear strength measured with such a device represents *apparent cohesion* (Rauws and Govers, 1988), because this device provides a measure of shear strength in the absence of any normal stress and thus, it departs from the classical Mohr-Coulomb analysis of the relationship between shear stress and normal stress used to derive cohesion. Nevertheless, from the results reported in Chapter 5, the Torvane shear device appeared to describe the variation in *J* better than the penetrometer resistance.

The difficulty in obtaining erodibility parameters for soil R and the lack of any dependence of *J* or β on soil strength for soil D (in Chapter 5) indicated that any routine measurement of soil strength (without consideration of some structural attribute) may not allow good prediction of erodibility. Therefore, a joint consideration of structure and strength of soil should be considered in future studies to interpret variation in erodibility for soils covering a wide range of structures.

### 8.4 NITROGEN LOSS

Nitrogen is one of the key nutrient elements required to maintain growth of trees in plantations. It is also the most common nutrient applied as fertiliser in plantations. In agroecosystem-based studies, it has been shown that the loss of N by erosion poses the greatest threat to decline in soil fertility (e.g. Rose and Dalal, 1988, White, 1986). For forest ecosystems and plantations, similar reports are unavailable. The loss of N due to erosion presented in Chapter 7 provides a baseline measurement for Tasmanian soils, and an upper limit for N-loss by erosion from other Australian plantations. This is because soils used for the erosion studies were temperate forest soils (hence, have a high organic-N concentration) and the simulated erosion events used in this study represented severe storm events (average rainfall rate of 116 mm $h^{-1}$) on bare soils.

The loss of N in 5 min of erosion ranged from 1-15 kg $ha^{-1}$ depending on the soil and erosion treatments used (Fig. 7.9), and was similar to the rate of N-loss reported by Palis *et al.* (1990a). Comparison of these data with that reported for uptake by *Eucalyptus* trees and N-mineralisation rate from *Tasmanian plantations* suggest that the highest value of N-loss by erosion in 5 min (15 kg $ha^{-1}$, in Fig. 7.9) represented a quantity similar to that accumulated in 10-month-old trees ($11.9 \pm 1.8$ kg N $ha^{-1}$, reported by Misra *et al.*, 1998) and close to the lower limit of the rate of net N-mineralisation in 1-2 year-old plantations in Tasmania (18-91 kg $ha^{-1}$ $yr^{-1}$, reported by Wang *et al*, 1998). At first, such comparison may seem unrealistic due to the size of erosion plots used in this study (0.8 $m^2$). However, most studies on direct measurements are from a small area (e.g. 50 mm diameter tubes used for N-mineralisation measurements by Wang *et al.*, 1998) or a few plants (6 trees for N-uptake by Misra *et al.*, 1998). These comparisons of N-loss by erosion with other sources and sinks of N in plantations, illustrate that losses of fertility due to erosion are as significant as that reported for agricultural landscapes.

The loss of N by erosion was closely related to the loss of sediment (Figs. 7.2 and 7.9), and therefore, any effect of structure on soil erosion carried through to N-loss. There was no specific advantage in estimation of N-loss via the enrichment ratio of sediment ($E_R$) as suggested in previous studies (Menzel, 1980). Due to the lack of any consistent pattern in the variation of the ratio of N-concentration of sediment to that in the uneroded soil with size for any soil (Table 7.2), the mechanism to explain the enrichment of sediment was uncertain. It is suggested that N-loss can be predicted directly from sediment loss assuming a constant N-concentration of sediment, because of (1) the uncertainty in interpretation of $E_R$ in this study, (2) a strong dependence of N-loss on sediment loss in a range of erosion studies (Fig. 7.13), and (3) an improvement in prediction of N-loss from the measured sediment loss, despite poor prediction of $E_R$ by Menzel's equation (Fig. 7.12). Thus, it is important that sediment loss is measured or predicted accurately to obtain good estimate of N-loss by erosion. Attempts to predict sediment loss with β (from Chapter 5) for the duration for which measured values of N-loss was available (for 0-5 and 35-40 min of erosion, in Chapter 7) were met with limited success due to the range of variation in sediment concentration observed with time of erosion for various events (Fig. 5.1), and the inability of the erodibility parameter β to represent short-term sediment flux.

## *8.5 CONCLUSIONS*

1. A quantitative indicator of the susceptibility of soils to wetting ($S_W$) developed in this study could describe variation in erosion and sediment characteristics for three forest soils. Such an indicator will be useful in future studies to discriminate the susceptibility of various soils to erosion. As average size of sediment decreased with an increase in $S_W$, this indicator can be useful to compare various soils for the fineness of sediment expected during erosion.

2. Erosion was influenced by soil structure and strength. For strongly aggregated soils with high depositability, erosion was low. When soils of contrasting structures were considered together, there was no correspondence between erodibility parameters and soil strength. As soil strength was influenced by soil structure, it is recommended that some quantitative attribute of structure need to be considered with strength to improve the correspondence between erodibility and strength.

3. For the soils studied, variation of nitrogen loss due to erosion was mostly due to variation in sediment loss, rather than due to the variation in concentration of N in sediment. Therefore, it is suggested that improvement in the accuracy of erosion prediction is of greater importance than the prediction of N-concentration of sediment with or without considering the nitrogen-enrichment of sediment.

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# APPENDIX

## *DEFINITION OF EROSION PROCESSES AND A GENERAL DESCRIPTION OF THE MODEL GUEST*

### *Definition of erosion processes*

Soil erosion is mainly caused by detachment of sediment from the soil surface due to the impact of raindrops, and the lift and transportation of sediment by overland flow. Soil that had not been eroded is referred to as uneroded soil. Soil material that has been eroded is referred to as sediment.

The process by which the impact of rainfall on the soil surface removes material from the bed of uneroded soil is termed *rainfall detachment*. The high local stresses produced by raindrop impact may break soil aggregates and detach material from the soil surface. Detached sediment may deposit back on the soil surface. *Re-detachment* is the detachment of deposited sediment. Less energy is required to re-detach a certain amount of deposited sediment than to detach the same amount of uneroded soil, because deposited sediment has a lower cohesive strength compared to the uneroded soil.

The processes by which overland flow (or runoff) removes sediment from the bed of uneroded soil and from the deposited layer are termed as *entrainment* and *re-*

*entrainment*, respectively. The ability of overland flow to entrain sediment is directly related to flow velocity, and hence its shear stress. Entrainment of soil occurs when the work done by overland flow exceeds the cohesive strength of the soil.

*Deposition* is the process of settling of eroded material in runoff water under the action of gravity. It is a size-selective process due to the dependence of settling velocity on size, shape, and density of sediment While detachment, entrainment, re-detachment, and re-entrainment are erosion processes that increase sediment concentration of runoff water, the process of deposition decreases it. The sediment concentration in an erosion event is determined by the net effect of detachment, re-detachment, entrainment, re-entrainment, and deposition.

### General description of the model GUEST

The model GUEST is a steady-state model, and the mass conservation of sediment of settling velocity of class *i* for a plane with uniform rainfall and runoff requires that (Misra and Rose, 1989):

$$\frac{\mathrm{d}q_{si}}{\mathrm{d}x} = \mathrm{e}_i + \mathrm{e}_{di} + \mathrm{r}_{ei} + \mathrm{r}_{ri} - \mathrm{d}_i \quad , \qquad \text{(A.1)}$$

where $q_{si}$ (kg m$^{-1}$ s$^{-1}$) is the sediment flux, which is the product of sediment concentration of settling velocity of class *i* ($c_i$) and volumetric flux of water ($q$), and $x$ (m) is the distance downslope of a bare plot of known area. The terms $\mathrm{e}_i$, $\mathrm{e}_{di}$, $\mathrm{r}_{ei}$, $\mathrm{r}_{ri}$, and $\mathrm{d}_i$ are the rates (kg m$^{-2}$ s$^{-1}$) of, respectively, rainfall detachment, re-detachment, runoff entrainment, re-entrainment, and deposition of sediment of settling velocity of class *i*. GUEST assumes that entrainment occurs only when stream power ($\Omega$, W m$^{-2}$) exceeds a certain threshold stream power ($\Omega_0$). Stream power ($\Omega$) is the energy associated with runoff water and is estimated from the shear stress associated with runoff water ($\tau$, Pa) and the velocity of flow (V, m $s^{-1}$). When $\Omega < \Omega_0$, the terms representing entrainment ($r_{et}$) and re-entrainment ($r_{rt}$) are absent in equation (A.1); therefore, the simplified equation describes sediment flux due to rainfall-driven erosion processes and deposition. GUEST assumes that sediment concentration reaches an equilibrium during an erosion event when the mass of the deposited layer does not change with time. In this case, the net effect of the erosion processes that influence the mass of deposited layer is zero, thus:

$$d_t - e_{dt} = 0 \tag{A.2}$$

When $\Omega > \Omega_0$, erosion depends on rainfall as well as runoff-driven processes, and deposition. Therefore, with no change in mass of the deposited layer with time, the sediment concentration at equilibrium is reached due to:

$$d_t - e_{dt} - r_{rt} = 0 \tag{A.3}$$

Erosion studies in flumes with simulated rainfall at constant rate indicate that intense rilling can lead to a situation where the active erosion surface is completely covered by a cohesionless deposited layer. Sediment concentration in that situation reaches a maximum that can be attained for a known flow regime. In this situation, the maximum sediment concentration is referred to as the *transport limit* ($c_t$), and is the net result of re-detachment, re-entrainment, and deposition processes. At transport limit, only a fraction F of the excess stream power ($\Omega$ - $\Omega_0$) is effective in the re-entrainment of sediment. Evaluation of F has been difficult and relies mostly on flume experiments (Proffitt *et al.*, 1993). An estimate of $c_t$ in the absence of rills can be obtained as (Misra and Rose, 1996):

$$c_t = \frac{1}{\phi}\left[a_d\mathrm{P} + \frac{\mathrm{F}\sigma(\Omega - \Omega_0)}{g\mathrm{D_w}(\sigma - \rho)}\right], \tag{A.4}$$

and in the presence of rills from:

$$c_t = \frac{1}{f\phi}\left[a_d\mathrm{P} + \frac{\mathrm{R_1F}\sigma(\Omega - \Omega_0)}{g\mathrm{D_w}(\sigma - \rho)}\right], \tag{A.5}$$

where $\phi$ (m s$^{-1}$) is the mean setting velocity of uneroded soil also referred to as *depositability*, $a_d$ (kg m$^{-3}$) is the rainfall re-detachability, P (m s$^{-1}$) is the rainfall rate, $g$ (m s$^{-2}$) is the acceleration due to gravity, $D_w$ (m) is the depth of water on the soil's surface, $\sigma$ (kg m$^{-3}$) is the wet-density of sediment, $\rho$ (kg m$^{-3}$) is the density of runoff water, $f$ is a dimensionless factor that indicates the efficiency with which re-entrained sediment can slide down the walls of a rill ($f$ = 1 for rectangular rills), and $R_1$ is the ratio of the width of the deposited layer to the wetted perimeter in a rill.

In general $c < c_t$, since the complete coverage of the soil surface by a deposited layer cannot usually be sustained for a long period of time. When only a fraction of the soil surface is covered by a deposited layer, the effective stream power, $F(\Omega-\Omega_0)$, is used in entrainment of the cohesive uneroded soil and re-entrainment of the cohesionless deposited material. In this situation, the sediment concentration is the net result of detachment, re-detachment, entrainment, re-entrainment, and deposition processes; and it is influenced by the strength of the uneroded soil. The sediment concentration, in this situation, is considered to be at *source limit*. The sediment concentration at the source limit relates to a parameter $J$ (referred to as the specific energy of entrainment, expressed in J kg$^{-1}$), that is implicitly related to the cohesive strength of uneroded soil (Hairsine and Rose, 1992a). There is some indication that $J$ increases with an increase in soil strength (Misra and Rose, 1995).

### ***Erodibility parameters of the model GUEST***

Variation in sediment concentration with time in an erosion event is interpreted by GUEST to yield three erodibility parameters. These are:

- Rainfall detachability ($a$, kg m$^{-3}$)

- Rainfall re-detachability ($a_d$, kg m$^{-3}$)

- Specific energy of entrainment ($J$, J kg$^{-1}$)

The first pair of parameters ($a$ and $a_d$) can be estimated from the measured data in situations where $\Omega \leq \Omega_0$, as these parameters relate to rainfall-driven erosion processes only. The parameter $J$ is obtained when both rainfall- and runoff-driven erosion processes occur (i.e, when $\Omega > \Omega_0$). The depositability of the uneroded soil is required to obtain the estimates of these erodibility parameters. Depositability is determined from the measurement of settling velocity characteristic of the uneroded soil with a Modified Bottom Withdrawal Tube technique (Lovel and Rose, 1986), or from sieving of wet uneroded soil (Lisle *et al.*, 1995).

When sediment concentration can not be measured as a function of time (as it is difficult in field studies), an empirical erodibility parameter (β) can be obtained with the model GUEST from the average sediment concentration ($\bar{c}$, kg m$^{-3}$):

$$\bar{c} = \overline{c_t}^{\,\beta} \quad , \tag{A.6}$$

where $\overline{c_t}$ is the sediment concentration at transport limit due to runoff -driven erosion processes only. The expression for $\overline{c_t}$ is similar to the equations (A.4) and (A.5), but it does not include the first term in the brackets of the RHS (*i.e.* $a_d$P).

Fig A.1 illustrates the typical relationship between $c$ and stream power ($\Omega$) for a range of $J$ and $\beta$. Note that $\Omega$ combines the effects of slope, slopelength and runoff rate on sediment concentration. In Fig. A.1 the contribution of rainfall detachment for sediment concentration has been neglected. This figure shows that $\bar{c}$ is approximately proportional to $J$.

The erodibility parameters of GUEST model have been reported for a number of studies on agricultural soils using field runoff plots and flumes, and simulated and natural rainfall (for example, Ciesiolka *et al.*, 1995; Proffitt *et al.*, 1991; Rose *et al.*, 1993; Misra and Rose, 1995; Hashim *et al.*, 1995; Paningbatan *et al.*, 1995; Fentie *et al*, 1997). However, there are no previous reports on erodibility parameters for forest soils. In Chapter 5 of this thesis, estimates of erodibility parameters are made, and their variation with soil properties, including soil strength, are examined.

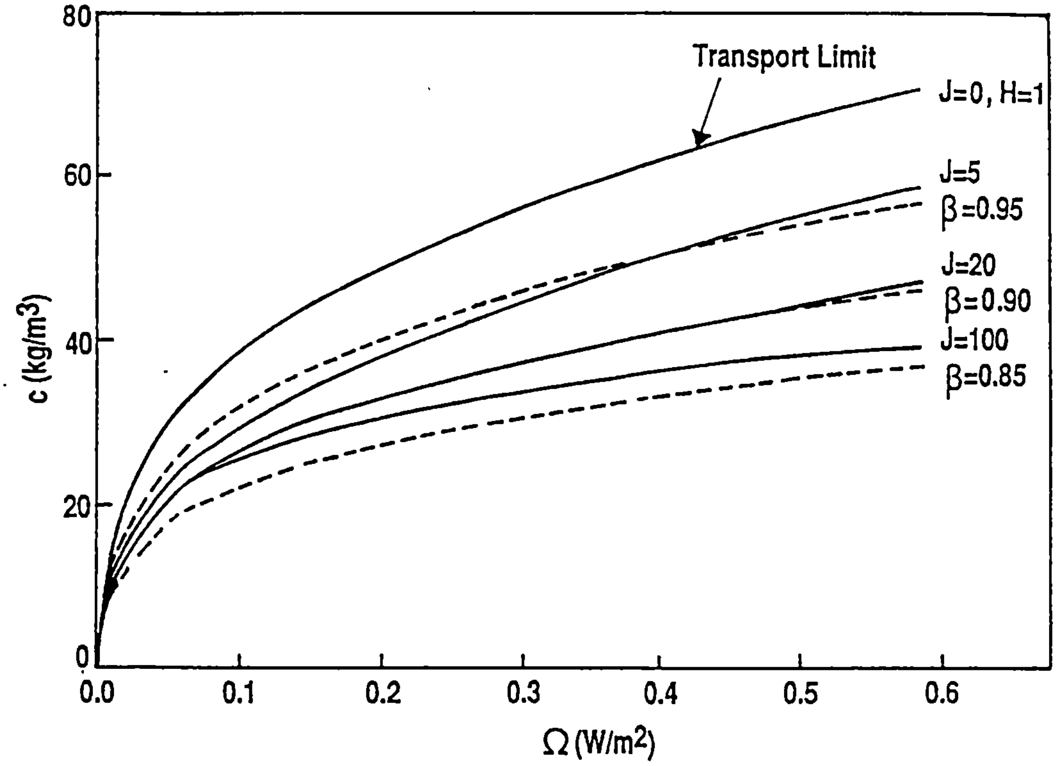


**Fig. A.1** - Typical variation of sediment concentration ($c$) or average sediment concentration ($\bar{c}$) with stream power ($\Omega$) for a range of values of parameters $J$ (J kg$^{-1}$) and $\beta$. The effects of rainfall-driven erosion processes on $c$ and $\bar{c}$ are neglected (adapted from Misra and Rose, 1989).