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Correlation of physicochemical properties and sub-erosional landforms with aggregate stability variations in a tropical Ultisol disturbed by forestry operations

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Abstract

The stability of soil aggregates against water erosion is a largely unstudied factor in research on the spatially-distributed impacts of tropical forestry. Soil strata with particularly weak aggregates are likely to be the focus for erosional activity whether buried or exposed by either natural processes or forestry activities. Understanding the location and likely cause of such intrinsic instability would, therefore, allow better spatially-distributed parameterisation of erosion models. This study has aimed to identify the range in aggregate stability within a region of Bornean Ultisol disturbed by forestry operations, and to identify the bulk soil properties associated with such stability variations. The sites sampled were subject to a range of denudational processes including piping, rilling and landslide-triggered erosion. Soil profiles with rates of erosion in excess of 10 mm a^{-1} were shown to have less aggregate stability (as characterised by the rainfall simulation survival index (RSSI)) than those with no visible signs of erosion. Further, large differences in aggregate stability between soil horizons of the same profile were observed and seen to be statistically correlated with the bulk soil properties of organic carbon and clay content (i.e. stabilising agents) and the dispersing agent of exchangeable sodium percentage (ESP) at sites undergoing erosion. Organic carbon appeared to be the most important governing factor, accounting for 56% of the variance in the aggregate stability. The presence of strongly and weakly expanding 2 : 1 clays within the soils may have been an additional de-stabilising factor. Reduced porosity was also observed at eroding sites with weak soil aggregates. The properties of EC_{25} and sodium adsorption ratio (SAR) were not correlated with the RSSI or the presence of soil erosion. Identification of the role of organic carbon, clay and ESP in the stability of these tropical forest aggregates is important in focusing future, more intensive research on the spatial parameterisation of models to simulate forestry impacts on erosion.

Keywords: Aggregate stability; Erosion modelling; Tropical forestry; Ultisol

1. Introduction

Aggregate stability can be a key indicator of the resistance of tropical soils to erosion (Bryan, 1968; De Vleeschauwer et al., 1978; Bryan et al., 1989).

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Identification of the critical soil properties affecting aggregate stability is, therefore, central to understanding where erosion might be located within a catchment as a result of natural or forestry-modified patterns in aggregate stability (de Roo, 1993). Bulk soil properties that have been shown to affect aggregate stability include the soil texture, organic content, clay mineralogy, and the presence of chemical dispersing agents. Lal (1981) and Bryan et al. (1989) have, however, stated that the relative importance of each of these properties to aggregate stability varies with the soil type and also with the type of sub-erosional process, that is, surficial-, gully-, piping- or landslide-triggered erosion (terminology from Bryan et al., 1989). While some soil properties affect aggregate stability, others such as the hydrological properties of porosity and saturated hydraulic conductivity are themselves affected by collapsing soil aggregates (Nys and Ranger, 1985; Baumgartl and Horn, 1991).

While there is wealth of research on aggregate stability of temperate agricultural soils, there is a dearth for tropical forest soils (Lal, 1980; Rose, 1993). This is important as the spatial variability of the stability of soil aggregates against erosion may be a key factor controlling sediment sources within tropical forest terrain (de Roo, 1993). The current high levels of disturbance of tropical forests has, in contrast to the limited work on aggregate stability, stimulated extensive research on the impacts on sediment mobilisation at the scale of experimental catchments (Bruijnzeel, 1990, 1996; Douglas, 1992; Douglas et al., 1992). Southeast Asia has been a particular focus for this work, and the most common soil within this region affected by such anthropogenic activity is the USDA Ultisol order. The Ultisol order is equivalent to the Acrisol–Alisol groups of the current FAO classification (FAO–Unesco, 1990; Landon, 1991). Indeed, these soils are the second most common soils of the tropics as a whole (Driessen and Dudal, 1991) and also cover extensive areas of southeast USA and southeast China. Ultisols are defined primarily on the basis of the clay accumulation (i.e. argillation), low base supply and non-perudic (i.e. wet throughout the year) moisture regime within the subsoil (Soil Survey Staff, 1975).

This exploratory study focuses on a region of southeast Asia comprising of an Ultisol beneath tropical forest disturbed by commercial, selective logging. The work has two aims: first to identify the spatial variability in aggregate stability observed within soil profiles experiencing soil loss by surficial-, gully-, piping- or landslide-triggered erosion. Secondly, the study aims to identify the bulk soil properties that may give rise to instability of soil aggregates comprising this tropical forest Ultisol. This preliminary work seeks to complement the plot and catchment scale erosion studies undertaken at the same site by for example, Douglas et al. (1992), Brooks et al. (1994), Walsh and Bidin (1995) and Chappell et al. (1998). Ultimately, understanding the spatial variability in aggregate stability and its causal factors will allow better spatial parameterisation of catchment-scale soil erosion models (de Roo, 1993).

2. Study region

The ca. 10 km^2 study region lies to the north of the research station – Pusat Luar Lembah Danum (or 'DVFC') in Sabah, Malaysian Borneo (Fig. 1). The climate at DVFC is equatorial with relatively little seasonality and an 11-year (1986–1996) mean rainfall of 2778 mm ($\pm\ \sigma = 320$). High rainfall intensities are relatively frequent with >50 mm h^{-1} (max. 5 min intensity) having a return-period of 23.3 days and in excess of 100 mm h^{-1} a return-period of 139.6 days (Sherlock, 1997).

According to Wright (1975) and FAO–Unesco (1979) Ultisol soils have developed over the whole of our 10 km^2 study region. All Ultisol soils sampled within this study were developed on sandstone members of the Kuamut geological formation, which also includes mudstone, tuff and chert (Leong, 1974; Clennell, 1996). The topography is undulating, and has an altitudinal range of 130–260 m (Fig. 1).

The research area is covered by 'lowland dipterocarp rain forest' managed by the Forestry Upstream Division of Yayasan Sabah as part of their 9728 km^2 logging concession (Marsh and Greer, 1992). The forest to the west of the River Segama (Fig. 1) does, however, form part of the Danum Valley Conservation Area, a 438 km^2 tract of undisturbed, Class I (Protection) Forest. The study area to the northeast of DVFC (largely the catchment of the Sapat Kalisun river) was, in contrast, selectively logged between 1988 and 1990. Timber extraction was by a 'selective' rather than a

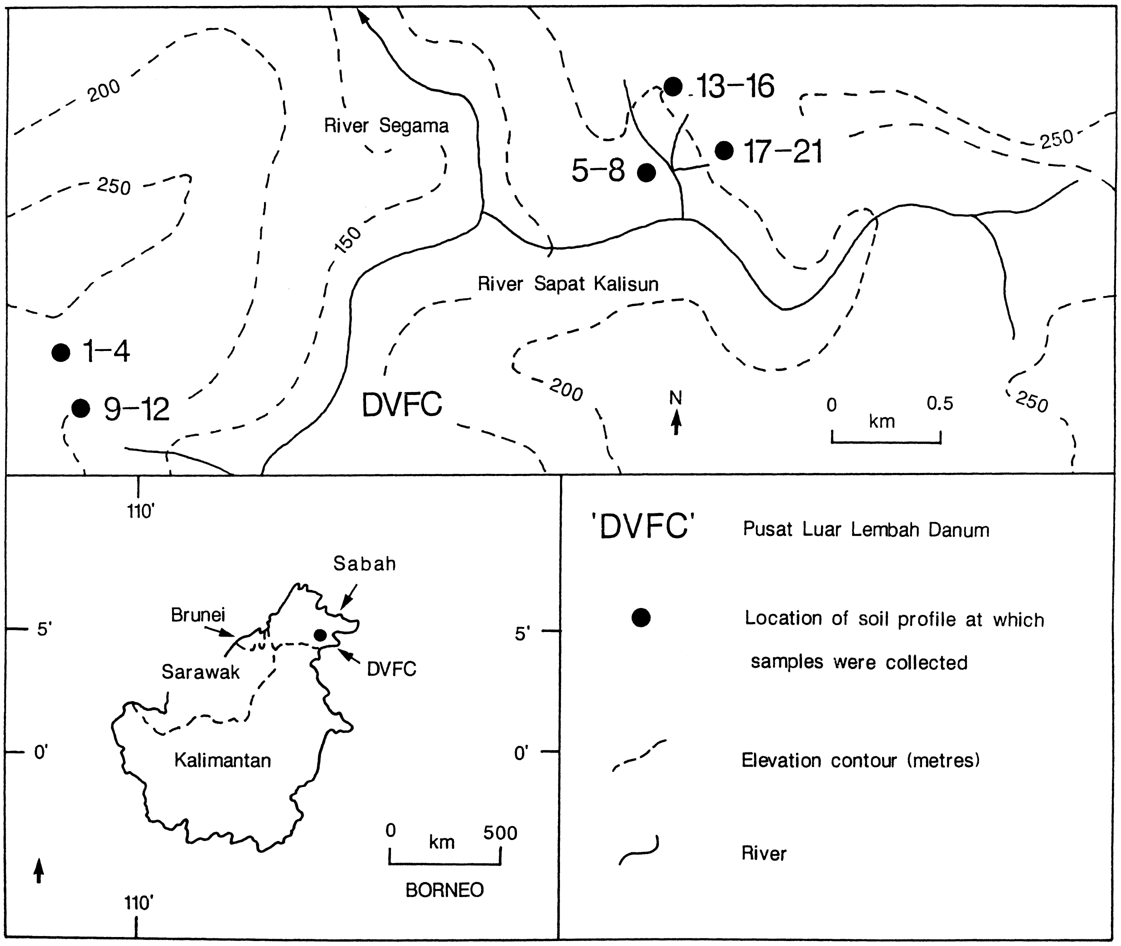

Fig. 1. The 10 km^2 sampling region north of Pusat Luar Lembah Danum ('DVFC'), Sabah, Malaysian Borneo. The map shows the sampling sites for the aggregate stability and bulk soil properties.

'clearfell' logging system, and has the aim of producing a sustainable harvesting coupe every 30 years. This type of logging system leaves a dense forest cover that is typical of selectively logged forests throughout southeast Asia (Marsh and Greer, 1992).

A total of 21 sampling locations within the 10 km^2 region (Fig. 1) were selected to encompass the likely extremes in aggregate stability of an Ultisol beneath disturbed tropical forest. Although small, this preliminary sampling frame is comparable to that used for correlation analyses by Morgan (1979), Chaney and Swift (1984), Goldberg et al. (1988), Levy et al. (1993b) and Levy and Miller (1997).

Sample Sites 1–8 have no visible erosion, while the Sites 9–21 have measured rates of erosion in excess of 10 mm a^{-1} (Table 1). Similarly high rates of erosion have been monitored on Ultisols (disturbed and undisturbed) elsewhere in south-east Asia (Bruijnzeel, 1990; Douglas, 1992). Furthermore, soil samples 1–12 were extracted from beneath undisturbed forest blocks, while soil samples 13–21 were associated with logging related terrain modifications.

Within the undisturbed forest, samples 1–8 were taken from two 20° linear slope segments, while samples 9–12 were taken from the head of an ephemeral channel or natural headwater gully. The head of this ephemeral channel contained numerous natural soil pipes, which can be observed at the head of many other stream channels in the area (Bidin, 1995; Sherlock, 1997). Indeed, there is a high density of ephem-

Table 1
Erosion rates at the soil sample sites

Soil samples	Landform	Rate of soil loss	Sub-erosional process (after Bryan et al., 1989)	Observation or measurement technique
1–4	20° planar slope	No signs of erosion	None	Visual only
5–8	20° planar slope	No signs of erosion	None	Visual only
9–12	Channel head	30 mm a^{-1}	Piping erosion	Erosion pins[a]
13–16	Skid trail (5°)	12 mm a^{-1}	Surficial erosion	Erosion bridges[a]
17–21	Landslide backwall	250 mm a^{-1}	Landslide-triggered erosion	Erosion pins[a]

[a] Walsh and Bidin, 1995.

eral channels in the region, which when incorporated with the perennial stream channels give a very high drainage density of 20 km km^{-2} (Walsh and Bidin, 1995).

Within the logged forest, soil samples 13–16 were extracted from beneath a trail used by a tracked forestry vehicle, known as a ‘skidder’, in 1989. The so called ‘skid trail’ is 40 m in length and the sample profile was excavated close to the junction with the road used by timber haulage lorries. Samples 17–21 were extracted from the arcuate backwall of a landslide that had occurred on 14/12/1994. This landslide, a complex slump-earthflow, similar to those described by Prior and Ho (1972), is approximately 0.3 ha in plan and is developed within roadfill material below a secondary haul road constructed in late 1988 and subsequently used for lorry haulage of timber in 1989. Qualitative evidence gained in mid-1994 indicated that a smaller scale landslide had taken place at this site prior to the failure in late 1994. Given the size of the secondary trees (*Macaranga* spp.) that had re-established on the landslide, the earlier failure probably occurred during the road construction or logging phase. On 19/1/1996, a 170 mm rainstorm triggered more extensive and larger-scale landsliding within the study area. Following this landsliding event very extensive rill erosion could be observed within the mobilised materials at the sampled landslide (samples 17–21; Table 1) and at other landslides within the region.

Soil pits were excavated to allow the extraction of the 21 soil samples. At each sampling location approximately 2 kg of soil was sampled from the same horizon. These samples were air dried and taken to the UK for subsequent analysis for aggregate stability and related physicochemical properties. Two intact soil cores were also extracted next to the 2 kg bulk samples for local analysis of porosity and soil moisture content.

3. Aggregate stability determination

Aggregate stability was determined on the samples using the standardised rainfall simulation techniques described by Ternan et al. (1996). This method involves testing soil aggregates against their response to simulated rainfall using a laboratory rainfall simulator inside an environmental chamber. Twenty five air-dried aggregates from each sample location were placed on a 2.8 mm sieve and equilibrated for 24 h at 50% relative humidity to ensure similar initial soil moisture status. Rainfall was simulated at ca. 45 mm h^{-1} intensity with a mean drop size of 580 μm ($\pm\,\sigma = 250$) as determined by the filter-paper technique adopted by Mason and Andrews (1960). The number of water-stable aggregates surviving at 0.5 min time intervals (a_t) over a period of 20 min was first determined. These data were then corrected for aggregates found to comprise of stones >2.8 mm using:

$$A_t = 100\left[\frac{(a_t - \sum s)}{(\sum a - \sum s)}\right] \tag{1}$$

where A_t is the per cent of aggregates surviving at time t (minutes) after correction for the number of aggregates containing stones >2.8 mm, Σs is the number of aggregates containing stones observed at the end of the whole test (see Table 2), and Σa is the number of aggregates tested (always 25). These corrected data were plotted as aggregate breakdown curves (Fig. 2) and the RSSI calculated from:

$$\mathrm{RSSI} = \frac{\sum(A_5 + A_{10} + A_{15} + A_{20})}{4} \tag{2}$$

Table 2
Physical properties of the sampled soil profiles, DVFC, Sabah, Malaysian Borneo

Sample	Profile	Sampling depth (m)	Σs	RSSI (%)	Stones (%)	Sand (%)	Silt (%)	Clay (%)	Texture class	OC (%)	η $(\mathrm{m}^3\,\mathrm{m}^{-3})$	s	θ% $(\mathrm{m}^3\,\mathrm{m}^{-3})$
1	1	0	0	100	0.2	36.1	48.4	15.5	L	1.53	0.585	2	67
2		0.25	2	100	4.9	31.4	50.8	17.8	SiL	0.10	0.505	2	63
3		0.51	1	100	30.2	39.5	40.5	20.0	L	0.28	0.459	2	63
4		0.88	2	94	28.7	45.5	41.3	13.2	L	0.37	0.512	2	64
5	2	0	1	100	0.2	36.6	50.1	13.4	SiL	1.22	0.519	2	66
6		0.25	0	100	0.1	25.4	55.3	19.4	SiL	0.49	0.547	2	84
7		0.50	0	100	0.1	24.6	55.0	20.5	SiL	0.41	0.566	3	57
8		0.92	0	100	0.1	21.5	55.7	22.8	SiL	0.35	0.520	3	77
9	3	0	0	100	0	27.6	56.6	15.8	SiL	1.45	0.605	2	55
10		0.25	0	100	0	25.9	55.4	18.3	SiL	0.73	0.515	2	62
11		0.65	0	77	0.2	26.5	55.2	18.4	SiL	0.51	0.541	2	68
12		0.94	0	60	0.5	21.8	60.9	17.3	SiL	0.48	0.523	2	61
13	4	0	0	87	3.1	25.4	58.1	16.5	SiL	0.77	0.514	4	60
14		0.25	0	100	0.1	11.4	63.8	24.9	SiL	0.40	0.557	4	65
15		0.51	2	86	0.6	10.1	66.4	23.5	SiL	0.39	0.574	4	70
16		0.71	1	92	14.1	12.8	67.2	20.0	SiL	0.38	0.557	3	68
17	5	0	3	81	24.8	47.4	38.9	13.8	L	0.18	0.385	3	68
18		0.55	3	45	18.5	35.1	51.0	13.9	L	0.11	0.427	2	82
19		0.96	5	36	25.5	53.0	33.7	13.3	SaL	0.08	0.399	2	73
20		1.25	1	92	0.4	30.9	49.4	19.7	L	0.28	0.504	3	93
21		2.20	0	93	0.4	34.7	48.5	16.8	L	0.41	0.501	3	100

'Σs' is the number of test aggregates containing stones; 'stones (%)' is the proportion of the bulk soil greater then 2 mm; 'sand (%)', 'silt (%)', and 'clay (%)' are the proprotions in the fine-earth-fraction.
The texture class 'L' is loam, 'SiL' is silt loam, and 'SaL' is sandy loam'.
Θ% is the degree of volumetric wetness relative to the porosity.
(η);s is the USDA soil structural class.

where A_5, A_{10}, A_{15} and A_{20} are the corrected percentages of aggregates surviving at 5, 10, 15 and 20 min respectively (Ternan et al., 1996).

4. Determination of factors potentially correlated with aggregate stability

The stability of soil aggregates against simulated rainfall is a function of the interaction between several soil properties. Development of a model to predict aggregate stability from a combination of bulk soil properties would demonstrate the likely sensitivity of aggregate stability (within a specific region) to particular properties and, therefore, allow natural or anthropogenically-induced spatial variability in aggregate stability to be better understood. The bulk soil properties that the previous studies have indicated are correlated with aggregate stability include (a) soil texture, (b) organic matter or organic carbon, (c) chemical dispersing agents of electrical conductivity, water-soluble and exchangeable sodium, SAR, and ESP, (d) clay mineralogy, and (e) porosity. These properties were, therefore, assessed at the same 21 sample sites as the aggregate stability.

For this study, soil texture was determined by wet sieving the greater than 0.06 mm fraction and laser sizing, using a Malvern Mastersizer 5002, the less than 0.06 mm fraction. Organic carbon content (OC) was measured by high temperature catalytic oxidation with non-dispersive infra-red detection (Sugimura and Suzuki, 1988). A Shimadzu TOC5000 was used for this analysis.

The bulk soil electrical conductivity (EC_{25}) was measured on 1 : 2 saturation extracts using a

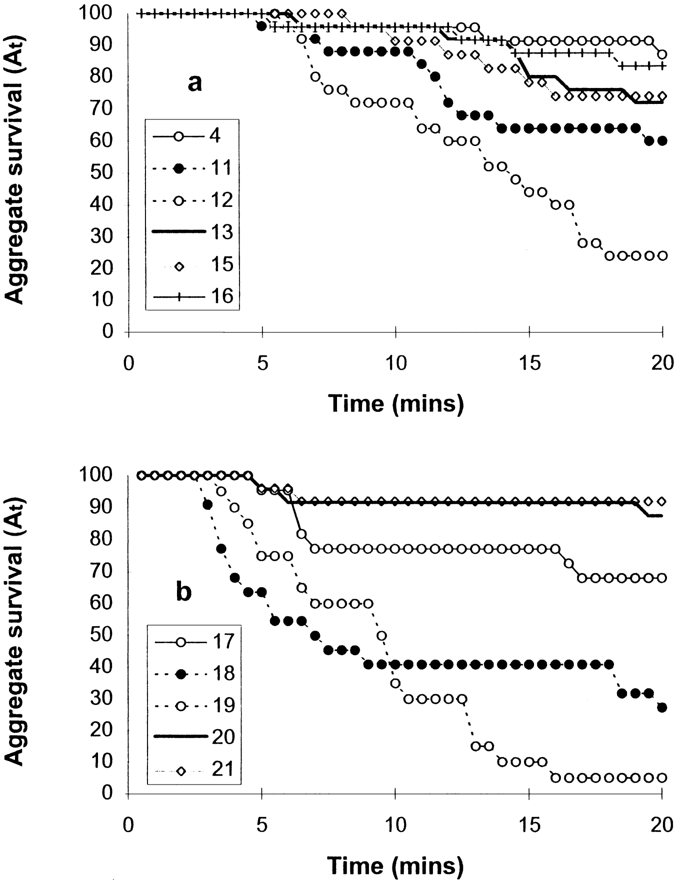

Fig. 2. Aggregate breakdown curves from the rainfall simulation tests (a) for samples 4, 11–13, 15–16, and (b) for samples 17–21. Samples 1–3, 5–10, and 14 are not illustrated as they have 100% aggregate stability.

four-electrode conductivity meter. Exchangeable sodium was determined using ammonium acetate extraction followed by analysis with flame photometry. The indices of the SAR and ESP were also determined. They are defined by:

$$\mathrm{SAR} = \frac{\mathrm{Na}^+}{\sqrt{(\mathrm{Ca}^{2+} + \mathrm{Mg}^{2+}/2)}} \tag{3}$$

$$\mathrm{ESP} = 100\left(\frac{\mathrm{Na}_x}{\mathrm{CEC}}\right) \tag{4}$$

where Na^+, Ca^{2+}, Mg^{2+} are the ionic concentrations of sodium, calcium and magnesium measured in 1 : 2 saturation extracts (mmol l^{-1}), Na_x is the exchangeable sodium (cmol kg^{-1}), and CEC is the cation exchange capacity of the soil (cmol kg^{-1}). The ionic concentrations of sodium, calcium and magnesium were determined by the analysis of sodium with flame photometry and atomic absorption spectrophotometry (AAS) for calcium and magnesium. The CEC was determined by the summation method, a technique considered by Rhodes (1982) to be more reliable than the more common ammonium saturation method. This approach, however, necessitated the determination of exchangeable sodium, potassium, calcium, magnesium, hydrogen and aluminium. Exchangeable potassium, calcium and magnesium were determined using ammonium acetate extraction followed by analysis of potassium with flame photometry and calcium and magnesium with AAS. The exchangeable hydrogen and aluminium were determined using 1M potassium chloride extraction followed by analysis with a pH electrode and flame photometry, respectively.

Table 3
Ionic concentrations of the sampled soil profiles, DVFC, Sabah, Malaysian Borneo

Sample	Profile	Sampling depth (m)	Exchangeable cations (cmol kg^{-1})						Water soluble salts: saturation extract (mmol l^{-1})			
			Na_x	Ca_x	Mg_x	K_x	Al_x	H_x	Na^+	Ca^{2+}	Mg^{2+}	K^+
1	1	0	0.220	5.028	1.208	0.164	11.634	5.956	0.335	0.142	0.310	0.192
2		0.25	0.601	0.197	0.075	0.087	4.895	0.652	0.235	0.049	0.012	0.033
3		0.51	0.257	1.409	0.148	0.130	7.712	4.539	0.135	0.006	0.004	0.033
4		0.88	0.226	0.696	0.072	0.288	15.480	0.875	0.274	0.010	0.019	0.036
5	2	0	0.157	0.622	0.223	0.127	9.833	4.070	0.652	0.077	0.052	0.097
6		0.25	0.275	0.341	0.209	0.257	12.516	5.345	0.283	0.035	0.021	0.028
7		0.50	0.343	1.239	0.422	0.321	14.000	1.400	0.209	0.056	0.016	0.041
8		0.92	0.377	1.153	0.231	0.277	13.907	1.134	0.265	0.039	0.004	0.018
9	3	0	0.407	0.346	0.363	0.210	6.742	1.774	0.387	0.089	0.014	0.256
10		0.25	0.346	1.134	0.419	0.210	8.541	1.399	0.170	0.023	0.007	0.069
11		0.65	0.307	1.357	0.433	0.222	7.667	1.928	0.226	0.011	0.010	0.061
12		0.94	0.875	0.482	0.497	0.246	7.761	2.058	0.139	0.006	0.006	0.031
13	4	0	0.297	1.903	0.373	0.578	11.275	3.443	0.165	0.031	0.009	0.051
14		0.25	0.398	1.428	0.160	0.233	15.202	2.960	0.135	0.005	0.002	0.010
15		0.51	0.197	0.142	0.053	0.236	14.216	2.346	0.117	0.009	0.002	0.013
16		0.71	0.208	0.438	0.084	0.249	14.955	1.933	0.152	0.008	0.002	0.013
17	5	0	0.577	0.235	0.085	0.197	6.483	0.204	0.130	0.030	0.003	0.020
18		0.55	0.415	0.536	0.066	0.280	10.813	0.657	0.252	0.016	0.002	0.018
19		0.96	0.523	0.067	0.024	0.167	6.022	0.491	0.113	0.005	0.001	0.010
20		1.25	0.579	0.707	0.089	0.138	12.280	1.258	0.144	0.046	0.004	0.015
21		2.20	0.273	0.392	0.131	0.124	6.805	1.033	0.217	0.039	0.008	0.031

Preliminary clay mineralogical analysis was undertaken on samples 3, 11 and 19 only. Sample 3 was extracted from a soil profile with no visible signs of erosion, Sample 11 was from a gully-head with piping erosion, and Sample 19 was extracted from the active percoline (i.e. zone of preferential subsurface flow) of a profile that had undergone landsliding (Table 1). A Enraf-Nonius X-ray diffractometer with a 120° position-sensitive detector and primary-beam-monochromated Cu $K\alpha_1$ radiation was then used to obtain the diffraction data. Ten minute scan times and three replicate measurements were undertaken. Phase identification was achieved by comparison with the UK Natural History Museum whole-pattern database. Quantification was then performed by sequential matching and stripping of standard patterns reduced in intensity from 100%. Differential adsorption effects were also corrected (Batchelder and Cressey, 1998).

For the porosity measurement, intact 69 cm^3 cores were extracted adjacent to the samples for the other analyses. These cores were saturated gradually for 24 h and the porosity determined gravimetrically. Supplementary analyses of the soil pH were undertaken on 1 : 2.5 soil : water suspensions, and the soil moisture content at the time of sampling was measured gravimetrically with undisturbed 860 cm^3 cores. USDA soil structural class (Landon, 1991) was estimated in the field at the time of soil sampling.

A total of 18 separate physicochemical analyses were, therefore, undertaken on samples taken from each of the 21 sample sites (Tables 2–5).

5. Aggregate stability variation

Most of the soil samples extracted from the locations with active erosion (samples 9–21) have

Table 4
Compound and derived chemical properties of the sampled soil profiles, DVFC, Sabah, Malaysian Borneo

Sample	Profile	Sampling depth (m)	EC_{25} (μS cm^{-1})	SAR	CEC (cmol kg^{-1} soil)	CEC′ (cmol kg^{-1} clay)	ESP (cmol kg^{-1})	Sum of bases	% base saturation	pH	Al^{3+} : CEC (%)
1	1	0	253	0.7	24.2		0.91	6.62	27.3	3.31	48
2		0.25	88	1.35	7.2	37	8.37	1.64	22.8	3.87	68
3		0.51	30	1.91	14.2	58	1.81	1.94	13.7	5.04	54
4		0.88	48	2.28	17.6	118	1.28	1.28	7.3	5.53	88
5	2	0	118	2.57	15.0		1.04	1.13	7.5	4.57	65
6		0.25	62	1.69	18.9	63	1.45	1.08	5.7	4.73	66
7		0.50	217	1.10	17.7	63	1.94	2.33	13.1	4.03	79
8		0.92	38	1.81	17.1	99	2.21	2.04	11.9	6.45	81
9	3	0	199	1.71	9.8		4.14	1.33	13.5	4.01	69
10		0.25	42	1.39	12.0	97	2.87	2.11	17.5	5.08	71
11		0.65	45	2.21	11.9	86	2.58	2.32	19.5	5.39	64
12		0.94	24	1.79	11.9	74	7.34	2.10	17.6	5.43	65
13	4	0	43	1.17	17.9		1.66	3.15	17.6	4.41	63
14		0.25	21	2.28	20.4	97	1.95	2.22	10.9	5.06	75
15		0.51	21	1.58	17.2	72	1.15	0.63	3.7	5.54	83
16		0.71	24	2.15	17.9	89	1.16	0.98	5.5	5.70	84
17	5	0	31	1.01	7.8		7.42	1.09	14.1	5.76	83
18		0.55	39	2.66	12.8	88	3.25	1.30	10.2	6.02	85
19		0.96	18	20.6	7.3	52	7.17	0.78	10.7	5.57	83
20		1.25	32	0.91	15.1	84	3.85	1.51	10.1	5.76	82
21		2.20	60	1.42	8.8	55	3.12	0.92	10.5	4.44	78

CEC′ is an estimate of the effective cation exchange capacity of the clay fraction (i.e., CEC(%clay/100)); it is not estimated for the topsoil, where high levels of organic matter would give an artificially high estimate. The % base saturation is calculated from 100(sum of bases/CEC).

Table 5
Clay mineralogy of selected subsoil strata near DVFC

Sample	3	11	19
Chlorite	7	8	7
Kaolinite-smectite	32	32	26
Illite (10 Angstrom mica)	29	30	37
Vermiculite (14.5 Ansgtrom)	32	30	–
Expandable illite–smectite (14 Angstrom)	–	–	30

Clay minerals are expressed as per cent of the crystalline silicate clays.

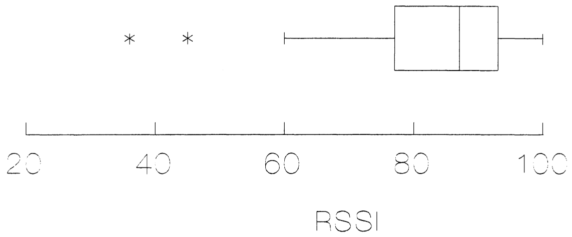

Fig. 3. Box-and-Whisker diagram (Wrinkler and Hays, 1975) of the statistical distribution of aggregate stability of the soils extracted from the eroding profiles.

aggregate stabilities less than 100% (Figs. 2 and 3, Table 2). This contrasts with the locations where signs of erosion are not visible (samples 1–8) where most sample aggregates are unaffected by the rainfall simulation tests (Fig. 2, Table 2). These two datasets are indeed statistically different at the $p = 0.033$ significance level. This association between erosion and instability in soil aggregates, therefore, supports the more detailed observations of Bryan (1968) and De Vleeschauwer et al. (1978).

Large changes in aggregate stability can be observed over small vertical distances within the soil profiles with active erosion (Table 2). This demonstrates that the stability of soil profiles (a) cannot be characterised by samples taken from a single point in a soil profile, and (b) may be governed by particular horizons with weak aggregates.

Identification of clear associations between aggregate stability and particular geomorphic landforms is beyond the scope of the dataset, however, comment on the location of soils with poor aggregate stability may be useful is focusing future research. The sampling sites at the gully head (samples 9–12) that has remained undisturbed by forestry activity exhibited a marked loss of aggregate stability with depth (Table 2). The piping erosion and marked channel-head retreat observed at this location has been associated with poor aggregate stability within other regions, for example, the Alfisols in UK (Jones, 1981) or Spain (Ternan et al., 1998). The least stable soil aggregates within our region were associated with roadfill material (samples 17–21), in particular Sample 19 extracted at 0.9 m depth. A few hours after failure of this roadfill material (on 14/12/1994) movement of fluidised grey (7.5 YR 6/2) soil was still observable over a 1 m^2 zone centred on 0.9 m depth. The water discharging through this failure backwall was probably sourced by ponded drainage on the upslope side of the haul road. Extensive rilling erosion of these exposed roadfill materials was then observed. Reduced aggregate stability was also observed with samples extracted from beneath the skid trail (samples 13–16), however, the stability was greater than observed with the samples extracted from the natural gully head (samples 9–12; Table 2). Future research might examine whether these natural gully heads are indeed more important sediment sources in comparison to anthropogenic sediment sources generated by forestry operations. Bidin (1995) has already demonstrated that the natural channel heads are very active elements of the soil hydrological system.

6. Correlation of aggregate stability with potential causal factors

The rainfall simulation tests used within this study appear sensitive to differences in the stability of weak soil aggregates. Relatively strong soil aggregates are unaffected by the test and give 100% stability. Given that the aggregate stability tests for samples extracted from the non-eroding sites failed to separate stability differences (i.e. most samples recorded 100% RSSI) then only the samples extracted from the eroding sites (i.e. samples 9–21) are considered appropriate for our correlation analyses. Prior to the identification of a multi-parameter predictive models for aggregate stability, single-variable regression analyses of aggregate stability against each of the bulk soil properties was first undertaken.

6.1. Soil texture

All sampled soils near DVFC have between 13 and 25% clay are, therefore, classified mostly as silt loams

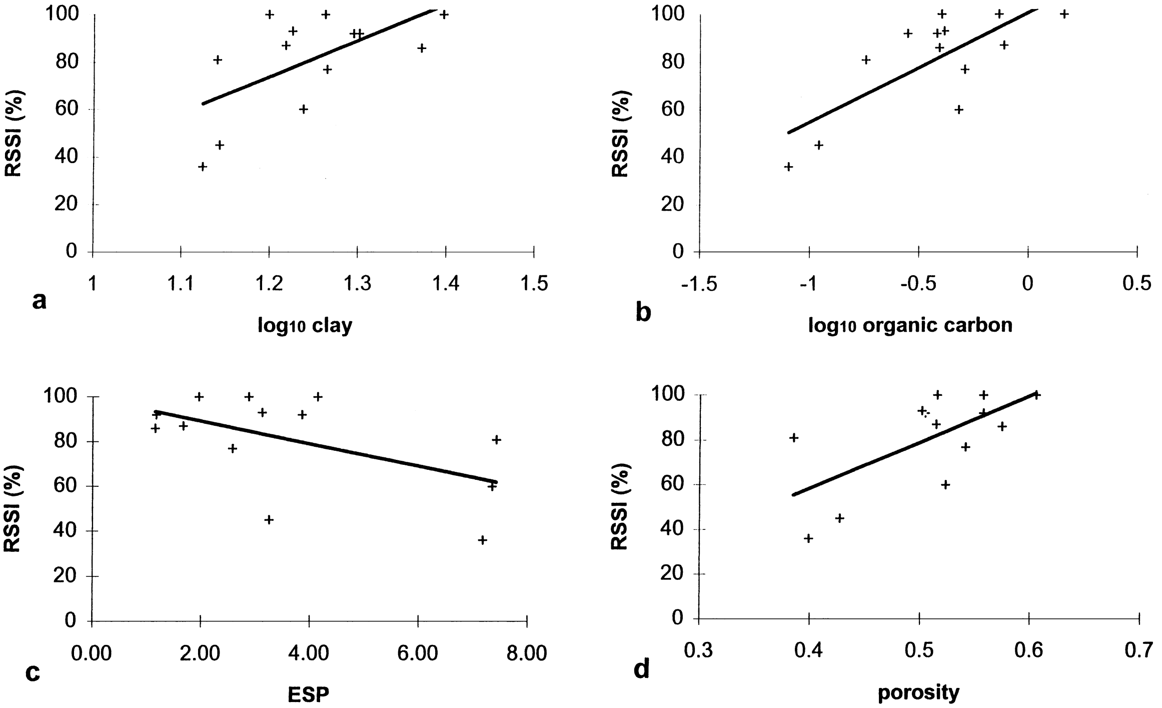

Fig. 4. Aggregate stability (RSSI) at the eroding profiles against (a) $\log_{10}$ linearised clay content, (b) $\log_{10}$ linearised organic carbon, (c) ESP, and (d) porosity.

(Table 2) or simply 'medium textured soils' (FAO–Unesco, 1990). Clay content generally increased from the ground surface to the subsoil, probably as a result of argillation. Clay content was not significantly different (at $p = 0.100$) between the eroding and non-eroding sites, however, a significant but weak positive correlation ($r^2 = 0.371, p = 0.027$) between the $\log_{10}$ linearised clay content and aggregate stability is observed at the eroding sites (Fig. 4(a)). This logarithmic relationship gave a greater degree of explanation in comparison to the linear model applied directly to the raw data ($r^2 = 0.328$, $p = 0.041$). Goldberg et al. (1988) reported very comparable correlations between aggregate stability and clay content ($r^2 = 0.380$, $p < 0.050$) and $\log_{10}$ linearised clay content ($r^2 = 0.410$, $p < 0.050$) following the analysis of 34 Alfisol and Aridisol samples. Levy et al. (1993b) reported an even stronger correlation ($r^2 = 0.651$, $p < 0.010$) between aggregate stability and the $\log_{10}$ linearised clay content within their 17 samples of Ultisol, Vertisol and Alfisol soil. Similarly, Levy and Miller (1997) reported a strong correlation ($r^2 = 0.71$, $p < 0.050$) between aggregate stability and clay content within 11 samples of Ultisol and Alfisol soil. The positive correlation can be explained in all four instances by the beneficial effect of the clay to soil cohesion and, therefore, to aggregate stability.

6.2. Organic carbon

The 21 sampled soils had between 0.08 and 1.53% (or $g\,\mathrm{kg}^{-1}$) organic carbon (Table 2). As expected, samples extracted from the topsoil of the undisturbed slopes (samples 1, 5, and 9) have significantly greater organic contents ($p = 0.016$) than the subsoil samples at the adjacent locations (samples 2–4, 6–8, 10–12; Table 2). While a statistical difference is not observed between the erodible and non-erodible sites ($p > 0.100$), there is a very significant but modest positive correlation ($r^2 = 0.562$, $p = 0.003$) between the $\log_{10}$ linearised organic carbon and aggregate stability at the eroding sites (Fig. 4(b)). Again, the $\log_{10}$ linearised data gave a greater degree of explanation in comparison to the linear model applied directly to the raw data ($r^2 = 0.305$, $p = 0.050$). Goldberg et al. (1988) similarly reported a modest positive correlation ($r^2 = 0.520$, $p < 0.010$) between aggregate stability and organic carbon levels ranging from 0.12 to 1.37%. Chaney and Swift (1984) observed a very much greater correlation ($r^2 = 0.864$, $p < 0.010$) between aggregate stability and organic carbon for the 26 British soils that they studied. These soils, including Inceptisols, Mollisols and Alfisols, had between 0.5 and 9.5% organic matter or ca. 0.3 to 5.5% OC. In some contrast, Le Bissonnais and Arrouays (1997) demonstrated that the strong effect of organic carbon on the aggregate stability ($r^2 = 0.710$) of their Inceptisol soils was only apparent beneath a threshold value of approximately 1.5% OC. In even sharper contrast, Levy et al. (1993b) were unable to identify a significant correlation between aggregate stability and organic carbon (with or without linearisation) with their 17 samples of Ultisol, Vertisol and Alfisol soil. For soils where strong correlations exist, the association can explained by the direct relation between organic carbon and the density of roots, micro-organisms and soil animals, as it is the decomposition products and secretions of these three features that help bind soil aggregates (Marshall and Holmes, 1979).

6.3. Chemical dispersing agents

High soluble salt content within soils, inferred from high bulk soil electrical conductivity (EC_{25}), has been shown to lead to reduced aggregate stability (e.g. Smith et al., 1992). Our statistical analysis does indicate that there is a statistical difference at $p = 0.046$ in EC_{25} between the eroding and non-eroding sites, however, the eroding sites have the lower arithmetic mean EC_{25}. Further, a significant correlation between the EC_{25} and aggregate stability at the eroding sites cannot be observed ($r^2 = 0.105$, $p = 0.270$). This result is supported by the work of Levy and Miller (1997) which demonstrated that salinity is not a dispersing factor where soil pH is less than 6.5. The arithmetic mean pH of the DVFC soil samples was 5.0 and all samples had a pH of less than 6.5 (Table 4). The poor correlation may also be explained by study soils not being particularly saline, having EC_{25} values of less than $4000\,\mu\mathrm{S\,cm^{-1}}$ (Table 4), and by the varying effect of different salts on soil instability.

Sodium salts, due largely to their univalency, are often seen to be the greatest aggregate dispersing agents. This effect acts through (a) greater osmotic potentials, and (b) greater distances between the ion swarm and soil particle surfaces, in comparison to higher valency ions (Holmgren and Flanagan, 1977). A significant correlation (where $p < 0.050$ or even $p < 0.100$) cannot, however, be seen between the aggregate stability and either the exhangeable sodium ($r^2 = 0.178$, $p = 0.151$) nor the water soluble sodium ($r^2 = 0.030$, $p = 0.569$). Statistical differences between the eroding and non-eroding sites for these two properties are similarly not observed at at the $p = 0.100$ level. This is possible, as the dispersing effect of sodium is moderated by the presence of other ions such as calcium and magnesium, in particular where 2 : 1 clays are dominant (Holmgren and Flanagan, 1977). This explains the need to calculate the indices of SAR and ESP (Eqs. (3) and (4)). For these soils, however, there is neither a statistical difference (i.e. $p > 0.100$) in SAR between the eroding and non eroding sites nor a correlation between the SAR and aggregate stability ($r^2 = 0.201$, $p = 0.124$). This poor correlation may result from marked variations in the unmeasured variable of soil carbonate and the more reliable adjusted SAR (Landon, 1991) could be calculated if carbonate levels were known. The relative dryness of the soil at the time of sampling (Table 2) may also make the SAR estimate relatively inaccurate.

In some contrast, the ESP, which is based on the balance between sodium and the other exchangeable cations (Eq. (2)), appears to be negatively correlated with the aggregate stability. While the 21 sampled soils have only low to moderate (i.e. 1–9%) ESP values (Levy et al., 1993b) there is a reasonably significant and weak negative correlation ($r^2 = 0.294$, $p = 0.056$) between the ESP and aggregate stability at the eroding sites (Fig. 4(c)). Other authors have observed the dispersing effect of ESP where levels are less then 1% (Levy and Torrento, 1995) or greater than thresholds of 1% (Levy et al., 1993a), and 5% (Emmerson and Bakker, 1973; Hodgkinson and Thorburn, 1996). A statistical difference between the eroding and non-eroding sites is, however, not observed (i.e. $p > 0.100$).

6.4. Clay mineralogy

Samples 3 and 11, extracted from sites 200 m apart, both have approximately one third of their crystalline silicate clays as kaolinite–smectite (which is 70–80% 1 : 1 kaolinite), one third as 2 : 1 illite and one third 2 : 1 vermiculite (Table 5). A mineralogical difference between the non-eroding site (Sample 3) and the piped site (Sample 11) was, therefore, not apparent with these two samples. Further, the clays at both of these sites exhibited almost no expansive properties following glycollation (M. Batchelder, pers comm.). In the same contrast, the sample extracted from the landslide percoline (Sample 19), the sample location with the least stable soil aggregates (i.e. 36% RSSI), included one third 2 : 1 illite–smectite clay (Table 5) that did show expansive properties under glycollation. Clearly, further sampling is required to identify the complex role of clay mineralogy in aggregate stability at this site. The presence of significant proportions of 2 : 1 clays, in particular expandable 2 : 1 clay (e.g. illite–smectite) might, for example, act against the positive effect of the clay fraction to aggregate cohesion, and also reinforce the role of ESP in aggregate dispersion (Holmgren and Flanagan, 1977).

6.5. Porosity

Evidence for the impact of poor aggregate stability on bulk soil structure was sought through measurements of porosity. The porosity of the 21 sampled soils varied from 0.39 m^3 m^3 to 0.61 m^3 m^3 (Table 2). A statistical difference between the eroding and non-eroding sites is observed at the $p = 0.054$ significance level, and there is a very significant if relatively weak positive correlation ($r^2 = 0.430$, $p = 0.015$) between the porosity and aggregate stability at the eroding sites (Fig. 4(d)). This correlation would support the idea of aggregate breakdown giving a collapse of the bulk soil structure which itself leads to a reduced porosity (Nys and Ranger, 1985; Baumgartl and Horn, 1991).

7. Multi-parameter model of aggregate stability

Soil aggregate stability depends upon the balance of disruptive and stabilising forces and as such presents a synthesis of the physical and chemical properties controlling aggregate dispersion. It is, therefore, not surprising that correlation of aggregate stability with the individual factors produced only relatively weak relationships. The single-variable regression analysis did, however, indicate that the bulk soil properties of clay content, organic carbon, ESP and porosity show some association with the compound property of aggregate stability. A multiple regression of all four properties with aggregate stability gives the following model:

$$\begin{aligned}\mathrm{RSSI} = 46.567 - (250.520\eta) - (1.731\mathrm{ESP}) \\ + (157.810[\log_{10}\mathrm{clay}]) + (65.765[\log_{10}\mathrm{OC}])\end{aligned} \tag{7}$$

where RSSI is the rainfall simulation survival index (%), η is the porosity (m^3 m^{-3}), ESP is the exchangeable sodium percentage, $\log_{10}$ clay is the $\log_{10}$ linearised clay per cent, and $\log_{10}$ OC is the $\log_{10}$ linearised organic carbon per cent. This model describes a relationship that is statistically very significant ($p = 0.012$) with a strong coefficient of determination ($r^2 = 0.766$). Examination of a correlation matrix for these regressors (Table 6) does, however, indicate that the porosity is strongly correlated with the $\log_{10}$ linearised clay ($r^2 = 0.526$, $p = 0.005$) and $\log_{10}$ linearised organic carbon ($r^2 = 0.673$, $p = 0.001$). The high correlation between the model regressors resulting from the inclusion of porosity increases the likelihood of multicolinearity, therefore, making Eq. (7) unreliable. Furthermore, the porosity is affected by rather than affecting aggregate stability unlike ESP, $\log_{10}$ clay and $\log_{10}$ OC. Porosity is, therefore, omitted from the analysis, so the RSSI model becomes:

$$\begin{aligned}\mathrm{RSSI} = 5.243 - (0.973\mathrm{ESP}) + (76.059[\log_{10}clay]) \\ + (35.845[\log_{10}\mathrm{OC}])\end{aligned} \tag{8}$$

which is still very significant ($p = 0.013$) and has a reasonably strong positive correlation ($r^2 = 0.680$). The predictive ability of the model is shown in Fig. 5 and the level of variance explained is not dissimilar to the levels of explanation afforded by the aggregate stability models of, for example, Goldberg et al. (1988). The RSSI model is intended to be a preliminary model focusing the parameterisation of future more spatially intensive sampling frames. A larger dataset may allow, for example, identification of

Table 6
Correlation matrix for potential regressors of aggregate stability in the eroding soils

	Log_{10} carbon	EC_{25}	Na_x	SAR	ESP	Porosity
Log_{10} clay	0.164	0.047	0.115	0.001	0.430^{**}	0.526^{**}
Log_{10} carbon	–	0.338^{**}	0.057	0.074	0.166	0.673^{***}
EC_{25}	–	–	0.009	0.005	0.000	0.174
Na_x	–	–	–	0.020	0.709^{***}	0.137
SAR	–	–	–	–	0.029	0.005
ESP	–	–	–	–	–	0.005
Porosity	–	–	–	–	–	–

r^2 $^*p < 0.100$.
$^{**}p < 0.050$.
$^{***}p < 0.010$.

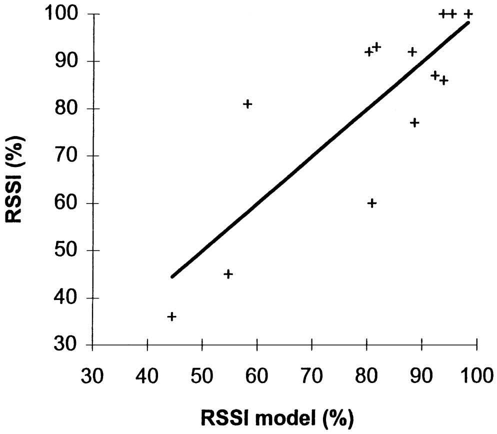

Fig. 5. Predictive ability of the three property RSSI model.

separate relationships for each sub-erosional process (i.e. surficial-, gully-, piping- or landslide triggered-erosion), so that where the process is known a priori, greater predictive certainty might be obtained. Furthermore, an aggregate stability model applicable to soils with relatively strong aggregates as well as weak aggregates needs to be developed. To achieve this, an aggregate stability measure may need to be derived by combining the results of two different aggregate testing procedures.

8. Possible association with soil type

The previous studies cited here have shown that unstable soil aggregates can be observed in Ultisols, Vertisols, Alfisols, Aridisols, Inceptisols, and Mollisols. Driessen and Dudal (1991) have, however, stated that some tropical soil types exhibit greater signs of erosion than others, and hence may be intrinsically more erodible. All of the soil profiles sampled near DVFC have argillic horizons with a base saturation of less then 35% (Table 4), lack spodic or plinthic horizons, albic tongues or a xeric water regime. Our physicochemical data, therefore, confirm that the soils should be classified as Ultisols (cf. Soil Survey Staff, 1975), which are equivalent to the FAO Acrisol–Alisol groups (Landon, 1991).

According to Driessen and Dudal (1991) the new separation of Acrisol and Alisol groups (cf. FAO–Unesco, 1979, 1990) is important for soil erosion studies. They state that Alisol soils have significantly more erodible soil aggregates than Acrisol soils as a result of their smaller proportion of stable kaolinitic clays. This smaller proportion of kaolinitic clays is manifested in a cation exchange capacity in excess of 24 cmol kg^{-1} clay (CEC′) in the Bt horizon. All of the soil profiles sampled near DVFC have a cation exchange capacity in excess of 24 cmol kg^{-1} clay (CEC′), and together with their argic B horizon and base saturation of less than 35% (Table 4) means that the soils can be classified in the Alisol group (FAO–Unesco, 1990). The relatively low proportion of kaolinitic clays in the three soil samples (Table 5) also supports their classification as Alisols. For context, the ACf Ferric Acrisol of central Singapore (Rengam series) is dominated by kaolinite (58%) and has a CEC′ of only 7 cmol kg^{-1} clay (Rahman, 1993; Chappell, unpublished). Alisol soils by definition also have a high concentration of free aluminium (FAO–Unesco, 1990). Aluminium dominates the exchange complex within all of the sampled soils at DVFC, with an Al^{3+} : CEC ratio greater than 60% in all soils, except Sample 1 (Table 4). Aluminium toxicity is expected with such levels and when combined with the noticeable deficiency in calcium and magnesium, with levels typically below 4 and 0.5 $\mathrm{cmol\,kg^{-1}}$, respectively, (Table 3) may reduce biological activity in the soil (Landon, 1991). The reduced biological activity may in turn reduce the aggregate stability (Marshall and Holmes, 1979; Tisdall and Oades, 1982; Driessen and Dudal, 1991).

The visual logging of the soil profiles during sampling for this study has allowed classification to the sub-group or 'FAO soil unit' level. With the exception sampling location 19, there was an absence of ferric, plinthic, strongly humic, gleyic and stagnic properties recorded. This indicates that samples 1–16 were collected from an ALh Haplic Alisol soil unit. The presence of localised gleying at sampling location 19, the landslide site, means that this soil profile (samples 17–21) can be classified as an ALg Gleyic Alisol.

9. Discussion and conclusions

This exploratory study has been undertaken within a 10 km^2 region where tropical Ultisol soils are subject to disturbance by selective, commercial logging, and where the stability of these soil aggregates against erosion may be critical to the estimation of the spatial variability of natural and anthropogenically-induced sediment sources.

1. The combination of soil physical and chemical data indicates that all of the soils can be classified in the FAO Alisol group, which Driessen and Dudal (1991) state have characteristically unstable soil aggregates. Within this single soil type those sampled profiles exhibiting erosion at rates in excess of 10 $\mathrm{mm\,a^{-1}}$ had significantly more unstable aggregates in comparison to the profiles without visible signs of erosion. There was, however, a very large variability in aggregate stability between horizons of the same soil profile, thereby, indicating that horizon-specific data may be important in the parameterisation of erosion models. Indeed, the preliminary data indicates that the horizons of weak aggregates within the channel-head and landslide profiles (profiles 3 and 5, respectively) seemed to be the focus for the erosion/instability, in the form of piping activity and soil fluidisation (during mass failure). Of equal importance was the observation that the variability between natural slopes (i.e. samples 1–12) was as great as the differences between those subject to vehicle compaction or disturbed by road construction (i.e. samples 13–21). This implies that research into the rates of erosion from forestry-induced sediment sources (e.g. skid trails, cut-and-fill slopes and culverts), must also address the variability in the equally important erosion (or instability) of natural sediment sources (e.g. channel-heads).
2. The correlation analyses indicated that the bulk soil properties of organic carbon and clay content may be significant stabilising influences on soil aggregates in the tropical Ultisol studied, while ESP may be a significant disrupting factor. It is, therefore, suggested that in future more intensive assessment of the factors governing the spatial variability in aggregate stability should focus on these three properties – organic carbon, clay content and exchangeable sodium percentage.
3. While the multiple-regression model of aggregate stability identified is preliminary, it may indicate that the identification of soil properties affecting different sub-erosional processes (i.e. surficial-, gully-, piping- and landslide-triggered erosion), may be achieved with a single model. This would have the advantage over process-specific models (cf. Bryan et al., 1989) of not requiring any a priori understanding of the sub-erosional processes taking place within a region.
4. The rainfall simulation test of aggregate stability did, however, prove relatively insensitive where the Ultisol soils were relatively resistant to dispersion. Therefore, additional aggregate stability testing methods need to be applied to these soils, and perhaps the results combined in a unified index. Ternan et al. (1998) demonstrated that the wet-sieve aggregate testing method (see Yoder, 1936; Kemper and Koch, 1966; Kemper and Rosenau, 1986) was more discriminatory for relatively stable soils, while the RSSI was more discriminatory for relatively weak soils. The combination of results

from RSSI and wet-sieving tests applied to the same soils is, therefore, suggested for further work to characterise the spatial variability of aggregate stability within tropical Ultisols.

5. This exploratory analysis has indicted that future more intensive sampling programmes ought to assess (a) whether statistically significant differences in aggregate stability exist between different tropical soil types as postulated by Driessen and Dudal (1991), (b) if certain soil horizons within these soils are particularly unstable, and (c) whether forestry activities produce instability in soils (through vehicle compaction or changed subsurface drainage) or simply expose unstable horizons.

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References

Batchelder, M., Cressey, M., 1998. Rapid, accurate phase quantification of clay-bearing samples using a position-sensitive X-ray detector. Clays and Clay Minerals 46, 183–194.

Baumgartl, T., Horn, R., 1991. Effect of aggregate stability on soil compaction. Soil and Tillage Res. 19, 203–213.

Bidin, K., 1995. Subsurface Flow Controls of Runoff in a Bornean Natural Rainforest. Unpublished M.Sc. Thesis, University of Manchester, UK.

Brooks, S.M., Richards, K.S., Nussbaum, R., 1994. Simulator experiments of the varied consequences of rain forest logging for runoff and erosion. Geografiska Annaler 76A, 143–152.

Bryan, R.B., 1968. The development, use and efficiency of indices of soil erodibility. Geoderma 2, 5–26.

Bryan, R.B., Govers, G., Poesen, J., 1989. The concept of soil erodibility and some problems of assessment and application. Catena 16, 393–412.

Bruijnzeel, L.A., 1990. Hydrology of Moist Tropical Forests and Effects of Conversion: A State of Knowledge Review, UNESCO, Paris, p. 224.

Bruijnzeel, L.A., 1996. Predicting the hydrological impacts of land cover transformation in the humid tropics: The need for integrated research. In: Gash, J.H.C., Nobre, C.A., Roberts, J.M., Victoria, R.L. (Eds.), Amazonian Deforestation and Climate, Wiley, Chichester, pp. 15–55.

Chaney, K., Swift, R.S., 1984. The influence of organic matter on aggregate stability of some British soils. J. Soil Sci. 35, 223–230.

Chappell, N.A., McKenna, P., Bidin, K., Douglas, I., Walsh, R.P.D., 1998. Upscaling suspended-sediment flows in disturbed rain forest terrain: role of localised new sources, In: Brooks, S., Longley, P., Macmillan, W., McDonnell, R., Abrahart, R. (Eds.), Proc. 3rd Int. Conf. GeoComputation (GeoComputation '98), University of Bristol, UK, 17–19 September 1998.

Clennell, B., 1996. Far-field and gravity techtonics in Miocene basins in Sabah, Malaysia. In: Hall, R., Blundell, D. (Eds.), Tectonic Evolution of Southeast Asia, Geological Society Special Publication No. 106, pp. 307–320.

de Roo, A.P.J., 1993. Modelling Surface Runoff and Soil Erosion in Catchments using Geographical Information Systems, Netherlands Geographical Studies 157, Koninklijk Nederlands Aardrijkskundig Genootschap/Faculteit Ruimtelijke Wetenschappen, Rijksuniversiteit Utrecht, Nederlands, p. 295.

Douglas, I., 1992. Hydrologic and geomorphic changes following rainforest disturbance with special reference to Borneo. In: Ismail, G., Mohamed, M., Omar, S. (Eds.), Forest Biology and Conservation in Borneo, Centre for Borneo Studies, Publication 2, pp. 261–287.

Douglas, I., Spencer, T., Greer, T., Bidin, K., Sinun, W., Wong, W.M., 1992. The impact of selective commercial logging on stream hydrology, chemistry and sediment loads in the Ulu Segama Rain Forest Sabah. Phil. Trans. Roy. Soc. London B 335, 397–406.

Driessen, P.M., Dudal, R., 1991. The Major Soils of the World, Agricultural University Wageningen, p. 310.

De Vleeschauwer, D., Lal, R., De Boodt, M., 1978. Comparison of detachability indices in relation to soil erodibility for some important Nigerian soils. Pedologie XXVIII 1, 5–20.

Emmerson, W.W., Bakker, A.C., 1973. The comparative effects of exchangeable calcium, magnesium and sodium on some physical properties of red-brown earth subsoils: 2. The spontaneous dispersion of aggregates in water. Aust. J. Soil Res. 11, 151–157.

FAO–Unesco, 1979. Soil Map of the World: Volume IX, Southeast Asia, Unesco, Paris, p. 49.

FAO–Unesco, 1990. Soil Map of the World, Revised Legend, FAO, Rome, p. 119.

Goldberg, S., Suarez, D.L., Glaubig, R.A., 1988. Factors affecting clay dispersion and aggregate stbility of arid-zone soils. Soil Sci. 146, 317–325.

Hodgkinson, R.A., Thorburn, A.A., 1996. Factors influencing the stability of salt affected soils in the UK – criteria for identifying appropriate management options. Agric. Water Manage. 29, 327–338.

Holmgren, G.G.S., Flanagan, C.P., 1977. Factors affecting spontaneous dispersion of soil materials as evidenced by the crumb test. In: Sherard, J.L., Decker, R.S. (Eds.), Dispersive Clays, Related Piping, and Erosion in Geotechnical Projects, ASTM STP 623, Am. Soc. Testing and Materials, pp. 218–239.

Jones, J.A.A., 1981. Piping effects in humid lands. In: Higgins, C.G., Coates, D.R. (Eds.), Groundwater Geomorphology, Boulder, Colorado, Geological Society of America, Special Paper 252, pp. 111–137.

Kemper, W.D., Koch, E.J., 1966. Aggregate stability of soils from the western portions of the United States and Canada. US Dep. Agric. Tech. Bull, p. 1355.

Kemper, W.D., Rosenau, R.C., 1986. Aggregate stability and size distribution, In: Klute, A. (Ed.), Methods of Soil Analysis, Part 1. Physical and Mineralological Methods. Am. Soc. Agronomy/ Soil Sci. Soc. America, Madison, USA, pp. 425–442.

Lal, R., 1980. Physical and mechanical composition of Alfisols and Ultisols with particular reference ro soils in the tropics. In: Theng, B.K.G. (Ed.), Soils with Variable Charge. New Zealand Soc. of Soil Sci., Soil Bureau, Dept. Sci. Indust. Res., Lower Hut, New Zealand, pp. 253–279.

Lal, R., 1981. Analyses of different processes governing soil erosion by water in the tropics. In: Walling, D.E. (Ed.), Erosion and Sediment Transport Measurement, Proc. Florence Symp., June 1981, IAHS Publication No. 133, pp. 351–364.

Landon, J.R., 1991. Booker Tropical Soil Manual. Longman, New York, p. 474.

Leong, K.M., 1974. The Geology and Mineral Resources of the Upper Segama Valley and Darvey Bay Area, Sabah, Malaysia. Geological Survey of Malaysia, Memoir 4 (revised), Government Printer, Kuching, p. 105.

Levy, G.J., Agassi, M., Smith, H.J.C., Stern, R., 1993a. Microaggregate stability of kaolinitic and illitic soils determined by ultrasonic energy. Soil Sci. Soc. Am. J. 57, 803–808.

Levy, G.J., Eisenberg, H., Shainberg, I., 1993b. Clay dispersion as related to soil properties and water permeability, Soil Sci. 155, 15–22.

Levy, G.J., Miller, W.P., 1997. Aggregate stabilities of some southeast US soils. Soil Sci. Soc. Am. J. 61, 1176–1182.

Levy, G.J., Torrento, J.R., 1995. Clay dispersion and macroaggregate stability as affected by exchangeable potassium and sodium. Soil Sci. 160, 352–358.

Le Bissonnais, Y., Arrouays, D., 1997. Aggregate stability and assessment of soil crustability and erodibility: II. Application to humic loamy soils with various organic carbon contents. European J. Soil Sci. 48, 39–48.

Marsh, C.W., Greer, A.G., 1992. Forest land-use in Sabah, Malaysia: An introduction to the Danum Valley. Philosophical Transac. Royal Soc. London series B 335, 323–457.

Marshall, T.J., Holmes, J.W., 1979. Soil Physics. Cambridge University Press, Cambridge, p. 345.

Mason, B.J., Andrews, J.B., 1960. Drop-size distributions from various types of rain. Quart. J. Royal Meteorol. Soc. 86, 346–353.

Morgan, R.P.C., 1979. Soil Erosion. Longman, New York, p. 345.

Nys, C., Ranger, J., 1985. Influence de l'espece sur le fonctionnement de l'écosyst_me forestier. Le cas de la substitution de'une essence resineuse áune essence feuille. Sci. Sol. 21, 203–216.

Prior, D.B., Ho, C., 1972. Coastal and mountain slope instability on the islands of St. Lucia and Barbados. Engineering Geol. 6, 1–18.

Rahman, A., 1993. Soil of Singapore. In: Gupta, A., Pitts, J. (Eds.), Physical Adjustments in a Changing Landscape. Singapore University Press, pp. 144–189.

Rhodes, J.D., 1982. Cation exchange capacity. In: Page, A.L., Miller, R.H., Keeney, D.R. (Eds.), Methods of Soil Analysis. Am. Soc. Agron., Madison, pp. 149–157.

Rose, C.W., 1993. Erosion and sedimentation. In: Bonell, M., Hufschmidt, M.N., Gladwell, J.S. (Eds.), Hydrology and Water Management in the Humid Tropics. Cambridge University Press, Cambridge, pp. 301–343.

Sherlock, M.D., 1997. Plot-scale hydrometric and tracer characterisation of soil water flow in two tropical rain forest catchments in Southeast Asia. Unpublished Ph.D. Thesis, Lancaster University, UK.

Smith, G.G., Foley, J.L., Loch, R.J., 1992. Effects of electrical conductivity of water used in wetting and in wet sieving on measured aggregate water stability. Soil Technol. 5, 177–184.

Soil Survey Staff, 1975. Soil Taxonomy, Agriculture Handbook No. 436, Soil Conservation Service, USDA, Washington, Wiley, New York, p. 754.

Sugimura, Y., Suzuki, Y., 1988. A high temperature catalytic oxidation method for the determination of non volatile dissolved organic carbon in sea water by direct injection of a liquid sample. Marine Chemistry 24, 105–131.

Tisdall, J.M., Oades, J.M., 1982. Organic matter and water stable aggregates in soils. J. Soil Sci. 33, 141–163.

Ternan, J.L., Williams, A.G., Elmes, A., Hartley, R., 1996. Aggregate stability of soils in Central Spain and the role of land management. Earth Surface Processes and Landforms 21, 181–193.

Ternan, J.L., Elms, A., Fitzjohn, C., Williams, A.G., 1998. Piping susceptibility and the role of hydrogeomorphic controls in pipe development in alluvial sediments, Central Spain. Zeitschrift für Geomorphologie 42, 75–87.

Walsh, R.P.D., Bidin, K., 1995. Channel head erosion in primary and logged forest in Sabah. In: Abstr. Int. Assoc. Geomorphologists, Southeast Asia Conference, Singapore, 18–23 June 1995, p. 79.

Wrinkler, R.L., Hays, W.L., 1975. Statistics: Probability, Inference and Decision, 2nd ed. Holt, Rinehart and Wiston, New York, p. 889.

Wright, P.S., 1975. The Soils of Sabah. Land Resources Division, Ministry of Overseas Development, UK, p. 71.

Yoder, R.E., 1936. A direct method of aggregate analysis of soils and a study of the physical nature of erosion losses. J. Am. Soc. Agron. 28, 337–351.