

Surface runoff and erosion

Experimental study of Horton overland flow on tropical hillslopes

1. Soil conditions, infiltration and frequency of runoff

by

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with 6 figures and 3 tables

Zusammenfassung. Künstliche starke Beregnungen wurden angewendet, um die Infiltration in stark erodierten Böden des Weidelandes von S-Kenia zu studieren. Bodentextur und -struktur ändern sich systematisch am Hang und steuern Infiltrationsraten. Eine Einwirkung der Vegetationsdichte auf die Infiltration konnte innerhalb der Variationsbreite nicht entdeckt werden. Auf ursprünglich trockenen, sandig tonigen Luvisolen schwankte die Infiltrationskapazität am Ende einer einstündigen Beregnung von 22–35 mm/hr, nahm aber ab während späterer Beregnungen aufgrund des Quellens der Kolloide, der Abnahme des potentiellen Gradienten mit dem Vorrücken der Feuchtigkeitsfront in größere Tiefen und der Entwicklung einer oberflächlichen Kruste. Immerhin, sehr intensive Niederschläge können genügend Flächenspülung produzieren, um diese Kruste abzutragen und die Infiltrationskapazität um bis zu 30 % zu erhöhen. In tonigen Vertisolen am Hangfuß wird die Infiltration gesteuert durch den Grad der Trockenrisse und die weitere Entwicklung dieser Risse, nachdem der Regen begonnen hat. Die Infiltrationskapazität kann anfangs 100 mm/hr übersteigen, nimmt aber sehr schnell ab bis auf weniger als 10 mm/hr, wenn die Durchfeuchtung andauert. Am oberen Hang erreicht eine tonige Rendzina, weniger aufgerissen als die Vertisole am Hangfuß, eine viel geringere Infiltrationskapazität, die einen konstanten Wert von 3–4 mm/hr nach der Infiltration von 110 mm Niederschlag annimmt. Ungefähr zweimal so viel Infiltration ist notwendig, um diese Konditionen in einem Vertisol zu erlangen. Niederschlagsüberschuß von einem sandigen, tonigen Boden tritt durchschnittlich weniger als 7–11 mal pro Jahr auf und erreicht weniger als 20 % des Jahresniederschlags. Wegen der großen saisonalen Variationen der Infiltrationskapazität in den aufreißenden Tönen ist es bis jetzt noch nicht möglich, den Umfang und die Häufigkeit des Niederschlagsüberschusses dort vorherzusagen.

Summary. Artificial rainstorms were used to study infiltration into severely eroded soils in the rangelands of southern Kenya. Soil texture and structure vary systematically along hillslopes and control infiltration rates; the effect of vegetation density on infiltration could not be detected within the range of cover available. On initially dry sandy clay Luvisols, infiltration

capacities at the end of a one-hour storm range from 22 to 35 mm/hr, but decline during later storms due to the swelling of colloids, the decrease in potential gradient as the wetting front penetrates to greater depths, and the development of a surface crust. However, very intense rainstorms may produce enough sheetwash to erode the crust and increase the infiltration capacity by as much as 30 percent. In clay Vertisols on footslopes infiltration is controlled by the degree of crack development and the evolution of the cracks after rain begins. Infiltration capacities which can initially exceed 100 mm/hr decrease rapidly to less than 10 mm/hr as wetting proceeds. On an upper slope, a clay Vertic Rendzina, less cracked than the Vertisols of the footslopes, has a much lower infiltration capacity which attains a constant value of 3–4 mm/hr after the infiltration of 110 mm of rainfall. Approximately twice as much infiltration is required to achieve this condition on the Vertisols. Precipitation excess from the sandy clay soils occurs on average less than 7–11 times per year and totals less than 20 percent of the annual rainfall. Because of the large seasonal variation in infiltration capacity in the cracking clays it is not yet possible to predict the magnitude and frequency of precipitation excess there.

Résumé. Des tempêtes de pluie artificielle furent utilisées pour étudier les conditions d'infiltration en sol très érodé des steppes du sud du Kenya. La texture et la structure du sol varient systématiquement le long des pentes et contrôlent les taux d'infiltration; par contre, l'effet de la densité de la végétation sur l'infiltration n'a pu être détecté sur la portée du couvert disponible. Sur les argiles sableux (Luvisols) sèches au départ, les capacités d'infiltration après une heure de pluie sont de 22 à 35 mm/hr, mais déclinent lors de tempêtes subséquentes dû au gonflement des colloïdes, à la baisse de gradient potentiel alors que le sol devient de plus en plus humide, et au développement d'une croûte en surface. Cependant, une pluie très intense peut éroder suffisamment pour vaincre cette croûte et accroître la capacité d'infiltration de plus de 30 %. Sur les argiles Vertisols des bas de pente, l'infiltration dépend du développement de fissures et de l'évolution de ces fissures lors de la précipitation. La capacité d'infiltration qui dépasse 100 mm/hr au départ décroît rapidement à moins de 10 mm/hr alors que le mouillage progresse. Sur une pente supérieure, l'argile Vertic Rendzina, moins fendillée que les Vertisols des bas de pente, possède une capacité d'infiltration très inférieure qui approche une constante de 3–4 mm/hr après infiltration de 110 mm de pluie. Près du double de l'infiltration est nécessaire pour atteindre cette condition dans les Vertisols. L'excédent de précipitation sur argiles sableux se produit en moyenne moins de 7–11 fois par an et totalise moins de 20 % de la précipitation annuelle. Dû à l'importance variation saisonnière de la capacité d'infiltration sur argiles craquelées il n'est toujours pas possible de prédire l'importance et la fréquence de l'excédent de précipitation.

Introduction

Recent environmental fluctuations in African rangelands have stimulated interest in hydrologic processes in tropical grasslands. There is particular concern about the intensification of Horton overland flow and of sheetwash erosion and the effects of accelerated erosion on soil properties including those which control moisture storage. As a first step in understanding the factors which control the water balance, storm-runoff hydrology, and rates of soil loss in rapidly eroding landscapes in Kenya, we have conducted an experimental study of infiltration and runoff on 5 m-long plots on three soils which characterize large areas of the country. Here we present a description of the physical conditions influencing the temporal and spatial variation of infiltration into these soils, and we combine measured infiltration capacities with climatic data to predict the frequency of runoff from two of the soils. In subsequent papers we will use the results of the experiments to characterize the spatial pattern of hydraulic roughness on the

slopes and to compute hillslope hydrographs (DUNNE & DIETRICH 1980) and to assess the controls on soil loss (DUNNE & DIETRICH, in preparation). We wish to emphasize the agreement between our results and the partial-area concept of Horton overland flow generation, formulated by BETSON (1964) and documented by YAIR et al. (1978). Our results indicate that the runoff generation and hydraulic characteristics vary spatially and from storm to storm.

The study area

Experiments were carried out in three regions of Kajiado District in southern Kenya: the Athi-Kapiti Plains; the Kimana area on the distal slopes of Mt. Kilimanjaro; and the Amboseli lowland (see fig. 1). The sites were chosen to represent the soils and other landscape characteristics of the three major rock types that underlie the region. At each site, distinctive systematic variations in gradient, soil properties, and hydraulic roughness have developed along hillslope profiles, and they strongly influence the local infiltration rates, runoff production, runoff characteristics, and soil erosion. In the following description we outline these controls on the spatial pattern of Horton overland flow.

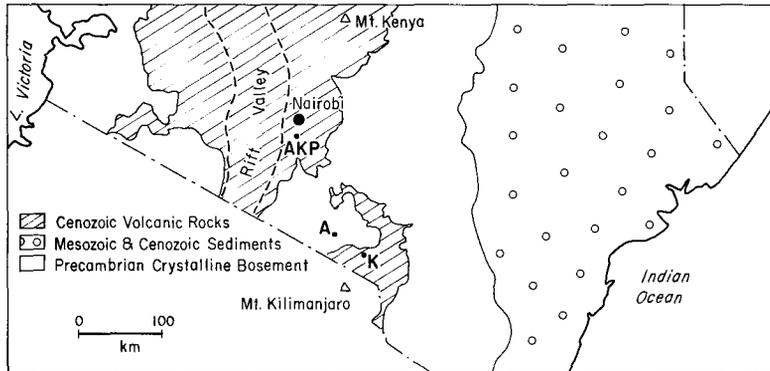


Fig. 1. Schematic geologic map of southern Kenya, showing the location of the study sites in the three geologic regions. AKP indicates the Athi-Kapiti Plains; A, the Amboseli region; and K, the Kimana area.

Geology, topography and soils

The Athi-Kapiti Plains (AKP, fig. 1) are underlain by a sequence of sodic lavas and tuffs which have been eroded into broad, flat-topped ridges and convex-concave hillslopes with lengths ranging from 100 to almost 1000 m. Most hillslopes have maximum gradients of 0.1 to 0.3 but their profiles are dominated by long, slightly concave footslopes with gradients of 0.015 to 0.025. The volcanic rocks weather to grey or black montmorillonitic clay soils with various degrees of cracking, character of surface material and depth according to their position on hillslopes. On ridge tops, the soils are Planosols, approximately 0.5 m deep, and consist of a surface debris of grey sandy loam, granules and grass litter

over a blocky, dense subsoil of black clay. No cracks are present at the surface. On the steeper slopes, 0.5 m-deep soils classified as Rendzinas, Phaeozems and Vertic Phaeozems have up to 2.0 major cracks per metre of contour, and a 1.5 cm thick layer of loose, aggregated grey silt loam with abundant pebbles and a fragmented surface crust like those described by McINTYRE (1958 a, b). Between 1.5 cm and 7.5 cm below the surface the soil consists of 1 cm x 1 cm blocks with 2 to 3 mm-wide interpedal cracks, overlying a subsoil with major vertical cracks, 1 cm wide and 13 cm apart. Between the major cracks the subsoil is divided into 3 cm x 4 cm blocks and the intervening cracks are 1 to 2 mm wide. Soils on the long, gentle footslopes are 0.8 to 1.0 m-thick clay Vertisols with cracks that are broad (up to 2 cm wide at the surface), numerous, and deep. At the surface, the spacing of major cracks ranges from 2 to 4 per metre of transect. The surface between the cracks is covered with a thin (less than 1 cm) layer of grey sand- and silt-size aggregates of clay. The soil has a strong angular blocky structure and exhibits slickensides. The upper 20 cm consists of 3 cm x 3 cm parallelepiped blocks separated by cracks 1 to 2 mm in width. Below this level the interpedal crack spacing increases uniformly to about 6 cm at a depth of 60 cm and 10 cm at the base of the profile. The surface is irregular due to swelling and shrinking of the clays.

Basaltic lavas from Mt. Kilimanjaro underlie a range of topography at Ki-mana (fig. 1). They extend from the south in layers, giving the land a stepped appearance. Steep escarpments covered with lava boulders separate hillslopes with lengths of 100 to 700 m and gradients less than 0.05. Some of the steeper portions of these hillsides (gradients > 0.03) are rilled and gullied. Soils are brown to dark reddish Luvisols composed of kaolinitic brown silty clays and sandy clay loams. Texture varies with distance from ridge crests and rock outcrops, and with respect to local gradient. On upper slopes with gradients of 0.04 to 0.05, soils range from a thin (< 0.1 m) gravel near rock outcrops to a gravel-covered silty clay with depths as great as 0.7 m. On the footslopes sandy clay loams exceed 1 m in depth and have a subsoil with weakly developed subangular blocky structure. The upper 10 to 30 mm of soil consists of thin ($\cong 1$ mm), brittle plates of fine sediment interlayered with coarser sands. The upper 2 mm of crust are probably formed by the breakdown of soil aggregates through the processes of slaking, bombardment, and washing-in of fine particles, as described by McINTYRE (1958 a, b), FARRES (1978), and others. A thin (0.1 mm) surface seal similar to that described by McINTYRE was widely recognized, although it was interrupted by clumps of vegetation, hoofprints, and patches of coarse sand where concentrations of overland flow had occurred.

Between the two volcanic outcrops of the Athi-Kapiti Plains and the Ki-mana region lie Basement schists, gneisses, quartzites and crystalline limestones of the Amboseli lowland (fig. 1). These rocks have eroded to a gently rolling landscape with broad flat hilltops and long slopes up to 1200 m in length and with gradients of 0.02 or less which typically end in swales. The soils are dark red to reddish brown kaolinitic sandy clays and sandy clay loams that are classified as Luvisols. Average soil thickness varies from 0.5 m on gradients of 0.02–0.03 near hilltops to about 1.0 m on the footslopes. On hillsides steeper than 0.02, the soil can become rocky or be entirely absent. The characteristics of the soil

surface are similar to those on the Kilimanjaro lavas, except for the more abundant occurrence of a thin surface layer of coarse sand. Below the widespread 20 mm-thick platy surface horizon the soil has a weakly developed subangular blocky to prismatic structure with thin cutans and discontinuous fine pores.

Gentle swales at the lower ends of the Amboseli and Kimana hillslopes are shallow, narrow bedrock depressions filled with 1 m-thick dark-brown soils rich in cracking clays which are dissected by sandy discontinuous gullies. As will be described later these soils are important runoff- and grass-producing zones.

Climate and vegetation

Mean annual rainfall varies from approximately 750 mm at the measurement sites on the Athi-Kapiti Plains to 450 mm at Kimana, and to 300 mm on the Basement at Amboseli (Government of Kenya 1970). Most of the precipitation occurs in two rainy seasons, separated by long, intense dry seasons. Potential evapotranspiration varies from 1800 mm/yr on the Athi-Kapiti Plains to 2000 mm/yr in the Amboseli area.

Over most of the study area, the vegetation cover is bushed grassland or

Table 1. Characteristics of the plots. Each of the plots was 5 m long.

| Geologic and physiographic area and soil type | Plot no. | Width of plot (m) | Gradient | Ground cover (%) | | Bulk density of upper 10 cm (g/cm ³) |
|---|----------|-------------------|----------|------------------|---------|--|
| | | | | Original | Reduced | |
| Athi-Kapiti Plains | | | | | | |
| Clay Vertic Rendzina | 2 | 2.0 | 0.085 | 65 | 32 | 1.07 |
| Clay Vertisol | 3 | 1.2 | 0.020 | 76 | 36 | — |
| Clay Vertisol | 4 | 2.0 | 0.025 | 64 | — | — |
| Clay Vertisol | 5 | 2.0 | 0.024 | 78 | 36 | 1.05 |
| Amboseli Basement | | | | | | |
| Sandy clay | | | | | | |
| loam Luvisol | 6 | 2.0 | 0.020 | 10 | 10 | 1.32 |
| Sandy clay | | | | | | |
| loam Luvisol | 7 | 2.0 | 0.020 | 8 | 8 | 1.41 |
| Sandy clay | | | | | | |
| loam Luvisol | 10 | 1.2 | 0.020 | 12 | 12 | 1.32 |
| Sandy clay | | | | | | |
| loam Luvisol | 11 | 1.2 | 0.050 | 41 | 16 | 1.62 |
| Kilimanjaro Lavas | | | | | | |
| Stone-covered sandy clay | | | | | | |
| loam Luvisol | 12* | 1.2 | 0.040 | 0 | 0 | 1.27** |
| Sandy clay | | | | | | |
| loam Luvisol | 13* | 1.2 | 0.015 | 0.04 | 0.04 | 1.05 |
| Sandy clay | | | | | | |
| loam Luvisol | 14* | 1.2 | 0.038 | 7 | 7 | — |

* 92 percent of the surface of plot 12 was covered with gravel with a median diameter of 11 mm and a maximum diameter of 256 mm. The surface of plot 14 had a sparse (several percent) cover of gravel with sizes up to 30 mm, and gravel up to 10 mm in diameter covered about 1–2 percent of plot 13.

** Upper 3 cm only. Bulk density of zone below is 1.13/g/cm³.

bushland (PRATT et al. 1966). The Vertisols and other fine-textured soils of the Athi-Kapiti Plains are covered mainly by grasslands with a sparse (< 1 percent) canopy cover of *Acacia drepanolobium*. The drier country to the southeast is covered with bushland and woodland of larger acacias and other species. The canopy here ranges from less than 1 percent on heavily settled ridgetops to about 30 percent on a few steep hills far from a water supply.

The ground cover on all these areas is clumped and varies seasonally with rainfall and grazing pressure. On the Athi-Kapiti Plains, the ground cover, measured with a pin frame (GREIG-SMITH 1957) at the time of the experiments, varied from 64 to 78 percent (table 1). Basal cover, measured after clipping vegetation down to the ground surface, ranged from 32 to 36 percent. At the Amboseli and Kimana sites, the ground cover had already been grazed down to ground level by the time of the experiments and for all but one site varied between zero and 12 percent (table 1).

The experiments

Field equipment

Details of the plot construction and of the rainfall simulator have been described elsewhere (DUNNE 1977, DUNNE et al., in press). Each plot had a length of 5 m and a width of either 1.2 m or 2.0 m. It was surrounded by a sheet-metal boundary, installed carefully and sealed. At the lower end of the plot a concrete apron linked the soil surface to a collection trough, the outflow from which was monitored. The rainfall simulator consisted of a wooden frame, 3 m high by 2 m wide and 5.3 m long, supporting a metal track along the center of the frame. A single nozzle was mounted on a trolley in the track and was moved rapidly back and forth over the plot to reduce the rainfall intensity to realistic levels while generating raindrops of approximately natural size. Runoff from the plot fluctuated slightly due to this effect and to errors in measuring the flow rate, but each hydrograph was well defined. Rainstorm kinetic energies averaged about 17 J/m²/mm for all storm intensities and Christiansens's uniformity coefficient for rainfall distribution on the plot exceeded 75 percent (DUNNE et al., in press).

Experimental design

We chose hillside plots at typical sites in the three geologic subdivisions and on each plot we measured infiltration and runoff resulting from the application of rainstorms of one-hour duration to soils in various conditions of antecedent moisture and vegetation cover on a range of hillslope gradients (table 1).

On each plot, the early experiments were conducted with a nozzle which generated raindrops with a median diameter of 2.0 mm at an average rainfall intensity of 70 mm/hr (and a standard deviation of 13 mm/hr between experiments). In Kajiado District, such a rainstorm has a recurrence interval of about 4–5 years (interpolated from data in TAYLOR & LAWES 1971: 19, DUNNE & LEOPOLD 1978: 64–65). The first simulated rainstorm (called the "dry run") was

applied to the soil with an antecedent moisture content at or below the permanent wilting point and with the ground cover at its typical late-dry-season density, listed in table 1.

The second experiment (called the "wet run") occurred one-quarter to two days after the first one. The delay allowed time for the topsoil to drain nearly to field capacity. No measurements of moisture content were made, but the soils were in a condition which characterizes most of the wet season when rain occurs frequently.

On four of the 11 sites the third experiment (called the "clipped run") occurred several hours to one day after the second one. It was conducted with the same nozzle and intensity as the first two experiments, but before application the grass cover was clipped as close to the ground as possible. These four sites (table 1) were the only plots on which the ground cover could be reduced. Elsewhere, the cover was so sparse (zero to 12 percent) that only root tops survived before the dry run. Even on the heavily vegetated plots, the cover could only be reduced to one-half of its original value before the hard stalk and dense basal parts of the plants were exposed. The antecedent moisture content was between field capacity and saturation, so that this experiment simulated the effects of a late-wet-season storm with a recurrence interval of about 5 years on heavily grazed, wet ground.

The final artificial rainstorm at each site was applied with a larger nozzle, which generated raindrops with a median diameter of 2.7 mm, at an average intensity of 132 mm/hr (± 30 mm/hr). Such a storm has a recurrence interval of approximately 100 years in Kajiado District (estimated from the sources mentioned above). These final experiments were designed to simulate the effects of a very large storm on wet soils with their lowest ground cover.

Infiltration

Sandy clay loams

a) Experiments using 2.0 mm diameter raindrops

Runoff from all plots on the sandy clay loams began less than 10 minutes after the onset of rainfall except on plot 12 which had a stony cover (see footnote in table 1). The delay was shorter in the later experiments (table 2). Infiltration capacity was calculated as the difference between the rates of rainfall and runoff, ignoring changes in detention storage. During each storm the infiltration

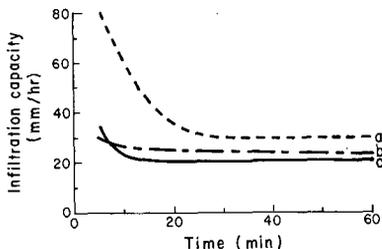


Fig. 2. Infiltration curves for successive artificial storms on plot 14 including the "dry run" (a), "wet run" (b) and "intense run" (c) referred to in table 2. The wet run occurred 17 hours after the dry run and the intense run followed the wet run by one hour.

capacity rapidly decreased to an approximately constant rate within 30 minutes at the most. As an example the infiltration curves for the consecutive experiments on plot 14 are shown in fig. 2.

The general form of the infiltration curves and the variations between consecutive experiments can be explained in terms of soil physics theory and our field observations. Apart from the relatively small influence of the accumulation of detention storage (which was measured with a thin scale), three processes contributed to the diminished infiltration capacity during a storm and in successive storms: the swelling of colloids; the decrease in potential gradient as the wetting front penetrated to greater depths; and the development of a surface crust.

The swelling of clay in the soil closes surface cracks and pores. An index of clay swelling during the dry run on plots 10 and 11 was measured by the change in elevation of the soil surface against ten stakes. The surface swelled by an average of 2.1 mm during the first experiment. Measurements in later experiments showed lower values, but the decrease could have resulted from slight movement of the stakes.

The influence of decreasing potential gradient as the wetting front penetrates to greater depths (FREEZE 1974) probably had an influence on the form of the infiltration curves early in the first experiments, before a continuous surface crust had developed. However, the soil physics theory that explains the declining infiltration capacity also predicts that the topsoil will be saturated to the depth of the wetting front when overland flow occurs. Immediately after the final experiment, we dug a hole in each plot and examined the soil. Our visual observations, as well as calculations of the moisture content from measured wetting-front depths, total volume of infiltrated water, and bulk density, indicated that the soil was not saturated (although the saturation percentages exceeded 80 per cent). Only the upper few millimetres of surface crust were saturated, except on plot 12 where a gravel cover protected the soil from crusting and infiltration was limited by a 3 cm-thick saturated platy horizon immediately below the gravel.

A third mechanism responsible for the decline of infiltration capacity during and between experiments was the development or augmentation of the surface crust referred to in the earlier description of soils. Several investigators (e. g., DULEY 1939, McINTYRE 1958 a, b, FARRIS 1978) have demonstrated experimentally that when bare soils are exposed to raindrop impact a thin surface crust is formed by dispersed fine particles which form a less permeable layer one to several millimetres thick. Soil aggregates are dispersed by slaking and bombardment and are washed into pores below. McINTYRE (1958 a) recognized a 0.1 mm-thick surface seal above a "washed-in" zone that was 1.5–3.0 mm thick. His laboratory measurements demonstrated that the underlying sandy loam had a saturated hydraulic conductivity approximately 200 times that of the washed-in zone and 2000 times that of the surface seal. FARRIS (1978) and McINTYRE (1958 a) documented in the laboratory the rate at which such a crust thickens and becomes less permeable during a rainstorm.

On the Kenyan hillsides we observed such crusts after the first experiment on each plot. In the early stages of the first experiment raindrops cratered the loose soil and broke down aggregates. The undisturbed crust that had survived

from the preceding rainstorm was not broken down by raindrop impacts. The change in the condition of the soil surface was most obvious after the first experiments during which the largest declines in infiltration capacity were measured. During the second and third experiments the initial infiltration rate was approximately equal to that at the end of the dry run. Dispersion of aggregates on the surface continued in the early stages of the wet run when the layer of overland flow was thin, but the effect of bombardment by drops declined during the second experiments (and during the clipped run on plot 11) as the layer of overland flow became thicker.

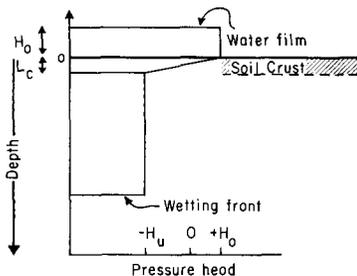


Fig. 3. Sketches defining the terms used in eq. 1.

The average saturated hydraulic conductivity (K_c) and thickness (L_c) of the crust (fig. 3) limit the final constant rate of infiltration (i) in the following manner (HILLEL & GARDNER 1969):

$$(1) \quad i = \frac{K_c (H_0 + L_c - H_u)}{L_c}$$

where H_0 is the positive hydraulic head imposed on the soil surface by ponded overland flow, and H_u is the pressure head just below the crust. The last value in turn depends upon the permeability-moisture-pressure relationship for the underlying soil. We were not able to make the necessary measurements of H_u to evaluate K_c from eq. 1, but future field experiments on infiltration into crusted soils need to incorporate measurements of the length terms in that equation in order to evaluate K_c and its temporal variation.

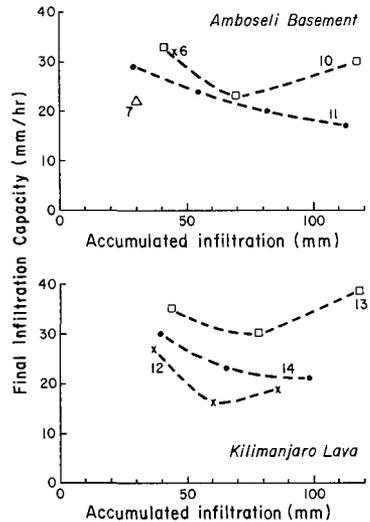
The shapes of the infiltration curves and the final rates for each experiment (fig. 4) were broadly similar for all of the plots underlain by sandy clay loams. Infiltration declined during the dry run for the three reasons outlined above, and at the end of the set of dry runs the final capacity averaged 29 mm/hr (with a range of 11 mm/hr) on the four Basement plots and 31 mm/hr (27–35 mm/hr) on soils derived from lava. If the seven values are consolidated, an overall mean of 30 mm/hr with a standard error of 1.6 mm/hr characterizes the initially dry sandy clay loams. During the wet runs (and the clipped run on plot 11) swelling and crusting continued (few aggregates remained on the surface at the end of the wet run) but at a decreasing rate. An overall mean infiltration capacity of 23 mm/hr ($s_{\bar{x}} = 2.2$ mm/hr) resulted from the five wet runs.

b) Experiments using 2.7 mm diameter raindrops

In the experiments with the larger nozzle the details of the results were con-

fusing. Although the final infiltration rates remained similar to those from the previous experiments, the values were higher in the final applications on three of the plots (fig. 4). Four reasons suggest themselves for this surprising result. The first is experimental error in the definition of runoff curves because fluctuations in rainfall and runoff were accentuated when the larger spray nozzle was used for the heavier application. This factor could explain the small apparent increase in infiltration capacity on plot 12, but it could not have been responsible for the magnitude of the observed changes on plots 10 and 13.

Fig. 4. Final infiltration capacities at the end of each artificial rainstorm as a function of accumulated infiltration on the sandy clay loams. Numbers refer to plots listed in tables 1 and 2. Only plot 11 had sufficient initial ground cover that a "clipped run" could be performed.



A second possibility is that the greater depth of runoff under the more intense rainfall provided a larger pressure potential which increased the force driving water through the soil surface, in effect, increasing H_o in eq. 1. However, the pore pressure beneath the crust is likely to be large relative to any realistic variation of H_o . For example, from table 2, the saturated hydraulic conductivity of soil on plot 13 must be at least 39 mm/hr. If we accept the earlier-mentioned result of McINTYRE (1958 a) that K_c was 0.5 % of the saturated hydraulic conductivity of the underlying sandy loam, and substitute a value of $K_c = 0.2$ mm/hr into eq. 1, then at the end of the wet run on that plot, $i = 30$ mm/hr and $L_c = 2$ mm imply that $(H_o + H_u) = 298$ mm, and since H_o on that plot ranged only between zero and 5 mm, its variation cannot significantly affect the infiltration capacity even if the original estimate of K_c contains a relatively large error.

MOLDENHAUER et al. (1960) suggested a third reason for an increase in infiltration capacity with rainfall intensity. As the depth of runoff increases, a greater proportion of the small topographic irregularities on the ground surface are flooded and the area available for infiltration increases. On plot 10, for example, only 70 percent of the ground surface was flooded at the end of the

Table 2. Rainfall, runoff, and infiltration data for the plots.

| Plot No. | Run | Rainfall Intensity (mm/hr) | Volume of Runoff (mm) | Final Infiltration Capacity (mm/hr) | Time to Onset of Runoff (min) |
|------------------|---------|----------------------------|-----------------------|-------------------------------------|-------------------------------|
| Sandy clay loams | | | | | |
| 6 | Dry | 79 | 34 | 32 | 6 |
| 7 | Dry | 69 | 39 | 22 | 5 |
| 10 | Dry | 73 | 32 | 33 | 7 |
| | Wet | 63 | 35 | 23 | 4 |
| | Intense | 166 | 118 | 30 | 1 |
| 11 | Dry | 64 | 35 | 29 | 4.5 |
| | Wet | 54 | 28 | 24 | 3 |
| | Clipped | 60 | 33 | 20 | 4 |
| | Intense | 127 | 96 | 17 | 1 |
| 12 | Dry | 45 | 8 | 27 | 19 |
| | Wet | 64 | 40 | 16 | 4 |
| | Intense | 107 | 82 | 19 | 1 |
| 13 | Dry | 71 | 27 | 35 | 9.5 |
| | Wet | 75 | 40 | 30 | 4 |
| | Intense | 123 | 83 | 39 | 3 |
| 14 | Dry | 82 | 42 | 30 | 5 |
| | Wet | 76 | 50 | 23 | 2.5 |
| | Intense | 111 | 78 | 21 | 1 |
| Cracking clays | | | | | |
| 2 | Dry | 57 | 1 | 51 | 32 |
| | Wet | 50 | 15 | 25 | 10 |
| | Clipped | 63 | 29 | 4 | 2 |
| | Intense | 141 | 111 | 3 | 1.5 |
| 3 | Dry | 81 | 4 | 71 | 40 |
| | Wet | 82 | 29 | 40 | 7 |
| | Clipped | 74 | 46 | 13 | 6 |
| | Intense | 96 | 80 | 10 | 3 |
| 4 | Dry | 78 | 1 | 76 | 48 |
| 5 | Dry | 106 | 4 | 91 | 27 |
| | Wet | 74 | 12 | 55 | 5 |
| | Clipped | 79 | 31 | 30 | 4.5 |
| | Intense | 183 | 141 | 2 | 1 |

dry run. This figure increased to 90 percent at the end of the second experiment, and 100 percent late in the third. On the other, smoother plots, however, there was no significant change between the second and third experiments. The proposed mechanism did not affect infiltration significantly in these studies.

A fourth possible reason for the increased infiltration capacity is that the heavier rainfall and runoff intensities and the larger raindrop size scoured the thin surface skin and increased the permeability of the surface crust. Most investigators (e. g., MCINTYRE 1958 a, fig. 4; BISAL 1967) conclude that the magnitude of raindrop kinetic energy affects the development of soil crusts and have attributed decreases in infiltration capacity to increases in energy application and soil dispersal. In our field experiments there were some important distinctions

from the conditions under which the frequently-quoted laboratory experiments were conducted. During the intense run, the rate, velocity and depth of runoff were greater than in the laboratory experiments or during our earlier field experiments. The rates of rainsplash and of erosion were much more intense (DUNNE 1977). Slaking and bombardment of aggregates and remobilization of fine particles already on the surface occurred, but a large amount of fine material was washed from the plot. The soil surface was thoroughly disturbed and experienced intense rates of bedload transport. Fine particles were winnowed out of the moving layer of sediment and a coarse lag of sand and water-stable aggregates was left behind at the end of the experiment. Silt and clay were lifted into the sheetwash, sometimes with an aerial trajectory. Most of the particles fine enough to cause crusting, therefore, could have been washed away by the intense sheetwash. At the end of each intense run patches of exposed crust frequently appeared to have been scoured and the skin seals on their surfaces were replaced by an armor layer of coarse sand.

McINTYRE (1958 a, b) showed that if dispersed fine particles do not remain on the soil surface the surface crust does not develop to the same extent. He also demonstrated that soils which experience the greatest splash for a given rainfall energy seal the least. Compaction and the washing-in of fines are necessary for the development of a crust. If the surface skin cannot develop because of turbulence and high shear stress, or if an existing crust is scoured, the conductivity of the crust should be higher than in the less intense storms.

On two of the plots, of course, there was no reversal of trend in the final experiment (fig. 4). On these plots (11 and 14) the surface crusts were much more strongly developed before the experiments, and survived the shear stress imposed during the intense run. On plot 11 the depth of runoff was less than elsewhere. In all cases the processes of crust destruction and formation oppose one another and the net result will depend upon detailed characteristics of each soil, which could not be measured in this study.

c) Variations downslope

In addition to the broad similarity in the infiltration capacities of the sandy clay loams, fig. 4 suggests a general increase in infiltration capacity downslope on each rock type, although each set of plots is not distributed along a single hillslope profile. On the Amboseli Basement, plot 11 (with a gradient of 0.05) represents the steeper upper parts of a few hillsides, that are grazed lightly because of their distance from water. Plots 6, 7, and 10 lie adjacent to one another about 300 m from a ridgetop and have a gradient of 0.02. The relatively thin sheetwash on a steep slope near a ridge top should leave the soil surface on plot 11 more exposed to raindrop impact than the surface further downslope, and the limited trampling by animals would allow the crust to remain intact. These factors may explain the lower infiltration capacity on the steeper slope. On the Kilimanjaro Lava, plot 12 is representative of gravel-covered soils within 100 m of bedrock outcrops. YAIR & LAVÉE (1976) have explained how a dense cover of gravel concentrates water onto a fraction of the plot surface, increasing the average water yield. Plot 14 represents conditions about 100 m further downslope, while plot 13 typifies the footslopes.

d) Variations with vegetation cover

There was no obvious effect of vegetation density on infiltration over the narrow range of vegetation cover sampled during these experiments. On plot 11 for example the decline in infiltration capacity between the second and third runs was probably associated with surface crusting as well as with a decrease in cover from 41 to 16 percent. The infiltration capacity on this plot declined further during the fourth experiment while cover conditions remained constant. Furthermore, the well-developed crust and higher bulk density on plot 11 were associated with a lower infiltration capacity than on plot 10, even though the former had a much denser grass cover. These data suggest that in some rangelands, changes of infiltration capacity result only from large changes in ground cover density. JOHNSON & NIEDERHOF (1941) and MARSTON (1952), working on soils with high infiltration capacities and a narrow range of cover density, also failed to discover any simple relationship between infiltration and vegetation cover. The data of KINCAID et al. (1963) similarly show little if any increase of infiltration capacity until the cover density exceeds about 20 percent. MOORE et al. (in press) found no difference between the infiltration capacities of clay and sandy clay soils in Machakos District, Kenya, for sites with zero and 57 percent basal cover, although ploughing of the surface strongly increased the infiltration capacity. However, SMITH & LEOPOLD (1942) and DORTIGNAC & LOVE (1961), working in the American West, documented significant changes in infiltration with modest changes in vegetation density. Further experiments are needed to clarify the physics of this effect and the circumstances under which it becomes important.

Cracking clay soils

Water entered the clays on the Athi-Kapiti Plains by two routes: by free gravity flow into the system of large cracks ($\cong 2$ cm) between prisms and smaller cracks ($\cong 2$ mm) between peds; and by flow within the porous peds influenced by gravitational, capillary and osmotic forces. During the early experiments infiltration was dominated by the first process, but as the experimental sequence proceeded the relative importance of the second process increased. Water could only flow into large cracks when it was under positive pressure, either as free water on the soil surface or after some horizon in the soil had become saturated. This water "short-circuited" the upper horizon by flowing down the outer surfaces of peds, which absorbed a portion of the water as indicated by the tracing experiments by BLAKE et al. (1973), BOUMA & DEKKER (1978) and BOUMA et al. (1978). At the end of the experiments water was found to have penetrated along cracks to depths of 55–90 cm, but the insides of many peds were dry.

During the experiments on dry soils, rainfall intensity soon exceeded the conductivity of the peds and water began to accumulate on the surface after 12 minutes on the aggregated topsoil of the Vertic Rendzina on plot 2 and after 3 or 4 minutes on the three Vertisols. This water flowed into cracks, with one large crack system on plot 3 absorbing all of the runoff from about 2 m² throughout the first storm. Runoff from the plots only occurred after half an hour or more. As the dry run proceeded, the infiltration rate decreased slowly, as shown in fig. 5. Throughout most of the first experiment, therefore, the infiltration ca-

capacity of these soils exceeded 60 to 100 mm/hr (table 2), although it will later be apparent that the saturated hydraulic conductivity of the peds themselves was less than 10 mm/hr.

In the later experiments the lag between the onset of rainfall and runoff declined strongly (table 2), and there was a marked decrease in the final, constant infiltration capacity (fig. 5), with the smallest change occurring between the third

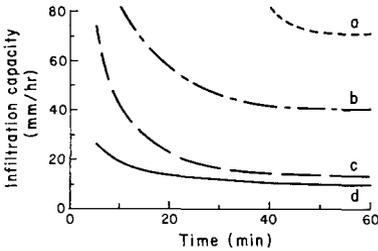


Fig. 5. Infiltration curves for successive artificial storms on plot 3 including the "dry run" (a), "wet run" (b), "clipped run" (c), and "intense run" (d). Although the ground cover was reduced from 76 percent to 36 percent between the second and third experiments, the dominant process causing reduced infiltration was crack closure. Run (b) occurred 20 hours after (a); run (c) followed (b) by 24 hours; and run (d) was 4 hours after (c).

and fourth experiments. During each experiment on the Athi-Kapiti Plains several processes combined to decrease the infiltration capacity. The 1 cm-thick, loose granular surface layer was dispersed and a surface seal developed as described previously. Because the depths of overland flow on these rough plots were great (5–10 mm), the surface seal developed less quickly than on the sandy clay loams, but was obvious after the third and fourth experiments. Infiltration into peds also decreased because of the declining head gradient (FREEZE 1974) as each became wetter. A third important process was the erosion of the lips of the cracks. Soil transported into cracks by this process swelled as it became wet and partially blocked the narrowing cracks. Finally, swelling of the clays (which raised the soil surface by 7–12 mm during the first experiment alone) closed the cracks quickly during the dry run and more slowly thereafter. Slight expansion occurred even after the end of the fourth experiment when some of the soil that had been eroded from the lips of cracks and had expanded as it became wet was extruded above the soil surface.

Of these four processes the dominant control on the infiltration capacity was the evolution of the cracks. This is clearly shown by the consistently lower infiltration capacities on the less-cracked Vertic Rendzina on the upper portion of the hillslope (plot 2) than on the Vertisols of the footslope (fig. 6). Even

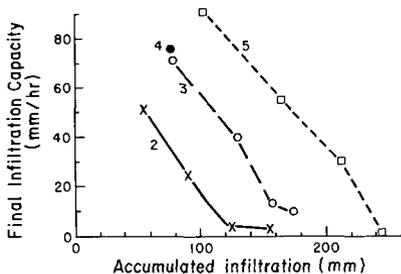


Fig. 6. Final infiltration capacities at the end of each artificial rainstorm as a function of accumulated infiltration on the cracking clay soils. Numbers refer to plots listed in tables 1 and 2. The soil on plot 2, which is typical of the upper parts of slopes, was less cracked than the footslope soils represented by plots 3, 4 and 5.

during the final experiment drainage into narrow cracks on both soil types maintained the final infiltration capacities at 2–10 mm/hr, which exceeds by about an order of magnitude the saturated conductivities of clay loam soils as reported by FREEZE (1972). Because of the dominance of the cracks, changes in infiltration capacity caused by reducing the ground cover were not detected.

Frequency of precipitation excess

There are no long-term records of runoff from plots or small catchments in the grazing lands of Kenya, and it is difficult, therefore, to compute the frequency and magnitude of Horton overland flow. Preliminary estimates of precipitation excess can be made from a comparison of rainfall intensity records with measured infiltration capacities. LAWES (1974) has published frequencies of rainfalls of various durations and intensities for 17 stations in Kenya. In order to apply these frequencies to the local study sites, the frequency of rainstorms with a particular magnitude and duration was correlated with the mean annual precipitation at each meteorological station (East African Meteorological Department 1966). The relationship was then used to estimate rainstorm magnitude and frequency at the experimental sites. The results for all rainstorm intensities and duration intervals available are listed in table 3. No data are available for rainfall bursts shorter than 15 minutes.

Table 3. Estimated average frequency (times per year) of the occurrence of rainfall intensities for storms of various durations at the three experimental sites.

| Intensity (mm/hr) | Duration (hr) | | | | |
|--------------------------|---------------|-----------|----------|----------|----------|
| | 0.25–0.49 | 0.50–0.99 | 1.0–1.99 | 2.0–2.99 | 3.0–2.99 |
| Athi-Kapiti Sites | | | | | |
| 25 – 29.9 | 2.8 | 3.0 | 0.1 | 0.7 | 0.1 |
| 30 – 39.9 | 3.6 | 1.8 | 0.6 | 0.1 | |
| 40 – 59.9 | 2.8 | 0.7 | 0.4 | | |
| 60 – 79.9 | 0.7 | 0.5 | | | |
| 80 – 111.9 | 0.4 | | | | |
| Kimana Sites | | | | | |
| 25 – 29.9 | 1.3 | 1.3 | 1.0 | 0.2 | 0.05 |
| 30 – 39.9 | 2.1 | 1.5 | 0 | | |
| 40 – 59.9 | 1.5 | 0.4 | 0.25 | | |
| 60 – 79.9 | 0.4 | 0.3 | | | |
| 80 – 111.9 | 0.3 | | | | |
| Amboseli Sites | | | | | |
| 25 – 29.9 | 1.1 | 1.2 | 0.2 | 0.1 | 0.03 |
| 30 – 39.9 | 1.3 | 0.6 | 0.3 | | |
| 40 – 59.9 | 1.0 | 0.3 | 0.2 | | |
| 60 – 79.9 | 0.3 | 0.2 | | | |
| 80 – 111.9 | 0.2 | | | | |

Sandy clay loams

Figure 2 and the other results from the sandy clay loams at Amboseli and Kimana indicate that infiltration capacity attains an approximately constant value within a few minutes after the onset of rainfall. Mass curves for typical rainstorms in Kenya (FIDDES 1975) and elsewhere (U. S. Soil Conservation Service 1972) indicate that the maximum intensity during a storm does not usually occur at the onset of rainfall. Therefore, to estimate the frequency of overland flow, it is appropriate to compare the intensities in table 3 with the final constant infiltration capacities for the sandy clay loams.

Figure 4 indicates that the final infiltration capacities of these soils do not exhibit major seasonal trends, and that of the 11 experiments on wet antecedent conditions, 8 yielded a value between 15 and 25 mm/hr, while in the other three experiments and the 7 dry runs the results usually lay between 20 and 30 mm/hr. Such variability is to be expected because of subtle differences in soil properties, and is the basis of the partial-area concept of storm runoff (BETSON 1964, DUNNE 1978, YAIR et al. 1978) which emphasizes that the generation of Horton overland flow is spatially variable to a greater extent than is commonly realized. However, because of the dearth of precise information on rainfall intensities in southern Kenya, a detailed comparison with infiltration capacities is not yet warranted. Instead, we have assumed an average capacity of 30 mm/hr (close to the mean of 29.7 mm/hr) for dry antecedent conditions, and 25 mm/hr (close to the average of 25.8 mm/hr) for all experiments on wet antecedent conditions. Plotting total runoff volume against rainfall (from table 2) for the one-hour storms also indicates threshold rainfalls of about 25 mm and 30 mm necessary to generate runoff under wet and dry antecedent conditions respectively. The infiltration capacity should vary between these two values during a wet season depending on the time since the preceding storm.

Table 3 and the measured infiltration capacities indicate that runoff is a relatively rare event on these rapidly eroding hillsides, even if the antecedent moisture conditions are always wet. The Kimana hillsides have a slightly higher frequency of runoff than those at Amboseli, although the influence of heavier rainfall may be offset by higher infiltration capacities on hillslopes similar to plot 13 (fig. 4). In fact, the variation of infiltration capacity along the hillsides and the consequent shedding of runoff from upper slopes onto more permeable footslopes adds another complication, the effect of which is probably less than uncertainties in the rainfall data. Subtle differences of this magnitude could only be resolved by extensive monitoring. At the level of approximation warranted by our field data and the scanty rainfall data, the low runoff frequencies on the two landscapes are similar. THORNES (1967) also concluded that runoff was rare on rapidly eroding areas of southern Spain.

The average annual precipitation excess at each site can be estimated from the rainfall intensities in table 3 and the average values of infiltration capacity for the dry and wet runs. If soils were always dry, the long-term mean precipitation excess at the Kimana sites would be 57 mm/year (13 percent of rainfall) and at Amboseli it would be 40 mm/year (13 percent of rainfall). If the soils were always wet, the results would be 84 mm/year (19 percent of rainfall)

and 59 mm/year (20 percent) respectively. These values fall within the scatter of points when mean annual rainfall is plotted against mean annual runoff from small catchments with a variety of soil types in semi-arid regions of the United States (LUSBY 1978; U. S. Department of Agriculture 1978). However, we will argue in the companion paper that on the long, gentle Kenya hillslopes most of the precipitation excess generated during a rainstorm infiltrates after the storm as water flows downslope. Channel runoff volumes are much lower than the values of precipitation excess given above, and more water enters the footslope soils than the soils further upslope. Water stored in the hillslope soils by infiltration is available for evapotranspiration, and therefore for sustaining primary production. The amounts of runoff shed from the hillsides represent an upper limit on the increase in soil moisture that could be attained if the infiltration could be raised by mechanical or vegetational means. The empirical relationship between evapotranspiration and net above-ground primary production presented by ROSENZWEIG (1968) suggests that if the 13 to 20 percent of rainfall that presently does not enter the soils in the upper portion of each hillside were to infiltrate, the average annual primary production could be increased by 26 to 45 percent. However, we stress that this is an approximate calculation about limits only, and the economic, technological, and ecologic aspects of the problem have not been addressed. The increased soil moisture recharge caused by the greater duration of runoff on footslopes must also sustain higher primary production there, but the calculation of this spatial pattern of infiltration requires the calculation of runoff hydrographs in later papers.

Cracking clay soils

Estimation of the frequency and amount of precipitation excess from the cracking clay soils of the Athi-Kapiti Plains is a much more complicated problem because of the dramatic decrease of infiltration capacity as the soils become wet (fig. 6). Short dry periods during the rainy season should reverse this trend somewhat, which further complicates the situation. At the beginning of a wet season, the Vertic Rendzina on plot 2, which is representative of the upper sections of hillsides, can absorb almost all of the rainstorms listed in table 3. Only two or three 15- to 30-minute bursts and one 30- to 60-minute burst per year and less frequent 1- to 2-hour storms could generate runoff if they were to occur at that time. The Vertic Rendzina becomes almost impermeable as the cumulative water storage above the permanent wilting point increases up to 120 mm, which amounts to 27 percent of the average precipitation of the main wet season or 65 percent of rainfall in the subsidiary wet season (East African Meteorological Department 1966). Figure 6 indicates that the Vertisols can absorb almost all rainstorms at the beginning of the wet season, but rainfall and recharge by runoff from upslope render them almost impervious after about 200 mm of infiltration.

Before the frequency, rates, and total amount of precipitation excess from the cracking soils can be predicted it will be necessary to define the seasonal variation of infiltration capacity and to use a long record of individual storms and their timing. The problem is complicated by the general downslope increase of crack density and width, and therefore infiltration capacity, which produces

yet another manifestation of partial-area contributions of Horton overland flow. YAIR *et al.* (1978) recognized a similar situation where bedrock outcrops shed large amounts of runoff onto a coarse colluvial mantle on hillslopes in the Negev Desert.

Cracking clay soils also occur in the swales between hillslopes in the Amboseli and Kimana regions. In the early part of the wet season infiltration into these deeply cracked soils causes hillslope runoff to be retained within the catchment. Once swelling of the clay has sufficiently reduced the infiltration capacities of these soils or when the soils become saturated, runoff out of the catchment will occur. During the late wet season these swales become regions of saturation overland flow (DUNNE 1978).

The cracking clay soils in the swales are also important sources of grass production. An estimate of the relative total annual primary production from hillslope soils and swale soils based on the total amount of water that can be withdrawn from each during evapotranspiration can be made from the empirical relationship reported by ROSENZWEIG (1968). On the upper parts of the Amboseli hillslopes roughly 250 mm of the 300 mm of annual rainfall infiltrate and become available for plant consumption; at Kimana roughly 380 mm are absorbed on the upper slopes. If the one-metre-deep cracking clays have an available water capacity of 300 mm and attain saturation twice each year, then at least 600 mm of water will be available for evapotranspiration. According to ROSENZWEIG's relationship, the plant production will be respectively more than 4 and 2 times greater in the swales than on the hillslopes. Importantly, this productivity will extend well beyond the end of the wet season and the exhaustion of available soil water on the hillslopes. Thus, although the swales occupy less than one per cent of the landscape they strongly influence the soil hydrology and grass productivity of the region.

Continued work on the spatial and temporal variation in infiltration capacity and frequency of runoff will contribute to the construction of a quantitative model for runoff, erosion and productivity of the intensively grazed rangelands of Kenya. For the model to be applicable on a geomorphic time scale the interdependency of erosion processes and soil properties must be defined.

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