The geometry of geomorphological forms as related to Late Pleistocene stratigraphy,

by R. PAEPE [1].

1. INTRODUCTION.

Increasing importance of quaternary research tends to diminish the historically grown gap between geomorphology and stratigraphy. While geomorphology, at least in its earliest days, was basically a descriptive method aiming to the study of landforms, quaternary geology was merely based on stratigraphical principles similar to the ones used in other fields of the geologic record. The use of a stratigraphical background in geomorphology was most often limited to the explanation of morphological features with respect to the nature of the substratum. On the other hand, within the scope of the stratigraphical sequence, environmental processes and forms were hardly concerned. Evidently, one and another are closely related, especially within the framework of continental quaternary features. Sedimentation and landscape erosion are contemporaneous processes, the effect of which is felt both ways. Slope development e.g. is controlled by the rate of evacuation of screes at its toe-slope or of the degree and nature of the weathering mantle (R. SOUCHEZ, 1966) while evolution in the upper part of a slope may lead to erosion of newly built deposits in the lower profile part. Also, through time, there may be periods during which accumulation is rather scarce while erosional processes are of a much greater importance, in both intensity and areal extent. It is appropriate to this paper to raise up geomorphology as a useful tool in stratigraphy, both qualitatively and quantitatively. Previously, many attempts in this direction have been undertaken (J. B. BAKKER, A. STRAHLER, A. E. SCHEIDEGGER, E. RICHTER, W. C. KRUMBEIN, etc.) but mostly on basis of theoretical considerations. Since a short time there exists a growing tendency to start mathematical quantitative analysis from empirical deductions (R. V. RUHE, R. SOUCHEZ, R. PAEPE). Empirical analysis was probably preceeded by theoretical concepts mainly as a result of the Davision descriptive-explanatory method based on natural didactical landforms. It is our conviction, however, that the impact of a mathematical-quantitative approach of geomorphology

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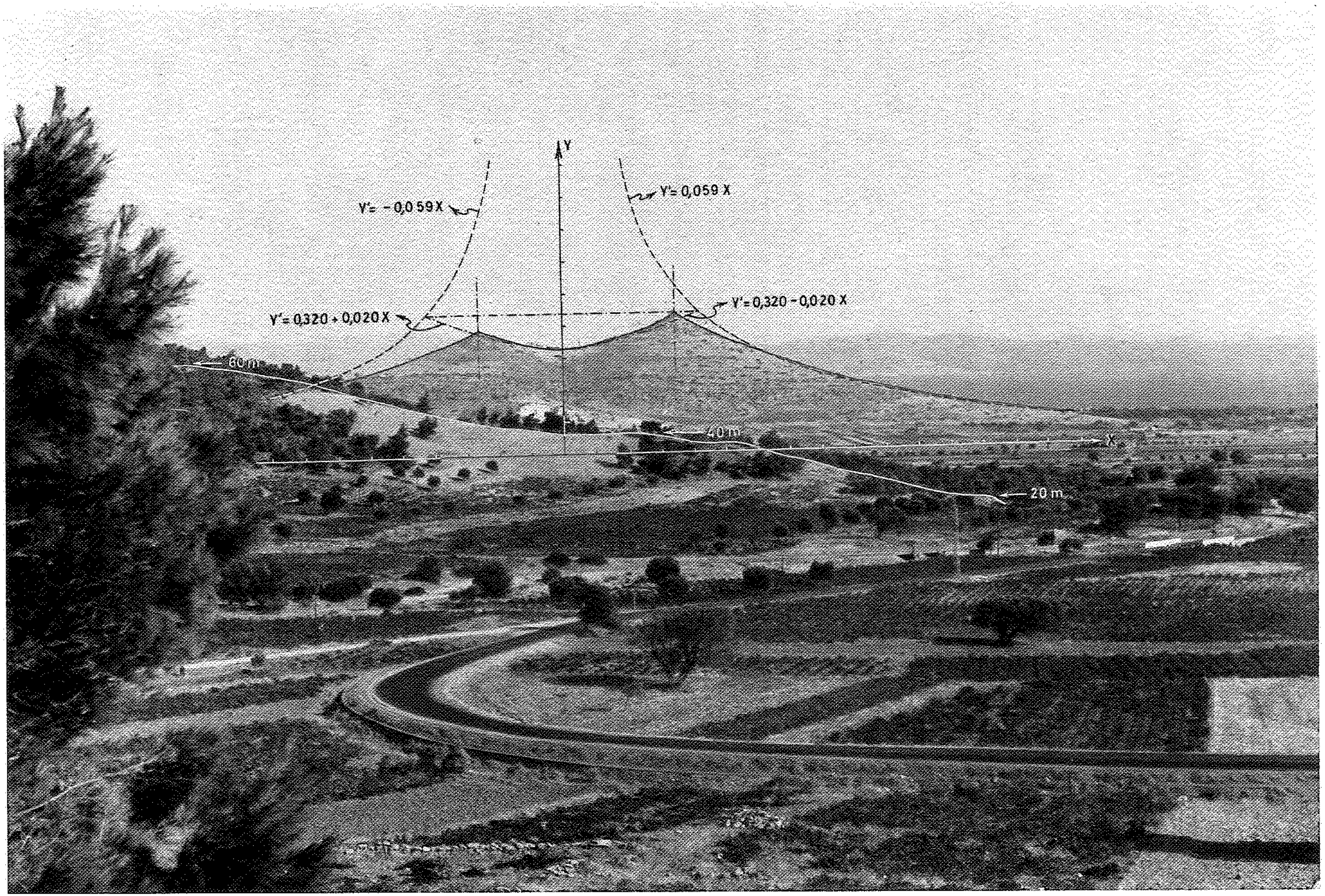

FIG. 1.

is the existing landscape forms. Then only, geomorphology may become a usefull tool of stratigraphy. In the following a distinction in the field is made between surficial (landscape) forms (external forms) as well as morphological features inherent of the deposits (internal forms).

2. EXTERNAL FORMS.

2.1. Mathematical formulation of geodynamic profiles.

The first question that has to be solved is the validity or not of mathematical expressions as a proof for a dynamic geomorphological equilibrium of a continuous slope profile.

The region of Lavrion in Southern Attica (Greece) offers a landscape sown with perfect inselbergs of which the double-peaked, conical shaped Vélatouri complex (fig. 1), dominating the bay of Thorikos, is a most representative one (R. PAEPE, 1968). Bedrock is composed of a heterogeneous complex of shales, limestones and igneous rocks, so that structural controll in the shaping of the slopes is excluded. From the sea (+5 m) the southern wall rises without interruption up to the highest peak (144 m) of the inselberg, describing a concave profile with steady growing curvature. But for a sporadically thin layer of detritus and some exceptional large boulders, bedrock crops out everywhere, so that it can be assumed that no deposition whatsoever, occurs along its surface. We thus deal here with an ideal erosional form the evolution of which has not been disturbed by subsequent mantling. In accordance with statements made by R. SOUCHEZ, 1966, the very upper part represents the zone of removal, while the lower part, down from about 120 m, is dealing with the zone of transport. The whole profile has reached a geodynamic equilibrium. Computing the latters mathematical form, this slope profile reads as :

$$Y' = 0{,}059\,X \qquad (1)$$

where $Y' = Y^{-1}$.

From the graph (fig. 1) we can readily see that the calculated form fits fairly well with the natural profile. Furthermore, expression (1) may be also written as :

$$XY = k \qquad (2)$$

which is the general equation of a hyperbola referred to its asymptotes, and where also the asymptotes are used as the coordinate axes. The Y-axis now is found to run through the basal part of a hanging valley, diametrical opposite to the southern wall and between the two peaks. This position is also the geometrical mid-point or line of asymmetry between the two peaks so that the conjugate hyperbola of (1):

$$Y' = -0{,}059\,X \tag{3}$$

tallies with the northern wall profile at least in its lower part; for stronger denudation of the peak lowered the profile in its upper part (100 m).

Next, separate computation of both members of the hanging valley, between the two peaks, on either side of the Y-axis, leads to the formulation of two other hyperbola:

$$Y' = 0{,}320 - 0{,}020\,X \tag{4}$$

and

$$Y' = 0{,}320 + 0{,}020\,X \tag{5}$$

In both the latter cases asymptotes are rectangular but shifted parallelly with respect to the coordinate axes; they are respectively:

$$X = 0;\quad Y = \frac{1}{0{,}32} = 3{,}125, \qquad X = 0;\quad Y = \frac{1}{0{,}32} = 3{,}125$$

and

$$Y = \infty;\quad X = \frac{0{,}32}{0{,}02} = 16, \qquad Y = \infty;\quad X = -\frac{0{,}32}{0{,}02} = -16.$$

It must be noticed that the asymptote at $Y = 3{,}125$ is tangent to the basal and midpoint of the curvilinear profile of the hanging valley. It can be assumed that this is the base level of erosion of the hanging valley as erosion-surfaces at that level do generally occur in the surroundings. By extension, it is evidenced that the X-axis, which is the horizontal asymptote of the first pair of conjugate hyperbola's, represents the base-level of erosion with respect to the slopes which extend to the bottom of the inselberg. Indeed, in the latter case part of the toe-slope is covered partly with alluvial deposits. These statements lead to a first conclusion in that at least two erosion cycles controlled denudation of the Vélatouri. This may also explain the asymmetrical shaping of the peaks.

Furthermore extrapolation of computed curves (4) and (5) until interception with the first pair of conjugate hyperbola's (1) and (3), provides two points which occur at the same altitude of the highest (southern) Vélatouri peak. It is hardly to believe that this is due to a pure coincidence with respect to the heterogeneity of the bedrock. Therefore, we rather think that the line through the three points, represents the former plateau level from which erosion started. Actually there exist also among the erosion-surfaces, one that is established at about 140-145 m and which is capped by a paleosol.

Furthermore, pure mathematical analysis leads also to the clarification of some aspects of the involved denudation processes. From equation (2) it follows that for decreasing values of X, there is gradual increment in Y. This means then that at a given moment the denudation is inversely proportionate to distance from the base level of erosion and that this is controlled by a hyperbolic function.

From the foregoing, it appears that the apparant geometry of the double-conical Vélatouri inselberg is subject to mathematical (analytic-geometrical) reconstruction. Conversely pure mathematical extrapolations, reveal the position of no longer existing landscape elements (such as base levels of erosion, erosion surfaces) which tally entirely with the geomorphic features of the environment. In the meanwhile, it could also be stated, that only those geomorphologic forms which are controlled by contemporaneous and identical processes are expressed by the same or related mathematical expressions. These statements corroborate to the viewpoint that natural forms which may be expressed by the same type of mathematical formulation at a given moment point to a common evolution of geodynamic equilibrium.

Taking the first derivative of expression (2), we obtain :

$$\frac{dy}{dx} = -\frac{k}{x^2}, \qquad (6)$$

where $\frac{dx}{dy}$ is the instantaneous rate of change in Y per unit change in X, at point X. Note that geometrically:

$$\frac{\Delta y}{\Delta x} = \tan \theta, \qquad (7)$$

where the first member is the average rate of change in Y per unit change in X, in interval x. As $\Delta x \rightarrow 0$ then we have that

$$\frac{dy}{dx} = \tan \theta = \text{slope of the tangent line at point X.} \qquad (8)$$

Next we may write:

$$\frac{dy}{dx} = -\frac{k}{x^2} = \tan \theta \simeq \theta \qquad (9)$$

for slopes inferior to 20°. In geo-dynamical terms, this means that with increasing values of X, the change in Y is negative and this at an increasing rate. In other words, the nearer the base level of erosion, the slower the erosional activity. Whereas equation (2) is a timeless expression of the profile shape, time is included in (9) and yet we have a mathematical measure for the effect of denudation through time. With respect to what preceeds: this also indicates that erosion is less active in the lower profile section than in the higher ones. Both laws, (2) and (9), explain on an empirical mathematical basis why at the greatest distance from and at the highest level above the base level of erosion, the greatest activity occurs.

These laws have been earlier qualitatively formulated. The fact that they also can be derived from pure empirical mathematical deductions adds to the validity of the above statements.

The next question to be solved is whether or not this is applicable to fossil forms. Again, we refer to the example of the Vélatouri for the slopes of the hanging valley correspond to a base level of erosion which is evidently older than the one controlling the slopes extending down to the foot of the inselberg. It implies that fossil equilibrium profiles, change little if any once the corresponding base level of erosion has dissappeared. They remain as well preserved relicts in the present landscape. This is also true to some extent for the curvilinear profile of the southern wall since it is already (partly) covered, as we stated above, by alluvial deposits. It is clear now that both slope profiles adapted their form to successive base levels of erosion and are persisting independently under present conditions. Consequently, we may extend the above statement as follows: *geo-dynamic profiles, fossil or actual, may be expressed in a mathematical form corresponding to a fossil or present base level of erosion.*

2.2. Age of geodynamic profiles.

Unless one deals with a zone of semi-arid climatic conditions, the chances to find uniform curvilinear erosion profiles, are poor. Especially, in so-called periglacial regions, slopes are most often covered with screes or weathered bedrock, even when they never have been mantled with alluvial or eolian deposits. Actually, the rate of evacuation of screes or weathered bedrock is slow and consequently the lack of fresh, active valley walls is a limiting factor for the development of new slopes. The probability of finding young, uniform curvilinear geo-dynamic profiles is low. Most of the slopes in periglacial areas are old and highly disharmonic. Eversince their establishment, they were subdued to various changes of climatic conditions and with this respect they are to be referred to as multicyclic.

The Focant plaine in the Belgian Ardennes yields good insight in slope development under periglacial climatic conditions. This depression is a somewhat aberrant feature in the morphological shaping of the Ardennes. It is situated, geologically speaking, on the contact between famennian (devonian) schists in the north and sandstones in the south. This is conjectured in the landscape by the smooth curvilinear rising northern slope opposite the rather abrupt southern slope (fig. 2). Also, the nature of the surficial deposits vary on both valley slopes : whereas the southern is merely bare bedrock (2) with only a thin, weathered clay mantle (1) along the toe-slope, the variety and succession is much greater on the northern one. Bare bedrock (2) is found only in the very upper part of the slope profile. Below follows a slope surface composed of a thin solifluction loess sheet overlying the weathered clay which rests on the schist substratum (4) within less than 0.80 m under the surface. On other parts along the northern slope, one may find between the two mentioned surface units, a zone with the weathered clay immediately cropping out, and then showing similar succession of surficial deposits as on the southern slope. From this, it may be delineated that the clay dips slope downwards under the loess cover. In the middle of surface (4) a knickpoint breaks the steep

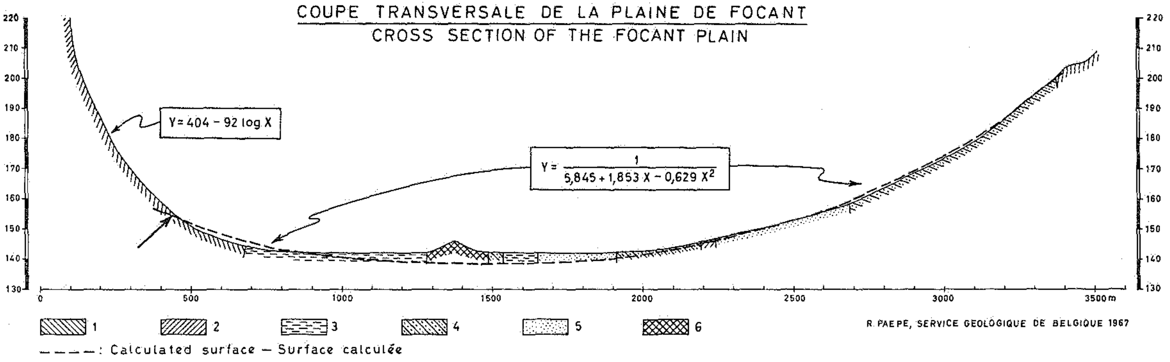
COUPE TRANSVERSALE DE LA PLAINE DE FOCANT
CROSS SECTION OF THE FOCANT PLAIN
Y = 404 − 92 log X
Y = 1 / (5,845 + 1,853 X − 0,629 X²)
220
210
200
190
180
170
160
150
140
130
0
500
1000
1500
2000
2500
3000
3500 m
1
2
3
4
5
6
– – – – : Calculated surface — Surface calculée
R. PAEPE, SERVICE GEOLOGIQUE DE BELGIQUE 1967

FIG. 2.

slope inclination; this point is to be considered as separating the sector of removal from the sector of transportation (R. SOUCHEZ, 1966) especially since below this point the surficial deposits show an increase in thickness. Another knickpoint appears at the lower limit of the surface (4). This one is then to be considered as a new point of separation between the removal-transportation sectors as a result of the downwards migration of the first knick-point due to levelling off (R. SOUCHEZ, 1966). From thereon, another surface (5) with still weaker slope descends to the flat plain; it is composed of a fairly thick (0.50 m) loessic cover, over-lying the weathered clay which in turns lies on a complex of fine stratified schist pellets (« Grèzes litées ») and zones of congeli-fracted schists extending downwards up to 2 m and more. Finally, in the valley, quaternary deposits attain usually a much greater thickness with the exception of some outcropping substratum domes (6) which may also appear as low, narrow ridges in the landscape. The valley deposits consist mainly of alluvial clays (3) up to 3 m followed by the mass of schists pellets and congelifraction schists up to about 5 m. When the alluvial clay is lacking, the sequence is built up by eolian loess, 4 to 5 m in thickness, under-lain by the weathered clay which separates the loess from the lowerlying schist complex. Finally, at the bottom, another loess may be found resting on congelifracted schists, sometimes also impregnated by a reddish colouring. The latter then shows great affinity to a reddish paleosoil, found in a brick yard exposure at Wanlin, underneath the same sequence of transported schists and loesses. The stratigraphical position points to an eemian age for the reddish soil.

Now, how does slope evolution fits into the above outlined strati-graphical sequence? By computing the profile form with data obtained mainly from the transportation-ablation sectors (between 2 900 and 3 400 m), we are sure to work within the reach of a geodynamic profile since the soil mantle in these sectors is thin (R. SOUCHEZ, 1966).

This leads to:

$$Y' = 5{,}845 + 1{,}853\,X - 0{,}629\,X^2, \qquad (10)$$

which is a second degree hyperbola. Projection of calculated points at distances inferior to 2 900 m up to about 400 m on the X-axis, results in a well fitting calculated profile with the natural one. However, in the section between 400 m to 0 m of the X-axis, from the point where the steep southern wall on bare bedrock starts to develop, fitting stops and another mathematical expression must be provided :

$$Y = 404 - 92 \log X, \tag{11}$$

which is a logarithmic function.

We now know that this break in the profile trend corresponds in the field with a lithological break : schist — sandstones. Thus, we may assume that, were it not of differential erosion, the computed form (10) would fit over the whole length of the cross-section. At least, it does on the schist substratum while it is also bending, conformly to the natural slope on the part of the southern slope consisting of schists. From the previous chapter we now derive that the entire slope cross-section on schists represents a geo-dynamic equilibrium. Next now comes the fact that where the computed profile passes underneath the alluvial (aggradation) cover at about 1 500 m, calculated profile depth and top of bedrock tally perfectly with each other.

Actually, in a boring, bedrock at this place was reached at 3.60 m below surface or at 138.40 m absolute depth whereas calculated bedrock surface was found to lay at 138.70 m. The error is negligeable. For that matter, it is quite conceivable to conclude as to the connection of the exuded profile slope with the burried one in one and the same geo-dynamic equilibrium profile. As we have seen, the cover deposits are related to the Weichsel Glaciation. On the other hand, the presence of a reddish soil overlying bedrock implies that no major-erosion later to its formation took place. As of then, it follows that both valley erosion and slope development are at the utmost, young Eemian in age but very probably still older. Indeed, it can hardly be imagined that, owing to the fact that V-shaped valley incision generally occurs under forested landscape conditions, a broad flat valley would have been built during the Eemian. In fact the valley slopes remember the ones to be known as resulting from cryoplanation processes, so that pre-eemian, cold-periglacial conditions are presumed as for their origin. With regard to the thin sediment mantle they mainly date back from the wettest part (solifluction) of the Last Glaciation with interference of extreme cold-dry climatic conditions (congelifracts).

2.3. **Structural conformity of geodynamic profiles.**

While studying the geology of Four Mile Creek in northeastern Iowa, several lobes of dissected erosion surfaces scattered on both sides of the Creek, were mapped. Very seldom, the surfaces of these lobes were flat while also basic topographic data were lacking. In such situation, one finds it difficult to disentangle which of the lobes belong together from the point of view of geodynamic equilibrium. It is obvious that in checking mathematical relationships between lobes of the two sides, a clarification may be

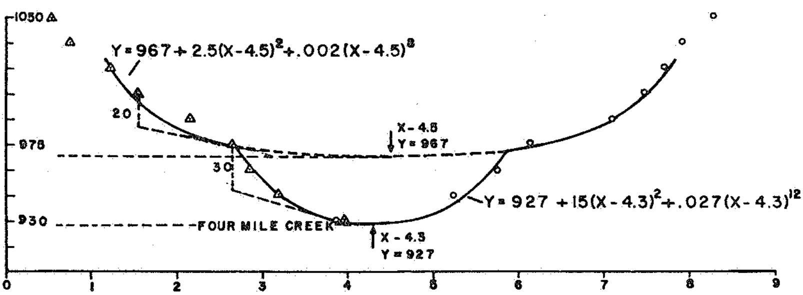

FIG. 3.

obtained. In order to do this, the best arrangement of topographic data was chosen to compute the valley slope profile whatever the side of the creek was dealt with. It was found that the calculated form of the upper right profile section (fig. 3):

$$Y = 967 + 2{,}5(X - 4{,}5)^2 + 0{,}02(X - 4{,}5)^8, \qquad (12)$$

could be fairly well projected through the less orderly arranged topographic data on the left side.

In the same way, computed shape of the lower left part (present Four Mile Creek valley):

$$Y = 927 + 15(X - 4{,}3)^2 + 0{,}027(X - 4{,}3)^{12}, \qquad (13)$$

went through the poorly arranged data of the other side.

Instead of using the curve-fitting method based on the least squares method, (as has been done herefore) an algebraic form description of curves (Ch. E. JENSEN) is used here. The choice for best fitting depends on the choice of the starting point along the X-axis. As of then, the constants are no more subject to statistical evaluation : the value of the first constant e.g. 967 ft is then an absolute number corresponding to the field data; it is the absolute height of the bottom of the geo-dynamic profile at the time it was not dissected by the Four Mile Creek. For the same reason, the first constant of (13), 927 ft is the calculated height of Four Mile Creek bed, which stands in fairly good agreement with the topographic datum at that point. From the graph, we may readily see that not only the present valley bottom lies about 40 ft below the previous local base level of erosion, but that it is also slightly shifted to the left.

The following main conclusions can be drawn from the above. First, it can be stated that the remnants of the various erosion surfaces and connected slope profiles occurring at more or less the same altitude, do belong to the same cycle of geodynamic evolution. Next, they are not affected by tectonic movements.

3. INTERNAL FORMS.

By internal forms is meant the geomorphological features which occur within the body of the deposits. During field investigation, it was many often emphasized, how much buried relief may be different from the present one i.e. buried flat erosion surfaces against the present undulating relief. Besides, with some luck, it was quite possible also to observe buried valleyforms over their entire length (R. Paepe, 1963, 1964, 1967). It was found that throughout the Late Pleistocene sequence, a differentiation of the channelform could be followed : whereas valleys within deposits dating from the Eemian, are delineated by deeply eroded rather V-shaped cross-sections, a steady flattening of the valley form throughout the Last Glacial (Weichselian) deposits was evident. It lies beyond doubt that this flattening is not the result of a fortuitous orientation of the valley cross-section closely to the longitudinal axis of the former river course. For this would imply then that for one or another reason all the brick yard walls or natural exposures would cut the geological sequence under a similar angle. This is hardly to believe. Moreover, it could be stated that several superposed channelforms, definitely of different geological ages, but all with the longitudinal axis in almost the same sub-parallel direction (Rumbeke, see R. Paepe, 1964), showed the above mentioned trend of valley flattening. For these reasons, it is our conviction that the observed channelforms are relating to a really existing morphology.

Now, the channelform is quantitatively given by a relationship known as the « width-to-depth ratio » (L. B. LEOPOLD and Th. MADDOCK, Jr., 1953). Actually, these parameters stand in direct relationship to some other hydraulic characteristics of stream channels as discharge, velocity, suspended and bed load, and scour. It is clear that those characteristics reflect prevailing climatic conditions. For the geologic constitution of an area which may evidently also influence the mentioned characteristics are to be accounted for as constant when dealing with drainage basins of limited extent. The latter statement surely holds for the stream channels occurring one above the other in an exposure. Another point is that all the observed stream channel cross-sections are occupying a midstream position so that with respect to this, variations of the hydraulic characteristics in a down- or upstream direction must not be taken into consideration.

It now becomes clear that in calculating the width-to-depth ratio of fossil valleys, we dispose of a geomorphological parameter relating to paleoclimatic evolution. The following data are obtained for the various lithostratigraphic units of the Late Pleistocene (R. PAEPE and R. VANHOORNE, 1967) :

Eem :

peat and gravels : Antwerp, 9; Poperinge, 10; Rumbeke, 9; Waarmaarde, 9.

Weichsel :

loams and coarse sands : Antwerp, 15; Poperinge, 17; Racour, 16; Waarmaarde, 16; Warneton, 16.

peaty loam formations : Poperinge, 35; Rumbeke, 39; Zonnebeke, 34.

cross bedded sands :

base : Poperinge, 77.
top : Poperinge, 125.

One may object against the poor number of records available. However, one must keep in mind that these are the result of more than ten years field work. On the other hand, the constancy of the values at each litho-stratigraphic level pleads in favour of their validity.

Let us now see how these data plot against a time scale. From the litho-stratigraphic evidence and the absolute dating of the

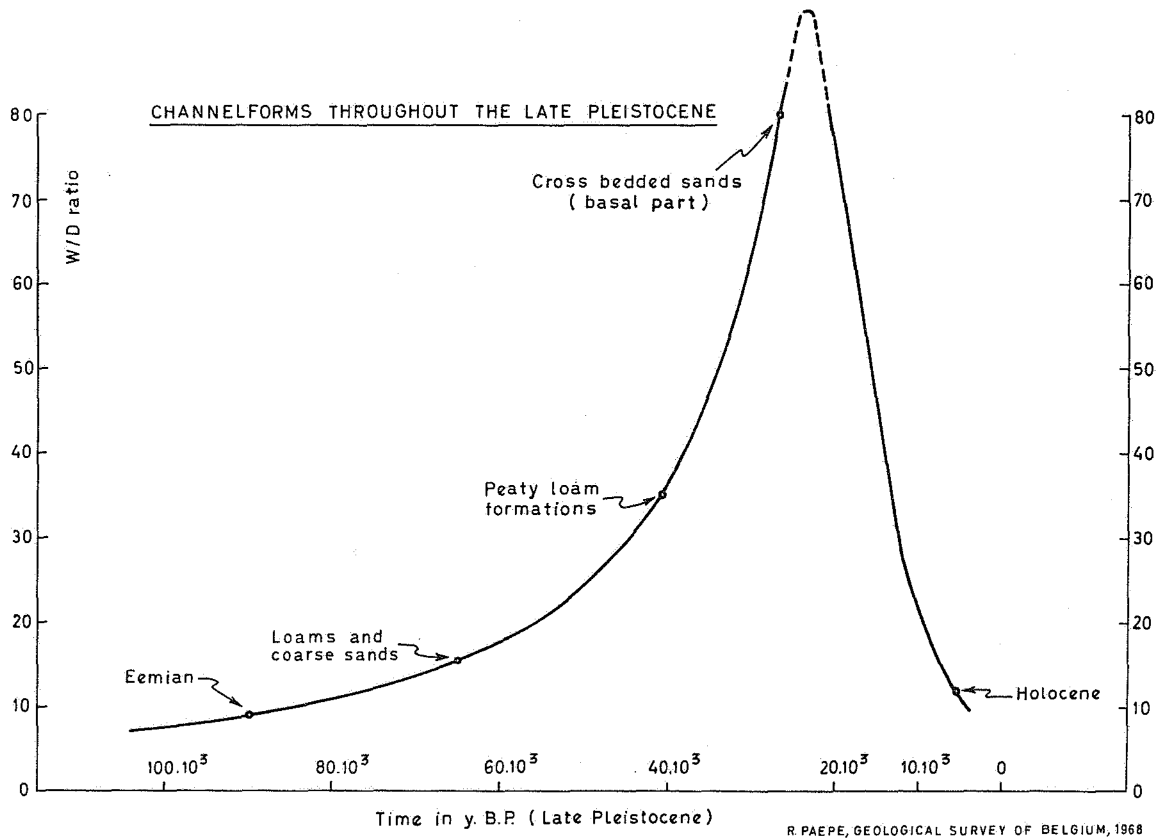
CHANNELFORMS THROUGHOUT THE LATE PLEISTOCENE
W/D ratio
80
70
60
50
40
30
20
10
0
Cross bedded sands
(basal part)
Peaty loam
formations
Loams and
coarse sands
Eemian
Holocene
100.10³
80.10³
60.10³
40.10³
20.10³
10.10³
0
Time in y. B.P. (Late Pleistocene)
R. PAEPE, GEOLOGICAL SURVEY OF BELGIUM, 1968

FIG. 4.

vegetation horizons (R. PAEPE and R. VANHOORNE, 1967), it can be assumed that successive periods of maximum erosion occurred at about :

90.000 y.B.P. for the *peat and gravels.*
65.000 y.B.P. for the *loams and coarse sands.*
41.000 y.B.P. for the *peaty loam formations.*
27.000 y.B.P. for the *cross bedded sands* (base).
22.000 y.B.P. for the *cross bedded sands* (top).

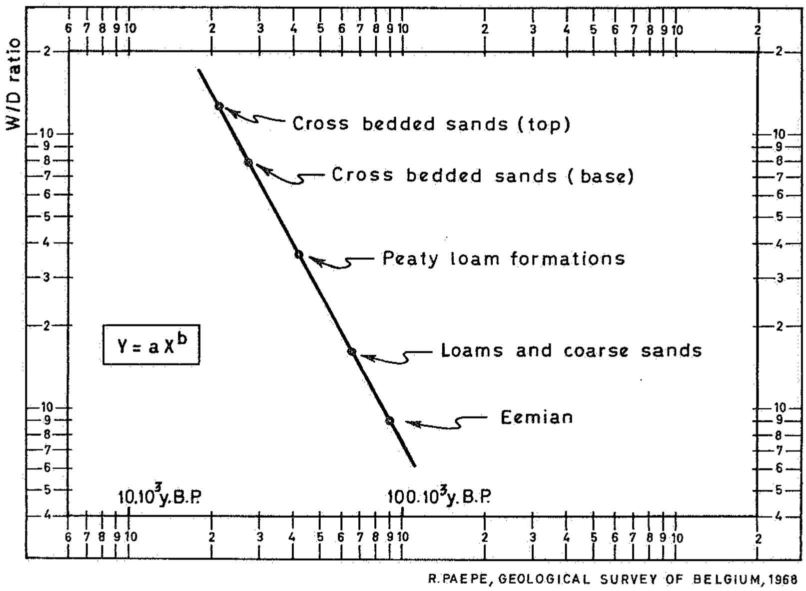

FIG. 5.

On regular paper, where the time-scale is set out along the X-axis and the average values of the width-to-depth ratio along the Y-axis, a curvilinear trend is obtained (fig. 4). On logarithmic paper (fig. 5), this succession shows a straight line pointing to a hyperbolic function which read as :

$$Y = a X^b.$$

This means that the effect of the changing climatic conditions on the geometry of the stream channels throughout the Late Pleistocene is a continuous evolution. In fact, the overwhole trend of the lowering temperature finds its expression through this statement. For at the time of the Eem, warm and humid climate permitted a dense vegetation cover resulting in concentrated water evacuation leading to V-shaped valley forms and consequently low width-to-depth ratio values. With the lowering of both temperature and precipitation at the beginning of the Weichsel, the vegetation cover diminishes and solifluction appears as a new form of rock waste removal under periglacial conditions. In the field at this stratigraphic level which corresponds with the *sands and gravels* and/or *loams and coarse sands*, solifluction sheets are seen to descend laterally into the stream channels (Antwerp, Racour, Warneton). As a result, there is a considerable increment of bed load which is known to influence positively valley widening (L. B. LEOPOLD and TH. MADDOCK, 1953). During the period of the *peaty loam formations*, decrease in water supply due to a further drop in precipitation and temperature, changes again the regimen of the stream. Consequently bed load increases once more but probably as a result of an important decrease in discharge and velocity, rather than of lateral supply by solifluction. Also the width-to-depth ratio increases not only thanks to the widening effect of this increase in bed load, but subsequently to a decrease in depth as well as in discharge. Actually, the thin ripple-mark structured morphology of the fine textured *peaty loam formations* points to an extreme shallow water flow (R. PAEPE, 1965, 1967). During the *cross bedded sands* deposition, we know that temperature continued to drop and that also humidity became considerably lower. Actually, the first important eolian loess deposits happen to occur at this level, laterally going over into the *cross bedded sands* (R. PAEPE, 1965, 1967) deposits. The latter formations are considered as melt water deposits which activity and presence are of limited extent. At that moment, high amounts of weathering waste due to prevailing frost activity, are to be evacuated only by these periodic melt water flows. As a result of such a heavy bed load, which may contain large blocks of flint, tertiary clay lumps etc., valley widening was a most important process. Scour seems to have been little as one may find usually the underlying top of the *peaty loam formations* undisturbed underneath these *cross bedded sands* (R. PAEPE, 1965, 1967).

With respect to the base and the top of the *cross bedded sands* there even exist an increase in the valley widening. In the same time interval, climate evolves towards even more severe conditions. This adds to the assumption that increasing valley flattening or rising values of the width-to-depth ratio is a result of more cold and dry climatic conditions.

In conclusion, whatever the fluctuations of the climatic conditions during the Late Pleistocene resulting in the establishment of interstadials and extreme severe cold periods, there exists a more general evolution characterized by a steady growing deterioration of the climate as a whole. As this statement follows from a feature so inherent to the sedimentation pattern as is the channel form, the Late Pleistocene may be looked at as a continuous evolution with respect to climatic conditions. As of then the sharp distinction which litho-stratigraphically is found between the Pleniglacial A, cold-humid, and the Pleniglacial B, cold-dry (R. PAEPE and R. VANHOORNE, 1967) is much more due to an adjustment of the sedimentation processes to climate than of a sudden break in the climate itself.

4. CONCLUSION.

Through mathematical analysis of landscape forms, it becomes possible to integrate geomorphology to stratigraphy. In computing external, at first view isolated landscape forms, their assimilation to a nearby stratigraphic sequence of a deposit succession is realized; on the other hand, it quite possible to link isolated features of a same geomorphological phenomenon as well. Computation of internal forms, as part of a stratigraphic sequence may lead to important statements with respect to sedimentological-climatological evolution.

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