PROCESS MODEL FOR SEDIMENTATION UNDER TURBULENT FLOW CONDITIONS

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ABSTRACT

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This study involves construction of a theoretical model of flow/sediment interaction. Described herein is a mathematical flow/sediment response model in which deposition occurs analytically rather than experimentally. The model consists of a series of discrete segments which can be combined into three functional sections: (1) definition of turbulent state, (2) turbulent state/sediment interaction, and (3) sedimentary deposit. Based on current understanding of the nature of flow and flow/sediment interactions, this model bears the promise of complementing future theoretical and empirical studies.

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INTRODUCTION

It is axiomatic in sedimentology that a deposited sediment bears the diagnostic imprint of the depositional environment. A great number of investigations have studied aspects of sediment such as size-frequency distributions and minor structure assemblages in order to deduce depositional environments. Laboratory studies and observations of recent sediment indicate that the flow regime is the primary variable from environment to environment. The complex motions within the flowing medium produce a sedimentary deposit which in some way mirrors this complexity. However, it is not known to what degree of fidelity the deposit mirrors the details of flow.

The fundamental limit for extracting information from sediment is a direct function of the faithfulness of the sediment in reflecting hydrodynamic processes. If the transfer of information was perfect, the fundamental hydrodynamic equations could be completely deduced from the sediment characteristics. If no information concerning the fundamental hydrodynamics equations can be deduced from a sediment, then study of sediment would be without value for interpreting depositional environments.

The real situation falls somewhere between these two extremes. The extent to which the quality of the record falls short of a perfect imprint of the hydrodynamic regime marks the absolute limit of interpretability of the sedimentary record. This limit plays the same role in sedimentology as does the uncertainity principle in physics.

The attempt to understand the relationship between flow regime and sediment has always been one of the primary goals of sedimentological research. Much insight into this problem has been provided by experimental work involving flumes and wave-tanks, as well as direct observations of natural systems. These studies as well as most others concerned with evaluation of the flow regime are hampered by the difficulty of monitoring flow regime without interfering with it. Therefore, the information such studies have provided has been mainly qualitative. In general, they have defined the evolution of a sedimentary accumulation through time but provide little information concerning the detailed events that constitute this evolution.

Another approach toward the same end involves construction of theoretical models of flow-sediment interaction. Described herein is a mathematical flow/sediment response model in which deposition occurs analytically rather than experimentally. This model, based on current understanding of the nature of flow and flow/sediment interactions, bears the promise of complementing empirical studies.

Explicit and implicit reasons exist for building such a process model. Explicitly, similarity between the product of the model and natural accumulations can provide information on the workings of an otherwise unobservable natural system. Implicitly, the lack of correspondence between output of the model and natural deposits represents a measure of the reasonableness of the commonly accepted framework of theoretical hydrodynamics upon which the model is based.

In order to be reasonable, such a model must be intimately tied to turbulent flow. An object of this study, then, is the development of a mathematical model of small scale sediment variation based on the concepts of fluctuating hydrodynamic energy regimes, and further, to demonstrate that the output compares favorably with empirical observations. With this established, a certain degree of validity can be assumed for the model. At this point, deeper analysis of the output concerning the quality and completeness of the sedimentary record is justified. Such analysis can provide a testable framework for future empirical and theoretical studies.

TURBULENT FLOW

In most depositional environments, the flow regime is turbulent. A wide variety of mathematical models of turbulent flow have been developed in hydrodynamics and aerodynamics. The vast majority of these, if not all, were developed for non-sedimentological objectives. A major problem in construction of the response model involves modification of these models to make them sedimentologically relevant without destruction of their theoretical validity.

Turbulence is that component of flow which is characterized by random rotational motions. As a result, the velocity at a fixed point in space is not constant but fluctuates irregularly about a mean value. The fluid elements which perform these fluctuations are macroscopic "lumps of fluid" (Hinze, 1959). In a fully developed turbulent flow, there is a hierarchy of such fluid packets varying in size from meters to microns. Each of these packets has its own intrinsic motion which is superimposed on the mean flow. These packets retain their identity and move as a unit through the mean flow for a short period of time before being assimilated by the fluid around them. The length of time the fluid packet retains its identity is a function of the packet size and the difference between the packet velocity and the mean fluid velocity.

As a consequence, both Eulerian and LaGrangian models have been developed for turbulent flow. In the Eulerian models, the reference frame is fixed and the primary variables of interest are the magnitude and direction of the velocity vector at a point In the LaGrangian models, a given fluid particle is followed through the flow field, thereby introducing the additional variable of location.

In either case, the precise functional relationship between these variables as well as the magnitude of various constant terms is not known with great precision owing to the difficulty of obtaining empirical data from the flow without perturbing it. As will be seen below, turbulent models are a blend of varying proportions of theoretical and empirical components. The model discussed below is developed for an Eulerian framework which concerns velocity fluctuations at a point.

GENERAL CHARACTERISTICS OF THE MODEL

The hydrodynamic response model described below simulates the behavior of turbulent fluid at one point in space through time. The fundamental aspect of the flow is the velocity-shear stress and the duration of any pulse.

The competency of the flow is directly related to the magnitude of the shear stress. The extent of erosion or deposition is a function of shear stress magnitude and duration of the pulse.

The model consists of a series of discrete segments. These individual components can be combined into three functional sections; (1) definition of turbulent state, (2) turbulent state/sediment interaction, and (3) sedimentary deposit. The turbulent state section contains those components judged necessary for modeling of the turbulent flow, and defines the duration and intensity of the horizontal component of fluid velocity. Discussion of the components of this section will include the criteria used for selection or rejection of alternative approaches to turbulent modeling.

The turbulent state/sediment interaction section relates the turbulent state to various size-frequency distributions and monitors the nature and extent of erosion or deposition from velocity fluctuation to fluctuation.

The sedimentary deposit section represents the accumulated result of a large number of turbulent flow/sediment interactions. The time sequence of erosional and depositional events as well as grain size analysis of this record will provide an estimate of the relative proportion of total turbulent events recorded in the sedimentary record, and the extent to which the record can be used to deduce the nature of the turbulent flow regime.

A generalized flow diagram of the model is depicted in Figure 1.

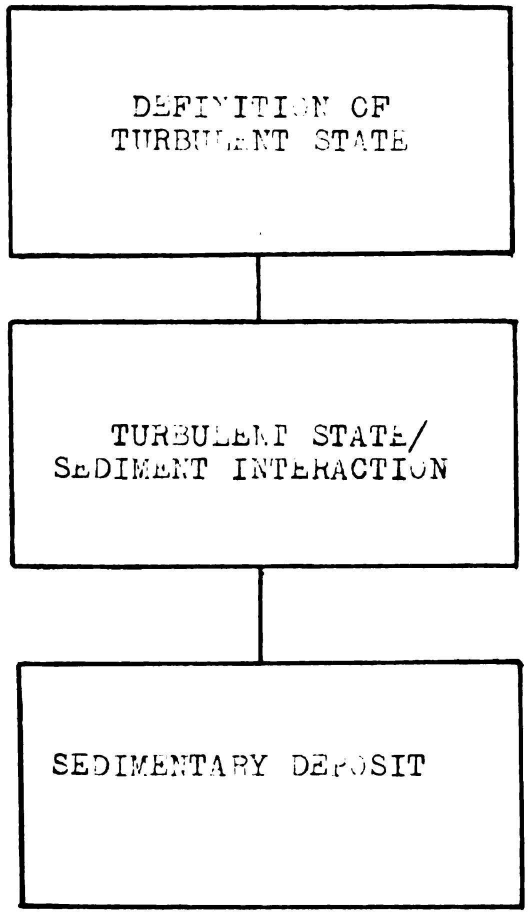

Figure 1. Hydrodynamic response model - generalized flow diagram

TURBULENT STATE

A wide variety of models of turbulence have been formulated. These models in general contain two common characteristics:

a. A treatment of turbulence ranging from near molecular motions to the macroscopic.

b. Incorporation of averages over time of such variables as flow direction and velocity, reflecting the inherent stochastic nature of turbulence.

With respect to the problem in hand, the presence of these factors requires modification of models to estimate instantaneous values for the flow velocity. In addition, those parts of general turbulence models involving aspects of turbulence whose dimensions are too great or too small to influence sedimentation will not be considered.

Determination of Flow Velocity

Flow velocity is a fundamental parameter for definition of the turbulent state. In accordance with general theory, this velocity can be considered to be composed of a mean velocity and a turbulent component. Setting up orthogonal coordinates, x, y, and z at a fixed point in space, where x is parallel to the direction of the mean current flow, we can write:

$$\begin{aligned} u &= \bar{u} + u' \\ v &= \bar{v} + v' \\ w &= \bar{w} + w' \end{aligned} \qquad (1)$$

where u, v, and w are the velocities measured parallel to the x, y, and z directions (Schlichting, 1960). The bar denotes the time averaged component of velocity. For our purposes we will assume that $\bar{v}$ and $\bar{w}$ can be set equal to zero. The mean velocity in the x direction $\bar{u}$ can be set at any value of interest to the investigator. The prime denotes the turbulent component of velocity. These components are random variables which may have either positive or negative sign.

Experimentally it has been shown that for the time averaged conditions described above, the horizontal component of turbulence transverse to the mean flow direction is negligible (Hinze, 1959). Therefore equations 1, for our purpose reduce to:

$$\begin{aligned} u &= \bar{u} + u' \\ v &= v' . \end{aligned} \qquad (2)$$

Additional experimental evidence has shown that u' is in general greater than v', and the magnitudes of the two components are roughly correlated (Schlichting, 1960). As a result, deductions concerning the nature of the horizontal component will provide substantial information concerning the vertical component. Therefore, as an additional simplification the model discussed herein uses as the primary variable of interest, the magnitude of the horizontal component of the velocity vector.

For any event based model, the determination of the probability distribution governing the turbulent components of velocity is of critical importance. Empirical investigations of aerodynamics and hydrodynamics indicate the fluctuations are normally distributed about the mean velocity (Hinze, 1959; Schlichting, 1960). Numerous studies have indicated that the standard deviation of the horizontal component of velocity is proportional to the mean (Kalinske, 1947; Clark, 1968; Kozyrenko, 1966).

The degree of turbulence as expressed by this variation about the mean would be expected to vary depending on the nature of the sediment-fluid interface , and the height above the interface at which the measurement was taken. Recent studies have indicated that, in general, the magnitude of the fluctuations decrease with increasing distance from the interface (Ishihara, Yokosi, Ueno, 1969).

Since this study is concerned with the movement of sediment as bedload, the nature of the fluctuations near the fluid-sediment boundary are critically important. In this region, Kalinske has experimentally determined that the velocity fluctuation is such that σ is about 0.25 where σ is the standard deviation of u about $\bar{u}$. This value of the constant of proportionality may, in fact, deviate somewhat from the true value, but the existence of proportionality of mean and standard deviation in such physical phenomena is generally observed. Empirical data from both laboratory studies and natural river studies indicate that near the boundary, the standard deviation of the velocity is in general, a small percentage of the mean (Clark, 1968; Ishihara, Yokosi, Ueno, 1969).

An instantaneous horizontal velocity is thus determined in the model by sampling from a normal distribution whose mean equals the mean velocity and whose standard deviation is set to an empirical constant.

Duration

Each velocity generated by a random model such as that described above persists for a certain duration. In actuality, the velocity varies continuously from moment to moment and duration of velocity in that sense would cover infinitely small periods of time.

The complex set of velocities and frequencies that constitute turbulent flow can be regarded as a superposition of eddies of various sizes and spatial orientations (Hinze, 1959). The behavior of velocity at any point in space over any interval of time is simply the sum of the individual eddy velocities.

The largest eddies have dimensions on the order of the scale of the mean flow. The smallest fluid packets in natural streams have dimensions in the millimeter range (Yokosi, 1967). It is obvious that many of these eddies and their associated velocities cannot appreciably affect sedimentation in the sand size range.

Although the velocity fluctuations with time at a point are in reality represented by a complex continuously varying wave form, in terms of sediment interaction the replacement of this wave form with one much simpler has little appreciable effect on the sedimentary deposit. This simpler time-average spectrum is associated with eddies of a particular size range. The relevant eddies must penetrate to the fluid-sediment interface and possess the combination of duration and intensity necessary to affect sedimentation in the sand size ranges.

As will be shown below, the magnitude of the turbulent shear stress is related to the correlation between the horizontal and vertical components of turbulence. Numerous experiments have shown that the smallest eddies may have considerable energy, but because they do not contribute to the correlation between u and v, they are of small significance in generating the turbulent stresses (Ishihara, Yokosi, Ueno, 1969). In addition, the larger size eddies would be expected to more likely penetrate to the sediment interface, and persist longer at a point once they impinged upon the surface.

From this analysis, it is concluded that larger size eddies with dimensions on the order of meters possess those attributes which make them sedimentologically the most important. Experimental studies have indicated that the dimensions of these eddies are in general controlled by the geometry of the channel (Yokosi, 1967).

In the model under discussion, eddy size is treated as an independent variable fixed by the investigator at a single value for a depositional cycle. The range of allowable eddy sizes is determined from empirical studies of a variety of environments. Once the eddy size is fixed, the length of time the eddy will spend over a point is simply the quotient of eddy size divided by its velocity. Later modification of this model will probably include eddy size as a random variable rather than as a constant.

The model velocity spectrum over time (Figure 2) resembles a "square wave function" with instantaneous increases and decreases of velocity. The physical picture which emerges from the velocity spectrum is one in which relatively large vortex structures, elongate in the streamwise direction, move past the sediment interface with individual convection velocities. These structures have widely varying size and strength, and the velocity fluctuations observed by Kalinske and others can be attributed to their passage by a point.

It is to be expected there would be a large

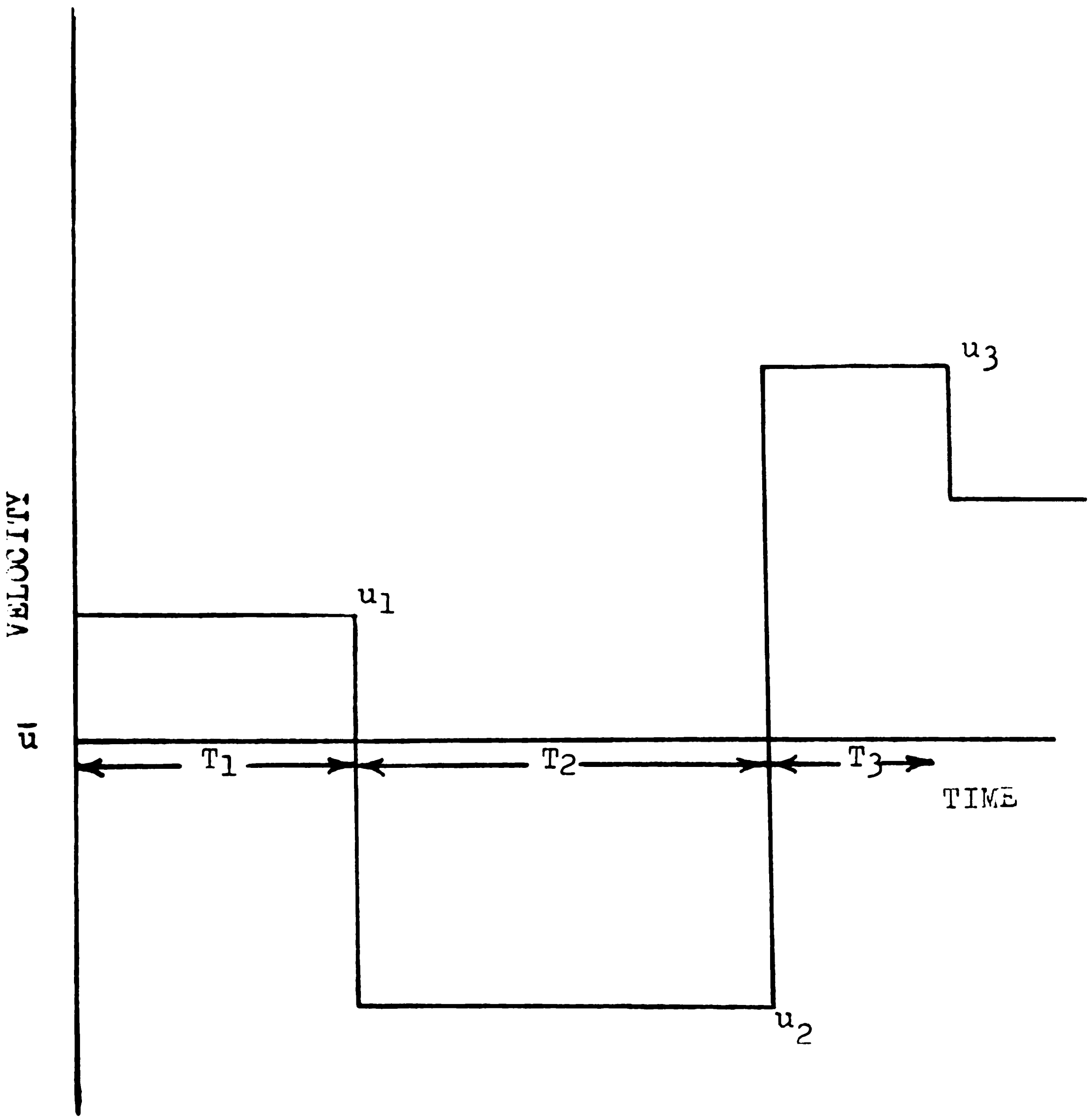

Figure 2. Model Velocity Spectrum

instantaneous value of turbulent shear stress which coincides with the passing of each structure (Clark and Markland, 1971). It is these stresses which are vitally important in the entrainment of sediment (Sutherland, 1966), and the following section describes a technique for indirectly estimating the fluctuation of turbulent shear stresses associated with the discontinuous shifts of velocity represented by Figure 2.

Shear Stress

A function of velocity, shear stress is the appropriate variable for dealing with flow/sediment interaction. Whereas velocity is concerned with length per unit time, shear stress is concerned with force per unit area.

The relationship between velocity and average shear stress can be expressed as:

$$\overline{\tau_{xy}} = -\rho\overline{u'v'} + \mu \frac{d\bar{u}}{dy} \qquad (3)$$

where
- x is the longitudinal direction parallel to the flow
- y is the vertical axis
- $\frac{du}{dy}$ is the velocity gradient in the y direction
- ρ is the density of water
- μ is the dynamic viscosity of the fluid
- u' is the velocity fluctuation in the x direction
- v' is the velocity fluctuation in the y direction
- $\overline{\tau_{xy}}$ is the force per unit area in the x direction on a surface normal to the y axis (Schlichting, 1960).

This relationship expresses the physical fact that momentum is being transferred by the movement of the fluid packets, in addition to the normal transfer of momentum through molecular motions. The first term of the equation represents the component due to turbulence and the second term is the laminar contribution. Under the velocities considered in this paper, the laminar term is negligible compared to the magnitude of the turbulent term so that we may consider the equation to be approximately

$$\bar{\tau}_{xy} = -\rho\overline{u'v'} \quad (4)$$

Because this function represents time averages, instantaneous shear stress or shear stress over short periods of time cannot be deduced from it directly.

According to Hinze (1959), shear stress can be envisioned to arise from the transfer of momentum across interfaces. Fluid packets carried from one region to another --- the velocity generally being different in the two regions --- will either gain or lose a small amount of momentum. If the packets gain momentum by being carried into a region of higher velocity, they will exert a corresponding retarding force upon the flow. If the packets lose momentum conversely, a corresponding force is exerted upon the flow. The difference between the mean velocity of the fluid packet and the velocity of the region into which it moves is recorded as the momentary velocity fluctuation.

Because of the mixing process in turbulent motion, momentum is constantly being transferred back and forth across any imaginary plane. As a result, the instantaneous shear stress fluctuates about the mean value expressed by equation 4. These fluctuations are closely tied to the velocity fluctuations.

Ideally, shear stress estimates should arise from calculations involving incremental velocity differences over time as the turbulent velocity changes. However, as discussed above, the velocity duration model used herein resembles a square wave function involving constant velocity for a certain duration, succeeded instantaneously by another velocity level fixed for a given duration. In the absence of a continuous velocity function, shear stress for the model was estimated in an indirect fashion.

The average shear stress was calculated for a particular mean velocity and standard deviation, using equation 4. The equation contains u' and v', horizontal and vertical fluctuations of velocity respectively. It has been shown that v' is correlated

with u' with a constant of proportionality equal to k. Therefore $|\overline{v'}| = k \cdot |\overline{u'}|$ and equation 4 may be rewritten as $\overline{\tau}_{xy} = \rho \cdot |\overline{u'}| \cdot K \cdot |\overline{u'}|$ (5)
Experimental evidence indicates k is roughly equal to the value 0.2 (Schlichting, 1960).

Experimentally it has been shown that shear stress varies as the square of the velocity (Schlichting, 1955; Kalinske, 1949). Using the above expression for average shear stress and this empirical relationship between shear stress and velocity, we are now in a position to determine changes in bottom shear stress caused by velocity fluctuations. We can express the empirical relationship between shear stress and velocity as $\overline{\tau}_{xy} = c \cdot u^2$ (6)
where c is the constant of proportionality. Assuming "c" can be evaluated, shear stress can be estimated directly from the horizontal component of velocity. This constant was determined in the following manner.

For each mean velocity used in the study, a series of horizontal fluctuations were generated in the manner described above in the section concerning determination of instantaneous flow velocity. These fluctuations were used in equation 5 to determine the average level of the shear stress. At the same time, the instantaneous magnitude of the velocity associated with each pulse was recorded. A series of instantaneous values for bottom shear stress were generated according to equation 6. This process was allowed to continue for a set number of velocity fluctuations.

An alternative expression for average shear stress can now be written:

$$\bar{T}_{xy} = \frac{C u_1^2 T_1 + C u_2^2 T_2 + \cdots \quad C u_n^2 T_n}{T}$$

where $\bar{T}_{xy}$ is the average shear stress
c is constant of proportionality
u_i is magnitude of the i^{th} velocity squared
T_i time i^{th} pulse persisted at a point
T total amount of time represented by n velocity fluctuations.

The constant of proportionality can be calculated according to the following relationship:

$$c = \bar{T}_{xy} \cdot T / \sum_{i=1}^{n} u_i^2 T_i$$

where n is the number of iterations.

Using this heuristically derived constant, the shear stress can be calculated and used to determine the competency of the system e.g., the largest grain size the system can move.

TURBULENCE-SEDIMENT INTERACTION

With the fixing of the horizontal component of velocity, the results of the preceeding section define the shear stress and time duration associated with that shear stress. If the relationship between shear stress and grain size can be established, it, along with the duration will allow us to translate the time series of turbulent velocities into a sequence of depositional or erosional events, involving specific grain sizes.

The relationship between flow velocity and critical grain size has been approached from many directions. A wide variety of experiments have shown that as velocity of flow over a bed of sediment is increased, there is a critical point at which the sedimentary grains start to move.

Earliest studies of the relationship between the flow and the movement of sedimentary grains centered around the velocity. Brahms (1753) suggested the following relationship between velocity and grain size: $v_{cr} = k \cdot w^{1/6}$, where v_{cr} was the critical pick-up velocity, k a coefficient, and w the submerged weight of the particles.

Realizing it was physically more correct to use force per unit area rather than velocity as the critical variable for defining the competency of a stream, duBoys (1879), White (1940), Shield (1936), and other investigators developed a wide variety of relationships between shear stress and grain movement. These studies indicate that movement will not take place until a critical shear stress that is a function of grain size is exceeded.

The duBoys equation relates the tractive force to the product of the depth of water and gradient of the stream, $\tau = \gamma \; D \; S$, where γ is the specific weight of water, D is the depth of water, and S is the gradient of the stream. Rubey (1938) showed in laboratory experiments that at low velocities a grain of a given size moves only if the product of depth and gradient exceeds some critical value.

Shields (1936), using experimental data, proposed a dimensionless form of the stress:

$$\phi_{crit} = \frac{\tau_{crit}}{(\sigma - \rho) D}$$

where D is the grain size, σ and ρ are the specific weights of the particles and fluid respectively, and ϕ_{crit} is a constant for a particular level of turbulence.

White (1940), by equating the fluid forces that just cause movement of the grain to the force holding the grain in place, arrived at the following relationship to define the critical threshold stress:

$$T_c = c\rho' g\, d \tan\phi$$

where c is a constant that depends on the grain packing, ϕ is the angle of repose of the grains, and ρ' is the difference between the particle density and fluid density. This expression holds for the case in which the grain is mainly enclosed in the turbulent flow.

Kalinske (1947) extended this relationship for sand mixtures, arriving at the following expression:

$$T_c = (.25)\, p\rho' g\, d_i \tan\phi$$

where p is the portion of the bed area taking the shear, ρ' is the difference between the particle density and fluid density, g is the force of gravity, ϕ is the angle of repose of the grains, and d_i is the grain diameter. The size, d_i , thus determined represents the diameter of a particle, presumably quartz, which just moves from a position of cohesionless contact with fellow grains.

Kalinske's relationship between shear stress and grain size is the one utilized in the present model. This relationship is more thoroughly grounded in theoretical concepts than the more empirical relationships of others (Shield, 1936; Leliavsky, 1959). The empirically-based estimates were derived from a simple inspection of the result of increasing flume velocity over a substrate of known grain size. Such results, by their very nature, represent a sort of time average of the turbulent flow-sediment interaction. Because the individual events that constitute this average are the principal elements of this model, it is felt that using such estimates would be inappropriate.

Assumptions for the Operation of the Model

The relationships discussed above are enough to construct a process model for deposition in turbulent flow at a point in space. As can be seen, it is a very simple model, indeed probably one of the simplest that could be constructed. In order to view it in a realistic fashion, a number of assumptions which underly the simple model should be mentioned:

(1) The total volume of sediment in transport with respect to the volume of the transporting medium is small so that complex grain to grain interactions are at a minimum.

(2) The assumption of a critical diameter carries with it an implicit assumption that the grains have uniform shape.

(3) The model takes no account of small grain entrapment between larger grains.

Size-Frequency Distributions

In any localized portion of a dynamic sediment situation, the size-frequency distribution available for interaction with the turbulent flow is not uniform. That is, not all grain sizes are present in equal proportions . Obviously the shape of the size-frequency distribution and the location of the most abundant grain size classes with respect to mean velocity will have a tremendous bearing on the deposited sediment.

We are, in this investigation, concerned with erosion and deposition occuring at a point in space. The real world is composed of an infinite number of such points which can be envisioned as selectively feeding sediment from one to another. The more distant a point, the less influence it has on the grain size-frequency distribution occuring at the selected point of interest at a given time.

Later parts of this discussion will concern the nature of the deposit at a specific point. The sediment being supplied to that point can be envisioned as emanating from a great number of points nearby. The grain sizes departing from this neighboring cloud of points are all those sizes less than some critical size. With this in mind, the size-frequency distribution fed into the point of interest was simulated by providing the neighboring cloud of points with a uniform size-frequency distribution and assuming a fixed level of mean velocity passing a fraction of that distribution.

Each size-frequency distribution generated from each member of the cloud differs from each of the others only in a random fashion. The sediment that reaches the central point can be considered the sum total of these events and tends to conform to some sort of limiting distribution.

SEDIMENTARY DEPOSIT

Erosion and Deposition

Although both deposition and erosion occur during model iterations, the dominant process is that of deposition. This is due primarily to a "lag effect", with the coarser portions of previous deposits protecting the underlying strata from erosion by any but extreme turbulent events. Even when such an event occurs as discussed above, the higher velocities tend to operate with least duration and the length of time for scouring is very limited.

In the real world, such upbuilding is probably limited. Indeed, it is very likely this upbuilding effects evolution of the bottom configuration which fundamentally changes the hydrodynamics of the system. This change could conceivably result in a shifting of the average velocity, making scouring of the existing deposit likely.

Each velocity pulse which results in deposition is assumed to deposit a layer whose grain size is a function of velocity, and thickness is a function of duration. Thickness of each layer was determined from an analysis of the rate of sediment movement into the area of our depositional site. The unit of area of concern around our point is assumed to be one square centimeter. This area was judged to be great enough to completely capture the result of a depositional event occuring at its upstream edge. The sediment entering this area is assumed to spread uniformly through the area. Then, assuming a constant mineralogy, and that total mass of sediment delivered to the area is known, thickness can be easily calculated.

Kalinske (1947), on the basis of theory, held that for any single grain size, the rate of sediment transport can be determined from the following relationship:

$$G = C d_i p_i \lambda_g u_g$$

where G is the rate of sediment movement per unit width in grams / sec-centimeter
C is a constant
d_i is the grain size
p_i is the proportion of the bed occupied by grain size d_i
λ_g is specific gravity of the particles
u_g is the difference between the flow velocity and the velocity required to start movement.

If a size-frequency distribution is partitioned into a number of segments each covering a small grain size range, Kalinske's relationship may be applied to the mean size of each segment without too great a sacrifice in precision.

A given velocity pulse will cause the sediment in each segment of the distribution to move with a characteristic velocity. The distribution in question is that portion of the size-frequency distribution supplied by the upstream cloud of points coarser than that defined by the critical shear stress at our location.

The rate of accumulation of each size fraction within our area of deposition can thus be determined. Because the duration of the turbulent pulse is known, the total weight of each size fraction is simply determined. The relative proportions in the size classes constitute the size frequency distribution that has resulted from the depositional event. The thickness of the deposit is calculated by the relationship:

$$Th = \frac{(\Sigma X_i) \cdot 1.4}{2.65}$$

where Th is the thickness, X_i is the mass of sediment delivered for size class i, and assuming a density of quartz of 2.65 gm/cm3 and porosity of 40%.

The size intervals used in this model are

.01 millimeters wide. For most distributions in the sand range, this would result in 50-100 divisions of the size-frequency distribution.

Whereas the conditions for deposition can be defined from the above discussion, the conditions for erosion require knowledge of the previously deposited grain size-frequency distribution. In the case of the present model, a homogeneous grain size of 1 mm. was assumed to exist at the model site. This grain size was too coarse for the vast majority of random fluctuations to move.

After a few pulses, sediment deposits accumulated on this coarse substrate. This material is potentially erodable under the turbulent regime of the model. Erosion was judged to occur when the turbulent shear stress exceeded the critical stress of the mean grain size of the surface packet of sediment. This was judged appropriate, rather than size fraction by size fraction erosion, because empirical studies have shown that initiation of erosion involves bulk movement of the body rather than grain by grain sorting (Bagnold, 1956).

The thickness removed is calculated in a similar manner as described above for deposition.

For a given level of turbulent shear, erosion will proceed until a sediment packet is uncovered which has a mean grain size which is too coarse to be moved, or until the fluid packet associated with the shear stress passes by the depositional site.

Accumulation

A series of turbulent pulses are generated by the model and each is translated into either an erosional or depositional event. The primary output of the model is then a complete listing of events with associated grain sizes and thicknesses.

From this list which includes every hydrodynamic event which occured over the time range, a second shorter sequence was compiled, containing only those sediment packets still remaining after erosion. This second list constitutes the simulated sedimentary record. The characteristics of this simulated deposit can be compared to those of the turbulent medium. In this way, the objectives for construction of the model can be realized.

SUMMARY

The flow diagram for the model whose components are discussed above is illustrated in Appendix I.

Two parameters are under control by the investigator. The nature of the incoming size-frequency distribution which is controlled by the specific mean velocity passing through the upstream cloud of points, and the mean velocity at the model observation point.

The model is incorporated into a computer program, Appendix II. This program was constructed in such a manner that modification of any or all parts of it can be easily accomplished. That is, the fundamental logical structure of the model should always be suitable. The functional relationships within each logical element can be modified or replaced according to the used to which the model is put, and contemporary understanding of the nature of water/sediment interaction.

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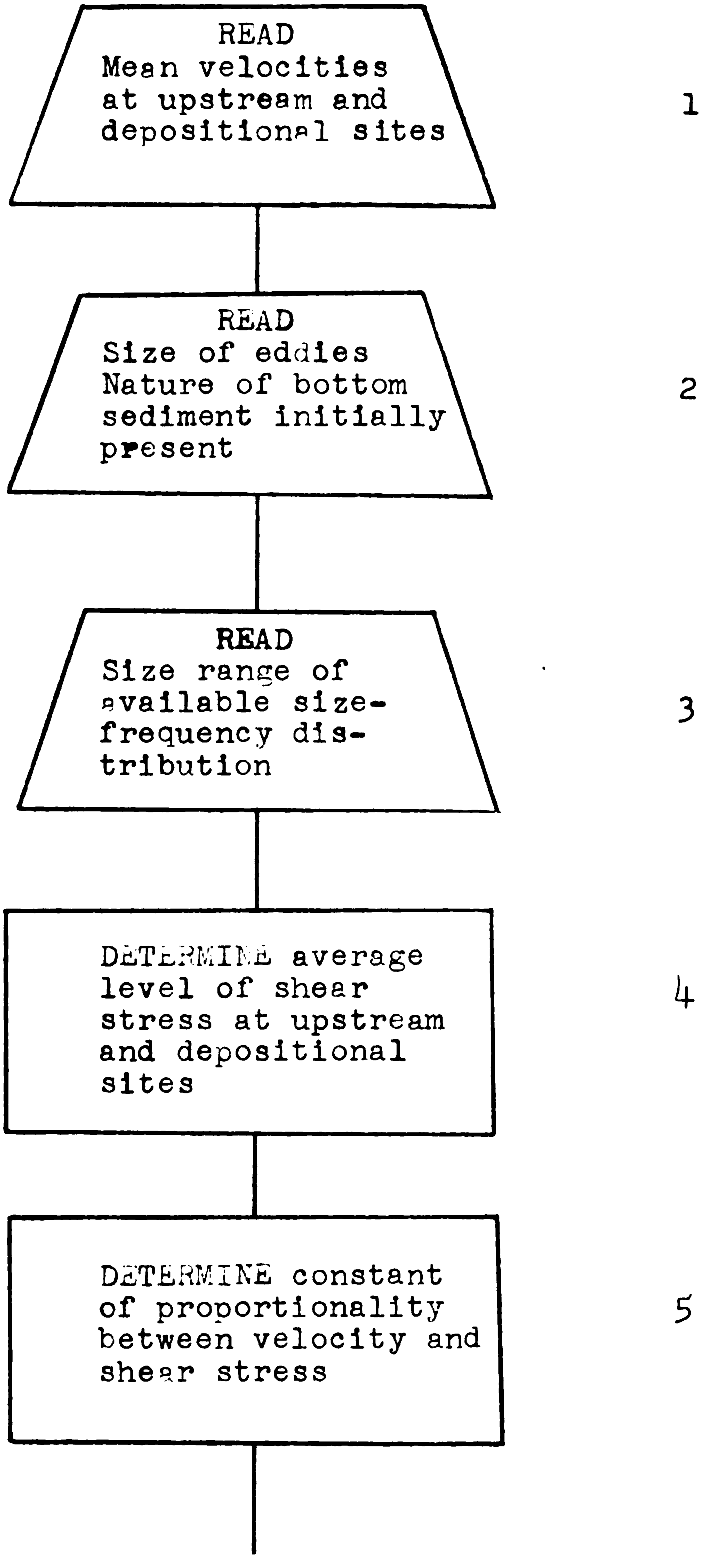
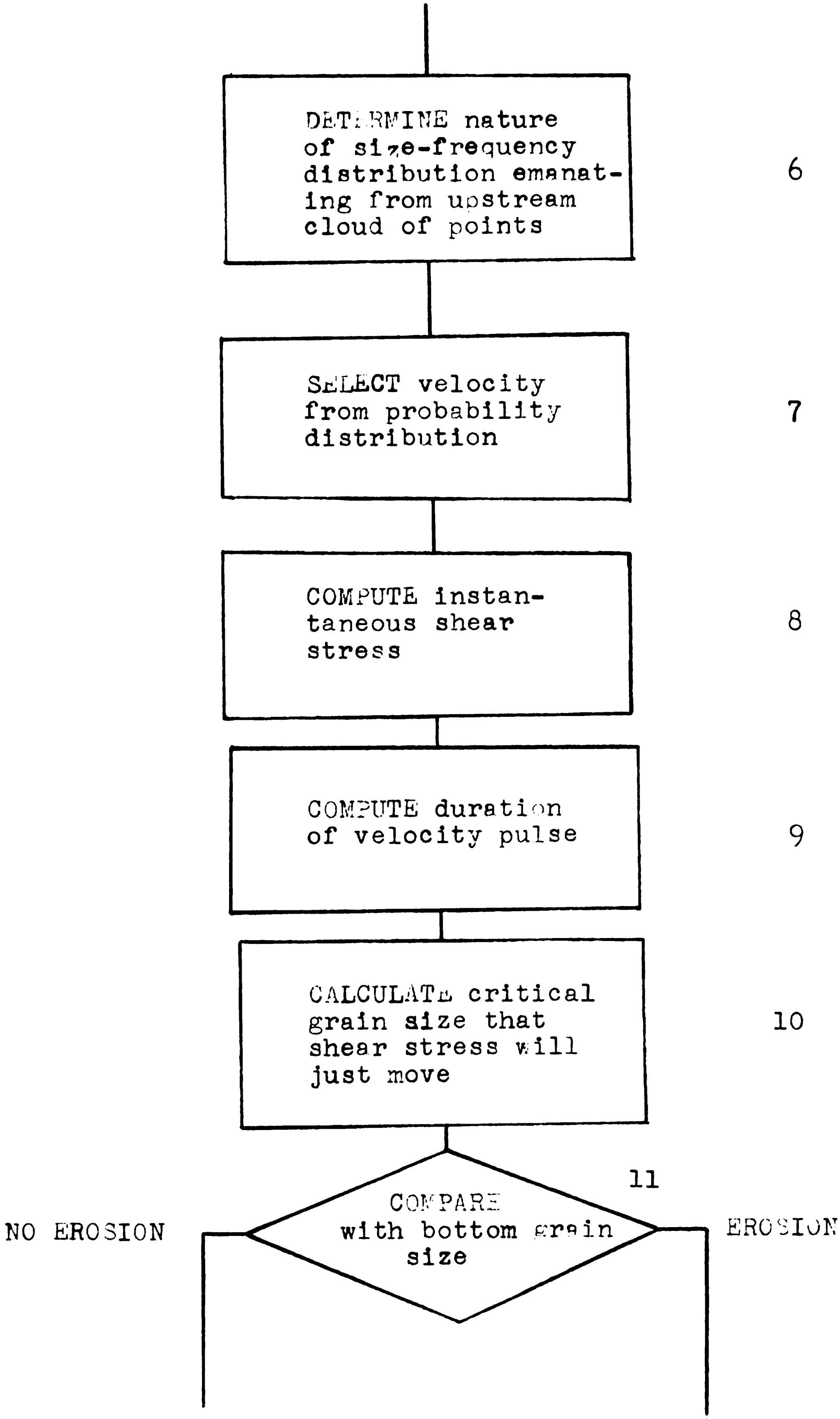
APPENDICES

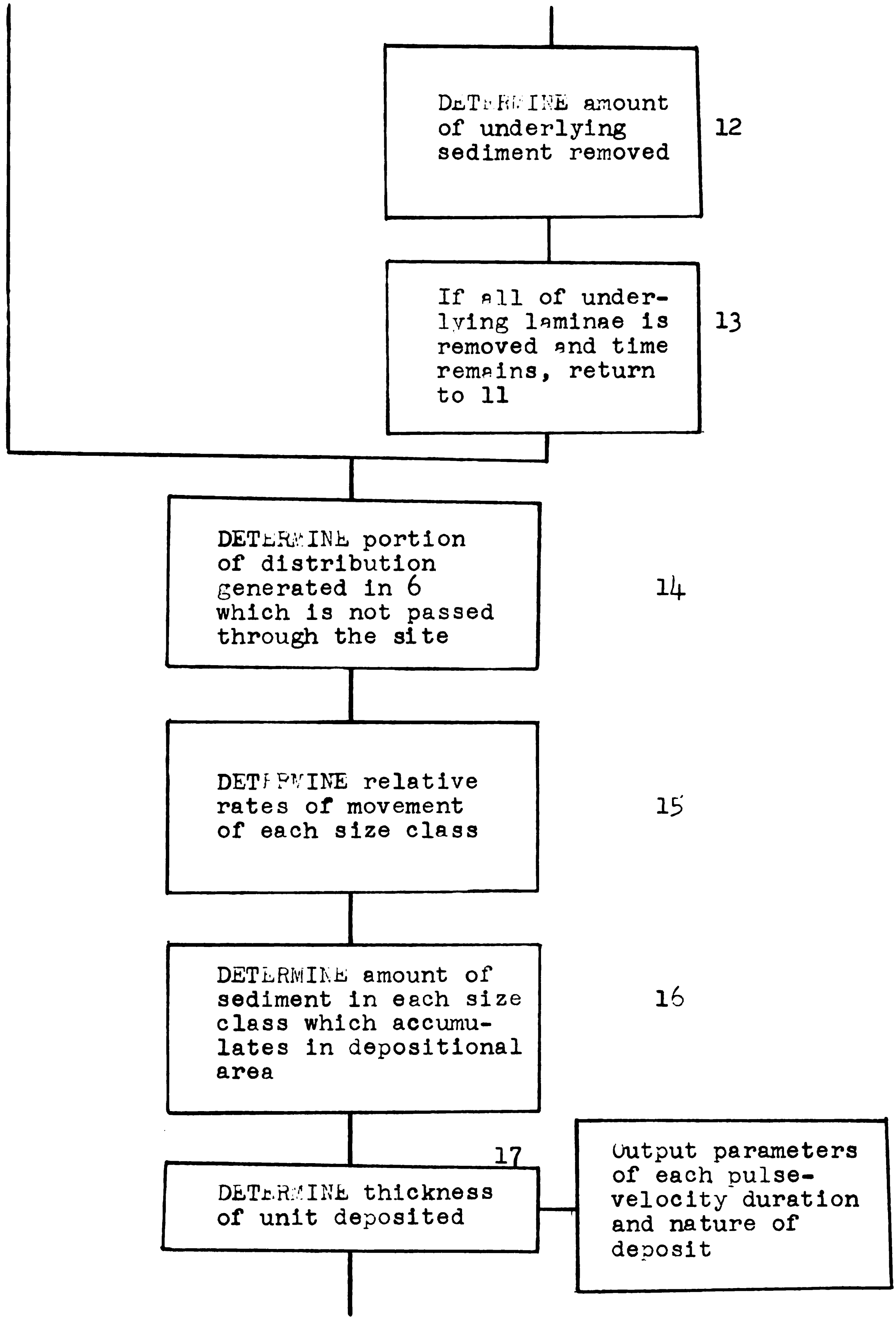
APPENDIX I

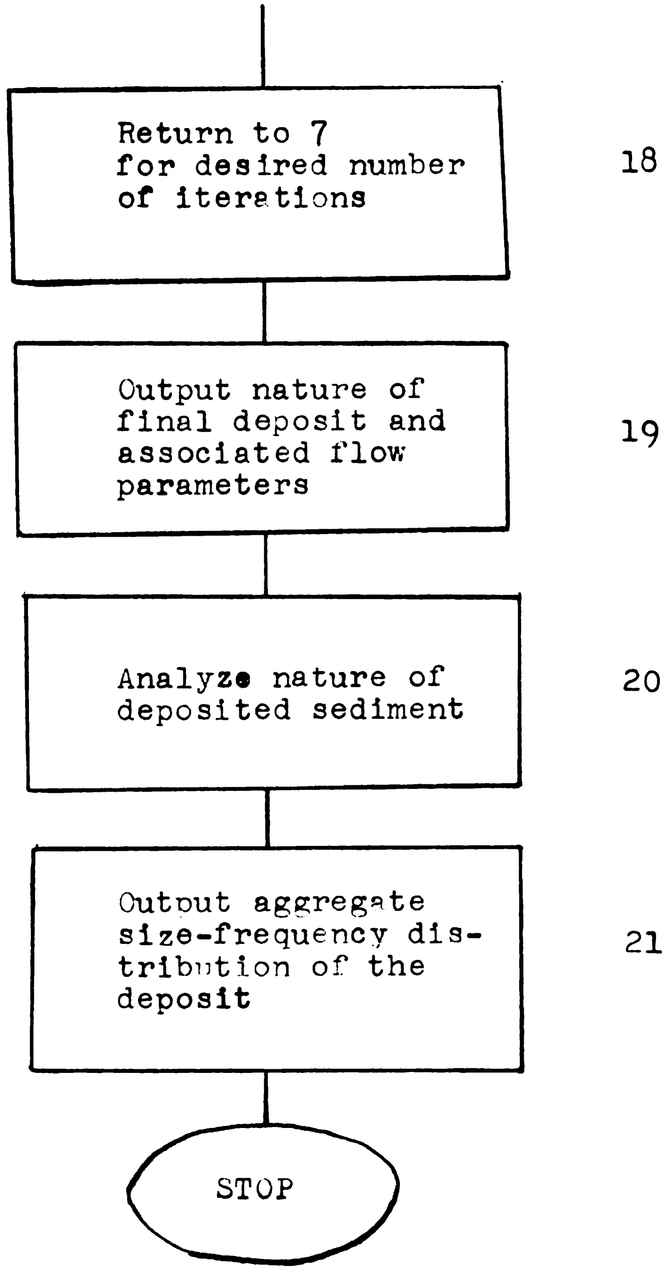
FLOW CHART FOR PROCESS-
RESPONSE MODEL

FLOW CHART FOR PROCESS-RESPONSE MODEL




DETERMINE nature of size-frequency distribution emanating from upstream cloud of points
6
SELECT velocity from probability distribution
7
COMPUTE instantaneous shear stress
8
COMPUTE duration of velocity pulse
9
CALCULATE critical grain size that shear stress will just move
10
11
COMPARE with bottom grain size
NO EROSION
EROSION


DETERMINE amount of underlying sediment removed
12
If all of underlying laminae is removed and time remains, return to 11
13
DETERMINE portion of distribution generated in 6 which is not passed through the site
14
DETERMINE relative rates of movement of each size class
15
DETERMINE amount of sediment in each size class which accumulates in depositional area
16
17
DETERMINE thickness of unit deposited
Output parameters of each pulse-velocity duration and nature of deposit


Return to 7 for desired number of iterations
18
Output nature of final deposit and associated flow parameters
19
Analyze nature of deposited sediment
20
Output aggregate size-frequency distribution of the deposit
21
STOP

APPENDIX II

LISTING OF COMPUTER PROGRAM

LISTING OF COMPUTER PROGRAM

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DEPOSE                                              CDC 6400 FTN V3.0-L284 OPT=1  01/31/72

      PROGRAM DEPOSE(INPUT,OUTPUT,TAPE60=INPUT,TAPE61=OUTPUT)
      DIMENSION U(200),D(500),G(500),SOD(500),HID(500),U(500),ICONT(500),
     1WTPERC(500,100),VOL(100),WTTOT(100),DMEAN(500)
      REAL KT,LOGRAIN,NUM
      READ(60,1)D(1),UXBAR,RHO,KT,G(1),N,UXBARUP,IRANF
    1 FORMAT(5F7.3,I5,F7.3,I5)
      READ(60,315)LOGRAIN,HIGRAIN
  315 FORMAT(2F5.3)
      READ(60,316)NN
  316 FORMAT(I5)
C THE FOLLOWING PROGRAM MIMICS BEHAVIOR OF A TURBULENT FLUID AT A POINT IN SPACE
C THROUGH TIME, THE FUNDAMENTAL ASPECTS OF THE FLOW ARE VELOCITY-SHEAR STRESS,
C SIZE OF THE DISCREET FLUID PACKETS, AND DURATION OF THE GIVEN VELOCITY PULSE
C AT A POINT.  BECAUSE OF THE TURBULENT NATURE OF THE FLOW, THE VELOCITY AT A
C POINT IS NOT CONSTANT BUT FLUCTUATES ABOUT A MEAN VALUE;
C D(1) DEFINES MEAN GRAIN SIZE OF SEDIMENT PRESENT WHEN MODEL STARTS;
C G(1) IS THE THICKNESS OF THAT LAMINAE.
C UXBAR IS THE MEAN VELOCITY OF FLOW AT THE POINT OF INTEREST.
C UXBARUP IS THE MEAN VELOCITY OF FLOW AT THE UPSTREAM CLOUD OF POINTS,
C N IS THE NUMBER OF VELOCITY FLUCTUATIONS ALLOWED,
C RHO IS THE DENSITY OF WATER.
C KT IS THE WAVENUMBER OF THE EDDY WHICH CONTRIBUTES MOST TO SHEAR STRESS;
C IRANF DETERMINES WHERE IN RANDOM NUMBER TABLE PROGRAM STARTS TAKING RANDOM
C DIGITS.
      DO 1500 KM=1,IRANF
 1500 AA=RANF(=1)
C AVERAGE SHEAR STRESS NEAR THE BOUNDARY IS CALCULATED IN THE FOLLOWING
C STATEMENTS.
C STDX DEFINES STANDARD DEVIATION OF THE VELOCITY DISTRIBUTION.
C KALINSKE SHOWED THE VELOCITY DISTRIBUTION WAS NORMALLY DISTRIBUTED WITH
C STDX=1/4 OF THE MEAN,
      STDX=UXBAR/4.0
      SUMFLUC=0.0
      VELOCSQ=0.0
      DO27I=1,250
      SUM=0.0
      DO 41 K=1,12
   41 SUM=SUM+RANF(=1)
C UOFM IS VELOCITY PICKED FROM THE DISTRIBUTION,
      UOFM=STDX*(SUM-6.0)+UXBAR
      VELOCSQ=VELOCSQ+UOFM**2
C FUOFM IS THE FLUCTUATION ABOUT THE MEAN VELOCITY,
      FUOFM=EX-UOFM
      FUOFM=FUOFM**2
C SUMFLUC IS THE SUM OF FLUCTUATIONS ABOUT THE MEAN,
   32 SUMFLUC=SUMFLUC+FUOFM
   27 CONTINUE
C FUOFMAV IS THE AVERAGE FLUCTUATION ABOUT THE MEAN VELOCITY FOR 250 EVENTS;
      FUOFMAV=SUMFLUC/250
C TAUTBAR IS THE AVERAGE SHEAR STRESS NEAR THE INTERFACE;
      TAUTBAR=(.2*FUOFMAV)
      VELOSQM=VELOCSQ/250
C USING THE CENTRAL LIMIT PROCEDURE, INSTANTANEOUS VALUES OF VELOCITY CAN BE
C DERIVED FROM THE VELOCITY DISTRIBUTION;
C EXPERIMENTAL EVIDENCE HAS SHOWN THAT SHEAR STRESS VARIES AS VELOCITY
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C SQUARED, (TAU=CONSTANT TIMES VELOCITY SQUARED.)
C FOLLOWING EVALUATES CONSTANT OF PROPORTIONALITY.
        CONSTA=TAUTBAR/VELOS2M
C FOLLOWING STATMENTS DETERMINE THE NATURE OF THE INPUT DISTRIBUTION;
C PROGRAM ASSUMES A FINITE NUMBER OF SITES AROUND THE DEPOSITIONAL SITE.
C SITES ARE ASSUMED TO BE FAR ENOUGH APART SO THAT THE VELOCITIES AT TWO SITES
C ARE NOT CORRELATED.
C UNIFORM SIZE DISTRIBUTION IS FED INTO EACH SITE.
C MM IS THE INDEX, EACH ITERATION A NEW VALUE FOR THE HORIZONTAL COMPONENT OF
C VELOCITY IS SELECTED - U(MM).
C EACH ITERATION REPRESENTS THE ENERGY CONDITIONS AT ONE POINT.
C THE COARSEST GRAIN SIZE THAT WILL NOT MOVE IS COMPUTED;  ALL GRAIN SIZES FINER
C THAN THIS ARE ASSUMED TO BE PASSED ON.  SIZE FREQUENCY DISTRIBUTION PASSED
C TO DEPOSITIONAL SITE IS ASSUMED TO BE THE SUM OF THESE INDIVIDUAL
C DISTRIBUTIONS.
      AVGRAIN=TAUTBAR/141.48
      WRITE(61,317)UXBAR
  317 FORMAT(1H1,*AVERAGE VELOCITY AT SITE=*,F7.3)
      WRITE(61,302)TAUTBAR
  302 FORMAT(1H0,*AVERAGE SHEAR STRESS AT INTERFACE=*,F7.3)
      WRITE(61,303)AVGRAIN
  303 FORMAT(1H0,*CRITICAL GRAIN SIZE MOVED BY THIS STRESS=*,F7.4)
      WRITE(61,304)LOGRAIN,HIGRAIN
  304 FORMAT(1H0,*RANGE OF GRAIN SIZES PRESENT*,F7.5,*-*,F7.5,*CM*)
      MM=2
   31 SUM=0.0
      DO 300 I=1,12
  300 SUM=SUM+RANF(-1)
      U(MM)=STDX*(SUM-6.0)+UXBARUP
C COMPUTE THE INSTANTANEOUS SHEAR STRESS
      TAUT=CONSTA*(U(MM)**2)
C SIZE OF EDDY OF INTEREST -- LOFM.
      LOFM=1.0/KI
C COMPUTE DURATION OF PULSE - TOFM
      TOFM=LOFM/U(MM)
C COMPUTE THE LARGEST GRAIN SIZE THAT WILL MOVE - D(MM),
      D(MM)=TAUT/141.48
C ASSUMING A GIVEN RANGE OF INPUT GRAIN SIZES ARE AVAILABLE, COMPUTE THAT
C PORTION OF THE DISTRIBUTION PASSED ON.
C B(I) REFERS TO INDIVIDUAL SIZES WITHIN THE INPUT DISTRIBUTION.
      NLO=LOGRAIN/.001
      B(NLO)=LOGRAIN
      NHI=HIGRAIN/.001
      DO 200 I=NLO,NHI
      B(I)=I*.001
      IF(B(I)-D(MM))200,200,3
  200 CONTINUE
    3 K=I
C HAVING DETERMINED THOSE GRAIN SIZES THAT WILL MOVE THROUGH THE SYSTEM, WE
C NOW NEED TO KNOW THE RELATIVE RATES OF MOVEMENT OF EACH SIZE CLASS;
C ASSUME A UNIFORM DISTRIBUTION BY WEIGHT AS BEING AVAILABLE.  COMPUTE THE
C INDIVIDUAL VOLUMES OF EACH GRAIN SIZE CLASS MOVED DURING THE PERIOD OF THE
C PULSE.  TOTAL VOLUME MOVED IS THEN THE SUM OF THE INDIVIDUAL COMPONENTS;
C KALINSKE DETERMINED THAT THE RATE OF SEDIMENT MOVEMENT WAS DEFINED BY THE
C FOLLOWING RELATIONSHIP -- RATE = K*D*(USHEAR-UC) WHERE K IS A CONSTANT,
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C D IS GRAIN SIZE, USHEAR IS SHEAR VELOCITY, AND UC IS CRITICAL SHEAR VELOCITY
C REQUIRED TO MOVE GRAIN SIZE D.
      DO503L=NLO,NHI
 503  VOL(L)=0.0
C NCLAS DEFINES THE NUMBER OF GRAIN SIZE CLASSES BEING PASSED ON.
      NCLAS=NHI-NLO+1
C WTP IS THE WEIGHT PERCENT OF EACH CLASS.
      WTP=1/NCLAS
C CONST IS THE SUM OF THE RECIPROCALS OF GRAIN SIZE.
      CONST=0.0
      DO700I=NLO,NHI
 700  CONST=CONST+1/B(I)
      DO701I=NLO,NHI
C P IS THE PROPORTION OF THE SHEAR STRESS TAKEN BY GRAIN SIZE B(I).
      P=.35*(1/B(I)/CONST)
C TAUTC IS THE CRITICAL STRESS WHICH WILL JUST MOVE GRAIN SIZE B(I).
      TAUTC=141.48*B(I)
      UC=SQRT(TAUTC/CONSTA)
C VOL(I) IS THE VOLUME OF GRAIN SIZE B(I) MOVED DURING THE PULSE.
C RATE OF MOVEMENT IS PROPORTIONAL TO THE DIFFERENCE BETWEEN THE VELOCITY AND
C THE SHEAR VELOCITY NEEDED TO MOVE THAT GRAIN SIZE.
 701  VOL(I)=B(I)*(U(MM)-UC)*.25*P*TOFM+VOL(I)
      SUMVOL=0.0
 500  DO501L=NLO,NHI
 501  SUMVOL=VOL(L)+SUMVOL
      DO502L=NLO,NHI
 502  TCONT(L)=VOL(L)/SUMVOL
      WRITE(61,306)UXBARUP
  306 FORMAT(1H1,*AVERAGE VELOCITY UPSTREAM=*,F7.4)
      WRITE(61,307)
  307 FORMAT(1H0,*SIZE FREQUENCY DISTRIBUTION INPUT TO DEPOSITIONAL SITE
     1*)
      WRITE(61,308)
  308 FORMAT(1H0,*   SIZE            FREQUENCY*)
      DO170K=NLO,NHI
 170  B(K)=K*.001
      DO309L=NLO,NHI
  309 WRITE(61,310)B(L),TCONT(L)
  310 FORMAT(1H0,F7.4,13X,F7.5)
C HAVING DETERMINED THE INPUT DISTRIBUTION, WE MOVE ON TO THE POINT OF INTEREST.
      M=2
C M IS THE ITERATION - EACH ITERATION A HORIZONTAL COMPONENT OF VELOCITY IS
C SELECTED. THE INPUT DISTRIBUTION FROM ABOVE IS FED INTO THE AREA. THOSE
C GRAINS COARSER THAN A CRITICAL SIZE ARE ASSUMED TO BE DEPOSITED.
C THE BOTTOM SEDIMENT IS ASSUMED TO BE ERODED IF ITS MEAN GRAIN SIZE IS LESS
C THAN THE CRITICAL SIZE
C TAVA IS THE MINIMUM SHEAR STRESS REQUIRED TO MOVE THE DISTRIBUTION.
      TAVA=B(NHI)*141.48
C UOFAMAX IS THE ASSOCIATED SHEAR VELOCITY.
      UOFAMAX=SQRT(TAVA/RHO)
      WRITE(61,311)
  311 FORMAT(1H1,*THE FOLLOWING OUTPUT GIVES THE PARAMETERS OF INTEREST
     1FOR EACH VELOCITY PULSE*)
      WRITE(61,312)
 312  FORMAT(1H0,*ITERATION    VELOCITY    DURATION    MEAN GRAIN SIZE    TH
```

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     1ICKNESS*)
   2 SUM=0.0
     DO21 I=1,12
  21 SUM=SUM+RANF(-1)
     U(M)=STDX*(SUM-6.0)+UXBAR
C TAUT IS SHEAR STRESS AT THE BOUNDARY,
     TAUT=CONSTA*(U(M)**2)
 401 TOFM=1/(U(M)*KI)
     LOFM=1.0/KI
C DETERMINE THE LARGEST GRAIN SIZE THAT WILL MOVE
C TAUT/141.48 DEFINES THE LARGEST GRAIN SIZE THAT WILL MOVE IN CM;
     D(M)=TAUT/141.48
C THE FOLLOWING STEPS DETERMINE EXTENT OF EROSION,
C PROGRAM CHECKS BOTTOM GRAIN SIZE -- IF D(M) IS GREATER -- EROSION TAKES PLACE;
C IF ALL OF THAT HSU IS REMOVED AND THERE IS STILL TIME REMAINING, PROGRAM
C CHECKS UNDERLYING HSU -- THIS PROCESS CONTINUES UNTIL TIME RUNS OUT;
     TUSED=0.0
     K=M-1
     DO50I=1,K
     IF(G(M-I))50,50,318
  50 CONTINUE
 318 IERODE=I
C TUSED IS THE TIME USED.
     TOFM=TOFM-TUSED
     IF(D(M)-D(M-IERODE))23,22,22
C TAUTC IS CRITICAL SHEAR STRESS WHICH WILL ERODE THE UNDERLYING STRATA,
  22 TAUTC=D(M-IERODE)*141.48
C UC IS THE CRITICAL SHEAR VELOCITY THAT WILL JUST MOVE THE GRAIN SIZE OF THE
C UNDERLYING STRATA,
     UC=SQRT(TAUTC/CONSTA)
     RATE=D(M-IERODE)*(U(M)-UC)*,25*,35
     THICK=RATE*TOFM*1,4
     G(M-IERODE)=G(M-IERODE)-THICK
      IF(G(M-IERODE))205,23,23
 205 TUSED=(G(M-IERODE)+THICK)*2,65/(RATE*1,4)
     G(M-IERODE)=0,0
     GO TO 50
C HAVING DETERMINED THE EXTENT OF EROSION, NOW TURN TO ANALYSIS OF WHAT
C HAPPENS TO INPUT DISTRIBUTION,
C IF D(M) IS GREATER THAN MAXIMUM GRAIN SIZE PRESENT IN INPUT DISTRIBUTION, ALL
C OF THE DISTRIBUTION IS SWEPT THROUGH,
  23 IF(HIGRAIN-D(M))20,20,25
  25 ADD=0.0
     SUM=0.0
C IN OUR INPUT DISTRIBUTION, WE HAVE CLASSES BETWEEN .1 AND 1.0 MM,
C THE RELATIVE PROPORTIONS OF EACH CLASS WAS CALCULATED EARLIER AND STORED
C IN TCONT(L) WHERE L IS THE NUMBER OF THE CLASS;
C NEED TO DETERMINE THE CLASS OF THE FINEST GRAIN SIZE NOT MOVED,
     JJ=D(M)/.001
     IF(JJ.LT,NLO)JJ=NLO
     SUM=0,0
     DO12IC=JJ,NHI
  12 SUMM=SUMM+TCONT(IC)/B(IC)
     DO13IC=JJ,NHI
     P=,35*(TCONT(IC)/B(IC))/SUMM
```

```
      TAUTC=B(IC)*141.48
      UC=SQRT(TAUTC/CONSTA)
      VOL(IC)=B(IC)*(U(M)-UC)*.66*P*TOFM
      NUM=VOL(IC)*6.0/(3.14*(B(IC)**3))
      SNUM=B(IC)*NUM
   13 SSNUM=SSNUM+NUM
C COMPUTE THE AVERAGE GRAIN SIZE
      DMEAN(M)=SNUM/SSNUM
C SUM THE INDIVIDUAL VOLUMES
      SUMVOL=0.0
      DO14L=JJ,NHI
   14 SUMVOL=SUMVOL+VOL(L)
C DETERMINE THICKNESS BY ASSUMING GRAINS FILL IN AN AREA OF
C ONE SQUARE CENTIMETER, AND ASSUME A POROSITY OF 40 PER-CENT.
      G(M)=1.4*SUMVOL
C RECORD THE WEIGHT PERCENTS OF EACH SIZE CLASS IN EACH HSU.
      DO15IC=JJ,NHI
   15 WTPERC(M,IC)=VOL(IC)/SUMVOL
      JJL1=JJ-1
      DO16IC=NLO,JJL1
   16 WTPERC(M,IC)=0.0
C SOD(M) IS SMALLEST GRAIN SIZE PRESENT IN HSU M.
      SOD(M)=B(JJ)
C HID(M) IS LARGEST GRAIN SIZE PRESENT.
      HID(M)=B(NHI)
      GO TO 28
   20 D(M)=0.0
      G(M)=0.0
      SOD(M)=0.0
      HID(M)=0.0
   28 WRITE(61,313)M,U(M),TOFM,DMEAN(M),G(M)
  313 FORMAT(1H0,2X,I5,6X,F7.3,4X,F6.0,9X,F7.4,8X,F7.4)
      M=M+1
      IF(M-N)2,2,1000
 1000 WRITE(61,314)NN
  314 FORMAT(1H1,*THE FOLLOWING OUTPUT GIVES THE NATURE OF THE DEPOSITED
     1 SEDIMENT AFTER*,I5,*ITERATIONS*)
      WRITE(61,100)
  100 FORMAT(1H0,*ITERATION   MEAN GRAIN SIZE   SMALLEST GRAIN SIZE PRES
     1ENT   LARGEST GRAIN SIZE PRESENT    THICKNESS*)
      DO 1001 M=2,N
 1001 WRITE(61,324)M,DMEAN(M),SOD(M),HID(M),G(M)
  324 FORMAT(1H0,2X,I5,8X,F7.4,19X,F7.4,22X,F7.4,14X,F7.3)
C THE FOLLOWING IS DESIGNED TO ANALYZE THE AGGREGATE SIZE FREQUENCY DISTRIBUTION
C DEPOSITED. PROGRAM MUST FIRST SORT THROUGH EACH HYDRODYNAMIC EVENT AND
C DETERMINE THOSE EVENTS WHICH HAVE A DEPOSIT ASSOCIATED WITH THEM.
      DO605L=NLO,NHI
  605 WTTOT(L)=0.0
      KOUNT=0.0
      DO600M=2,NN
      IF(G(M))600,600,601
  601 DO 602IC=NLO,NHI
      KOUNT=KOUNT+1
  602 WTTOT(IC)=WTTOT(IC)+WTPERC(M,IC)*G(M)
  600 CONTINUE
```

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      SUM=0.0
      DO603IC=NLO,NHI
  603 SUM=SUM+WTTOT(IC)
      WRITE(61,322)
  322 FORMAT(1H1,*THE FOLLOWING IS AN ANALYSIS OF THE AGGREGATE SIZE FRE
     1QUENCY DISTRIBUTION DEPOSITED*)
      WRITE(61,321)KOUNT,NN
  321 FORMAT(1H0,I5,* EVENTS OUT OF *,I5,* HYDRODYNAMIC EVENTS ARE RECOR
     1DED*)
      WRITE(61,323)
  323 FORMAT(1HL,*          GRAIN SIZE      WT. PERCENT*)
      DO604IC=NLO,NHI
      WTPEC=WWTOT(IC)/SUM
  604 WRITE(61,320)B(IC),WTPEC
  320 FORMAT(1H0,10X,F8.4,5X,F8.4)
      END
```