CHAPTER TWO

CAUSES AND MECHANISM OF MUDFLOWS

2.1 Definition of mudflow

A mudflow is a process in which gravel, boulders and rocks, mixed with clay and water, move readily, almost like a liquid, down a slope. A variety of other terms, such as *debris flow*, *debris avalanche*, *mud avalanche*, *lahar*, *rocky mudflow*, *mudslide*, *earth flow*, etc., have also been used in technical journals and newspapers to describe this phenomenon. In many cases, relatively minor quantities of mud are involved in these processes, and hence, the term *debris flow* has been preferred to *mudflow* in many technical articles. Nevertheless, the term *mudflow* is still widely used by the news media and the general public to describe most such flows; hence, it is this term which has been retained here and should be understood in a broad sense, covering all types of flows referred to in this monograph.

2.2 Causes of mudflows

Most mudflows are triggered by one or more of the following agents:

(*a*) Volcanic activity;
(*b*) Heavy rainfall;
(*c*) Landslides;
(*d*) Earthquakes;
(*e*) Snow and ice melt;
(*f*) Breach of man-made or natural dams;
(*g*) Underground water (seepage).

Many large-scale mudflow disasters originated from recently deposited products of volcanic eruptions mixed with water; e.g., Mt. Agung (Indonesia, 1963), Mt. Kelut (Indonesia, 1919, 1966), Mt. St. Helens (USA, 1980), Nevado del Ruiz volcano (Colombia, 1985) and Mt. Pinatubo (Philippines, 1991), etc. A mudflow resulting from volcanic activities is more specifically known as a *lahar* (a term originating in Indonesia, designating a debris flow over the flank of a volcano). The water component of a lahar may be the result of intense precipitation (because of reduced absorption of the precipitation into the slopes of the volcano when it is covered by a blanket of deposited volcanic ejecta), snowmelt caused by volcanic activities, or the bursting of crater lakes (Smith and Lowe, 1991). (*Note: See Table 1.1 for additional details on these and other specific mudflows referred to in this monograph.*)

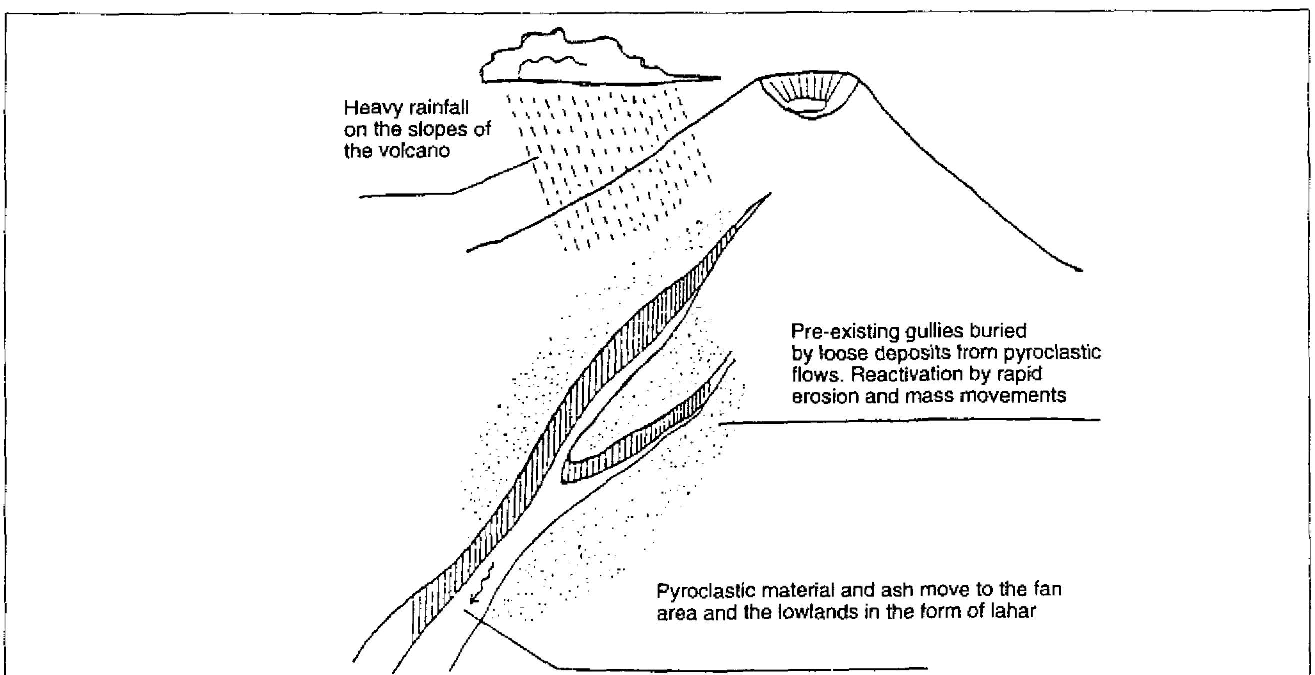

FIGURE 2.1 Lahar formation in pre-existing valleys filled with pyroclastic deposits (modified from Supangkat, 1989).

Mudflows occurring immediately after an eruption and containing hot pyroclastic material are known as *primary lahars* (or *hot lahars)*. Their water content is usually derived from the bursting of the crater lake or snowmelt. As huge volumes of hot material are transported by the primary lahars, they are responsible for most of the immediate destruction and loss of life. However, *secondary lahars* triggered by rain water can also transport large quantities of volcanic material, deposited in pre-existing gullies and channels over several years following the eruption, and can cause extensive damage. They are sometimes called *cold lahars*, in contrast to the hot lahars described above. *Figure 2.1* illustrates schematically the initiation of a lahar.

Mudflows in non-volcanic areas frequently originate from single or multiple landslides, which may be caused by a variety of factors; e.g., rain water, earthquakes, erosion, undercutting, underground water, etc. The debris (soil, rocks) brought down by landslides can turn into mudflows after mixing with river water. The debris then moves rapidly along the canyon. After emerging from the mouth of a gorge, the debris flows across an alluvial fan or cone and comes to a halt. Alluvial fans and cones are in fact the result of such repeated deposits of mudflow material.

Some examples of mudflow disasters resulting from landslides caused by heavy rainfall are Nagasaki (Japan, 1982), Wollinitzbach and Gradenbach (Austria, 1965, 1966). In the case of Mt. Huascaran (Peru, 1970) and Mt. Ontake (Japan, 1984), the mudflow disasters were triggered by earthquakes.

Dams may fail by overtopping due to heavy rainfall, earthquakes, gradual erosion, infiltration, avalanches, etc., leading to mudflows resulting from the sudden release of large quantities of water, sediment and/or sludge. The dams may be man-made or natural (such as morainal dams or those created by massive landslides). For example, an avalanche descending into a lake in the glacial valley in the Mt. Huascaran region (Peru, 1941) resulted in a morainal dam breach. This, in turn, caused the morainal dam of another lake downstream to burst. The resulting massive mudflow killed some 8,000 people. Examples of mudflow disasters due to the failure of man-made dams are Stava (Italy, 1985) and Izu Peninsula (Japan, 1978).

2.3. Initiation of mudflows

A mudflow may develop in an area where a combination of the following three prerequisites exists:

(a) *Gradient of slope*

The minimum slope of a terrain which is required for a mudflow to be initiated is, in general, about 15°. Local conditions such as the type of debris, base roughness, channel topography, etc., can modify this figure considerably. Takahashi (1978) made an analytical study of the upper and lower limits of slopes in which mudflows are possible. The upper limit is determined by the slope at which the shear failure of the debris occurs before the debris is fully saturated. In this situation, there is a sliding movement rather than a flow. This should thus be called a landslide rather than a debris flow or mudflow. A landslide can turn into a mudflow later, however.

The lower limit is defined by the slope beyond which the debris can no longer move even when it is fully saturated. In this case, there occurs only an ordinary dilute stream flow (or flood flow) which transports the sediment along its bed at a much reduced rate compared to a mudflow. The transitional types of flows between mudflows and dilute streamflows are called *hyper-concentrated flows* (Smith and Lowe, 1991). *Figure 2.2* shows these limits for a river channel draining Mt. Pinatubo (in the figure, the Greek letter θ represents the slope).

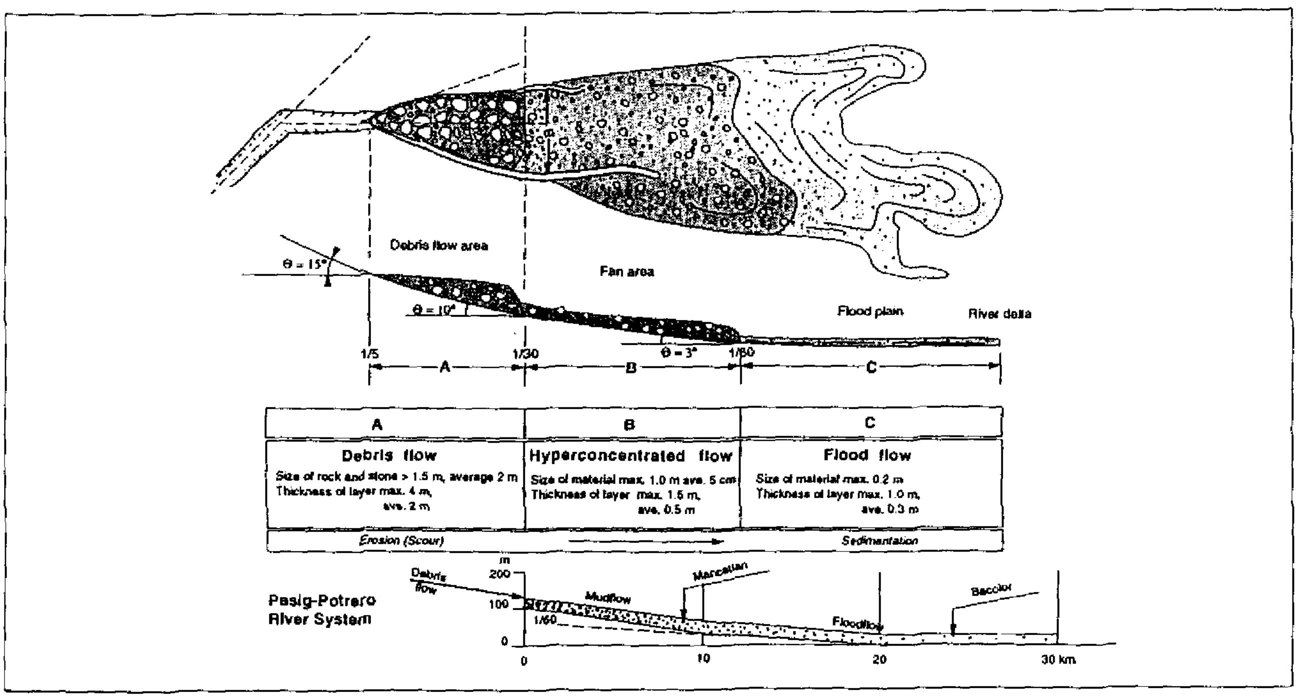

FIGURE 2.2 Characteristics of flow process as a function of river bed gradient.

(b) *Water content*

Water content is a key factor in the development of a mudflow. The water may come from intense rainfall, rapid snowmelt and/or the bursting of a natural or man-made dam. When the pore spaces between grains of the soil material become fully saturated and the high pore pressure reduces the internal cohesion of the material, a mudflow may start. More often, however, the rock fragments and soil brought down by a landslide become more compact as it continues to slide, and the voids between the solid grains become filled with water, which acts as a lubricant. The resulting fluidity of the material changes it into a mudflow. When a sliding mass does not have sufficient water content and the hill slope reduces before the mass reaches a torrent, it comes to a halt. If the hillsides are steep enough, such a landslide can still turn into a mudflow after mixing with the torrent water.

Many mudflows take place during or shortly after intense rainfall. The scatter diagram shown in *Figure 2.3* shows the relationship between the rainfall amount and duration for some reported mudflow incidents (Innes, 1983).

(c) *Potentially mobile material*

The third necessary condition for a mudflow to develop is the existence of potentially mobile material. The solid content of a mudflow may come from a landslide on an adjacent slope or a volcanic eruption. There may already be substantial amounts of potentially mobile solids on the torrent bed. In some cases, the breaching of a morainal or man-made dam supplies the necessary material. In general, a mudflow may contain not only boulders, rock fragments, mud and water, but miscellaneous debris such as branches or even whole trees, fragments of bridges, houses, crushed cars, etc., picked up along the way.

2.4. Characteristics of mudflows

2.4.1 Physical measurement of mudflows

How can one identify a mudflow and distinguish one from what is merely a fast-flowing mass of muddy water? The following are some important physical characteristics of a mudflow which give the observer of an approaching flow information which can help to determine whether or not it is in fact a mudflow:

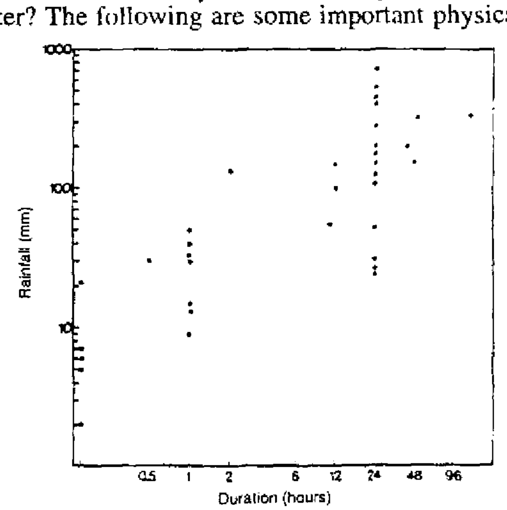

FIGURE 2.3 Scatter diagram illustrating the rainfall amount-duration relationship which has been reported as triggering a mudflow (Innes, 1993).

Speed:	The average maximum speed of a mudflow usually ranges from a few metres per second (e.g., 3 m/sec. for the 1991 Ormoc mudflow disaster in the Philippines), to tens of metres per second (especially on steep slopes).
Volume:	A mudflow usually consists of a large volume of dislodged material. For example, in the case of the Vincent mudflow disaster (Italy, 1963) the volume was 250 million m^3. In the case of the Murgab rock avalanche and mudflow (Tajikistan, 1911) triggered by an earthquake, it was 2.5 billion m^3.
Density:	In general, a mudflow is of high density. The mass of the mudflow materials per volume is usually measured in tons per cubic metre. It may range from 1.2 to 2.5 tons/m^3, and even higher. In the latter case, it can be categorized as a high-density mudflow, capable of transporting or removing (floating) bridges, steel structures, foundations, etc.
Energy:	The kinetic energy of the dislodged material in a mudflow is normally very high. It can be determined by the formula: $$E = \frac{M V^2}{2}$$ (where E is kinetic energy, M is mass, and V is velocity).
Discharge:	The maximum discharge rate of a mudflow is usually very high. It is measured in material per second. In the case of the Almaty mudflow, the recorded maximum discharge exceeded 10,000 m^3/sec. It may be even higher in the case of a dam failure or debris avalanche.
Boulders:	The speed at which the first boulders of the mudflow travel is normally very high; it may be close to the speed of a falling body. In the case of the Mt. Huascaran disaster (Peru, 1971) the first boulder's speed was assessed at 320 km/h. (*See the case history of the Mt. Huascaran mudflow in Chapter 8.*) These boulders may also be very large, as noted in the following paragraph.
Wave:	The front part or forerunner of a starting mudflow is usually composed of very large boulders or rock debris followed by mud. The size of these front boulders may be very impressive; in an extreme case, the recorded size of a transported boulder in the 1993 Nepal mudflow was 10 x 10 x 20 metres, weighing about 4,000 metric tons (the equivalent of around one hundred fully loaded railroad cars). (See also Section 2.4.3, below.)

2.4.2. Pulsating flow

A mudflow does not usually come in a single surge; it generally comes in a number of successive surges, as illustrated in *Figure 2.4* (Johnson, 1970; Pierson, 1980; Li, *et al.*, 1983).

Multiple surges may indicate the following phenomena higher up in the catchment:

(*a*) Initiation of mudflows at different spots along the upper water course;

(*b*) Occurrence of successive landslides;

(*c*) Successive collapses of lateral levees or successive arrival of large volumes of material in the flow path;

(*d*) Interim formation of natural dams due to landslides or the damming effect of drifting timber and the subsequent collapse of such hindrance to free flow.

However, the fact that a mudflow resulting from a single initiating event still consists of several large pulses, suggests an intrinsic instability of a mudflow in an open channel. (Niyazov and Degovets, 1975; Li *et al.*, 1983; Davies, 1986).

2.4.3. Capacity to carry large boulders

A mudflow is able to carry huge boulders, a property known as "carrying capacity" or "transport capacity". The extreme case of the 4,000 metric ton boulder transported in the 1993 Nepal mudflow was noted in Section 2.4.1. Takahashi (1981) also mentions a boulder of approximately 3,000 tons which was transported several kilometres by a mudflow in Japan. The front wave of a mudflow often includes very large boulders, as shown in *Figures 2.5* and *2.6*. Sometimes the momentum of these boulders causes them to tumble ahead of the main body of the flow, soon to be overtaken, however, and reincorporated into the front. The massive impact of the boulders often dislodges other boulders which are in the path of the oncoming mudflows, in which they too then become incorporated. The kinetic energy of the boulders is sometimes so great that upon hitting an obstacle, they leap several metres high and travel considerable distances. The awesome boulder-studded front wall accounts for most of the damage and destruction caused by a mudflow.

Three possible reasons have been put forward to explain the ability of a mudflow to carry large boulders, but the question remains an open one:

(a) *Cohesive strength of fine material*

The view that the cohesive strength of fine material (clay) supports the large boulders in a mudflow has been largely based on experimental work using artificial mudflows with large clay content (Hampton, 1975). However, in reality mudflows with high clay content are relatively rare; thus, this explanation is not plausible (Innes, 1983; Davies, 1986).

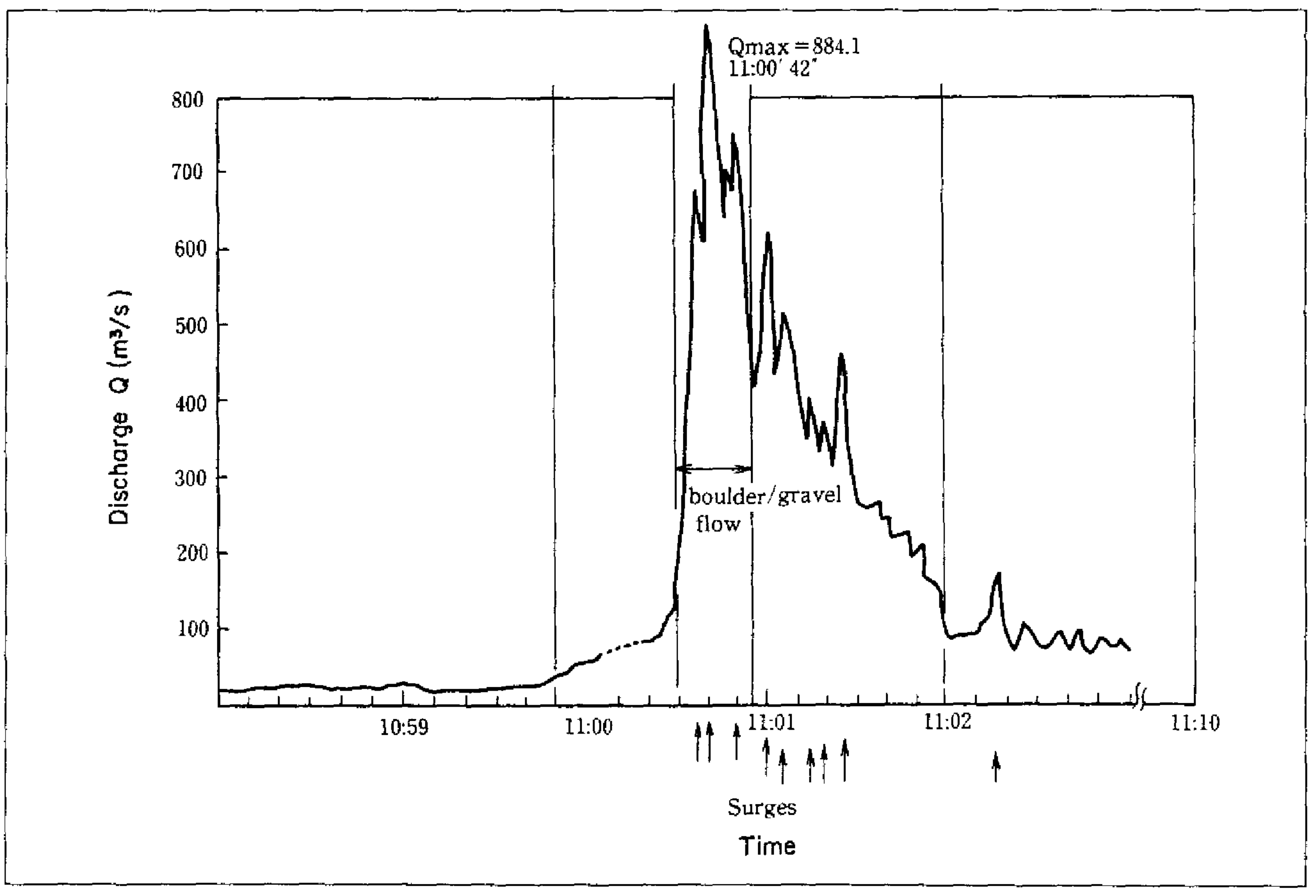

FIGURE 2.4 Hydrograph of a mudflow, monitored at Namerikawa, Japan (Ishikawa, 1985).

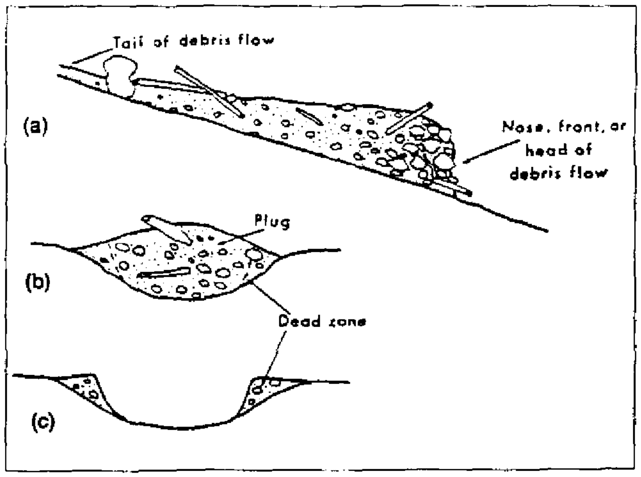

FIGURE 2.5 Characteristics of a moving debris flow:

(*a*) Longitudinal section showing variable concentration of solids in different parts of the flow.

(*b*) Cross-section through debris flow in motion, indicating ventral plug and marginal dead zones.

(*c*) Cross-section of torrent channel after passage of debris flow; note debris levees left behind by the flow.

(b) *Buoyancy due to excess pore pressure*

The second explanation for the carrying capacity of mudflows is that buoyancy due to excess pore pressure within the body of the flow can support large boulders (Hampton, 1979; Pierson, 1981). This is possible only when excess pore pressure develops in a debris mass and, further, when it is dissipated slowly during the mudflow. Nowadays, the buoyancy theory is not thought to be realistic.

It has also been suggested that buoyancy combined with the cohesive strength of the material could support large boulders. In any case, the reduced effective weight results in smaller frictional resistance. When compounded with the high-impact forces associated with mudflows, this may often explain the high transport capacity (Innes, 1983).

(c) *Dispersive pressure*

When solid particles are present within a shearing fluid, contacts between grains result in a tendency for these particles to be forced apart. This type of dispersive pressure is believed by many researchers to be responsible for the ability of mudflows to move large particles.

Bagnold (1954, 1968) argues that the dispersive pressure is proportional to the square of the particle size for a given shear rate. Therefore, the dispersive pressure causes the smaller particles to move towards the area of greatest shear, while the larger particles are forced to move towards the area of least shear; i.e., towards the surface. Since the flow surface moves fastest, the large solid particles or boulders which come to the surface drift to the front of the flow, as observed in many mudflows. Rodine (1974) and Naylor (1980) question the role of dispersive pressure in mudflows, while Lowe (1976), Takahashi (1980, 1981) and Davies (1986), among others, maintain that it is the main factor behind the transport capacity of mudflows. Further research work is required to determine the importance of dispersive pressure in mudflows.

It is clear that there are various possible mechanisms which could support large particles in mudflows. It could also be that each of them contributes to some extent to the boulder-carrying capacity of mudflows.

2.5. Mechanics of flow (rheology)

One early attempt to understand the mechanics of mudflows was in terms of "fluidization", in the sense used by chemical engineers. That is, it was suggested that mudflow debris becomes "fluidized", meaning it acquires the characteristics of a liquid, because the interstitial fluid moves upward so rapidly that granular particles become suspended in it. As a result, the debris mass flows like a fluid. However, experiments and field observations show that this is not the case, and that fluidization does not occur in mudflows in this way, because a relatively small content of water, always present in a mudflow, is itself sufficient to fluidize it. (Johnson and Rodine, 1984).

The second approach was to treat a mudflow as a case of shear failure, like that of soil, expressed by Coulomb's law in terms of cohesive strength and internal friction. Hence, the critical thickness of debris which can flow could be derived. This model of mudflow was also found to be unrealistic because a flow condition referred to as "quasi-static", required for shear failure, does not exist in a mudflow.

The current approach in the field of rheology of mudflows is to use the Coulomb-viscous model (Johnson, 1970; Johnson and Rodine, 1984). In this model, the viscous resistance term is added to the Coulomb equation. In a simplified form of this model, known as the Bingham model, the internal friction is neglected. This, however, leads to some discrepancies (Takahashi, 1981).

The Coulomb-viscous model predicts a central "plug" or "raft" of relatively rigid debris moving at a uniform velocity. In the transition zone between the moving plug and the stationary channel walls, the velocity of flow reduces parabolically to zero (*Fig. 2.7*). Johnson (1970) reports that the actual velocity distribution compares well with the one predicted by the Coulomb-viscous rheological model of mudflows.

2.6. Deposition of mudflows and geomorphological aspects

The process of mudflow deposition has a decisive impact on the future morphology of a valley. A mudflow slows down and comes to a halt once the gradient becomes sufficiently low. In general, this happens when the slope is less than 10°. The actual value of the gradient of the stopping slope depends on a number of factors, such as the volume, water content and material composition of the mudflow and the topographical features of the slope. Mudflows may, in some cases, continue to advance even where the gradient is as low as 3°

FIGURE 2.6 (*a*) View of a typical mudflow: Boulder-studded front which arrives without any precursory waves

FIGURE 2.6 (*b*) View of the mudflow: The same mudflow some time later showing the metre-sized material of the front. Location, Nagano, Japan. *Photo: Ministry of Construction, Japan.*

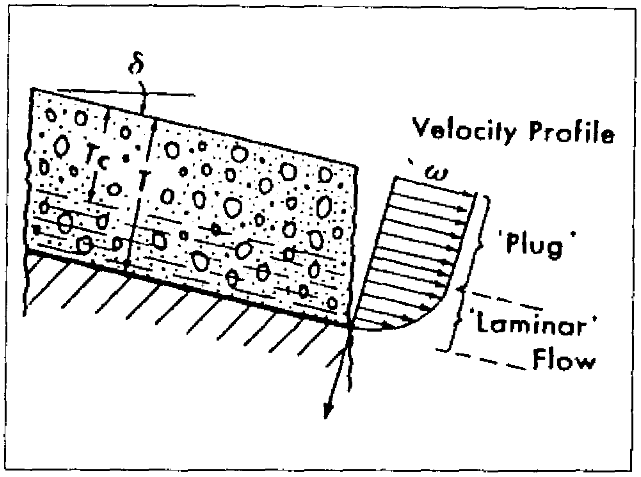

FIGURE 2.7 Velocity distribution in a mudflow. The observed profile compares well with the one predicted by the Coulomb-viscous rheological model of mudflows (Johnson and Rodine, 1984).

if they contain enough fine particles like volcanic ash, or when they are flowing over a flat and solid surface, such as a paved road, because of less resistance to flow.

The location beyond which the reduced gradient causes debris deposition is known as the *intersection point* (*Fig. 2.8*). As more debris continues to be deposited, the intersection point moves upstream. If subsequent flows pick up material and lower the bed, the intersection point moves downstream. The deposition of debris at and below the intersection point, also termed *aggradation*, causes the bed to rise. This reduces the carrying capacity of a river leading to overtopping and flooding of the surrounding areas. In an extreme case, the river bed may even become higher than the neighbouring terrain.

The boulder-studded front of a debris deposit has a lobate shape (Johnson, 1970; Innes, 1983). The front of the lobe is also referred to as a *snout* (*Fig. 2.9*). As discussed earlier (Section 2.4.3 (*c*), dispersive pressure is believed by many researchers to cause the boulders to rise to the surface where the velocity is higher, and hence, they drift to the front of the flow. Owing to the high concentration of boulders, the flow speed of the front tends to decrease, which could explain why the front part of a mudflow becomes swollen. This is reflected in the shape of a debris lobe after deposition.

The deposition process usually occurs at the exit from a gorge. The material brought down by successive mudflows forms a cone-shaped deposit as the transported material spreads out on emerging from the gorge. Such a deposit is called an *alluvial fan* (alluvial fans can be seen in *Figs. 2.2, 2.10* and *2.11*). A conical alluvial fan with steep slopes is also known as an *alluvial cone.* A series of debris lobes can have an appearance similar to the scales of a fish. If the topography permits, a mudflow may take a course lateral to an existing alluvial fan, which broadens as a result.

On a map, debris flow deposits can usually be recognized by their wavy form (*Fig. 2.12*). *Lateral deposits* form when snout material moves laterally and boulders are pushed to the sides (Sharp, 1942). Such lateral debris

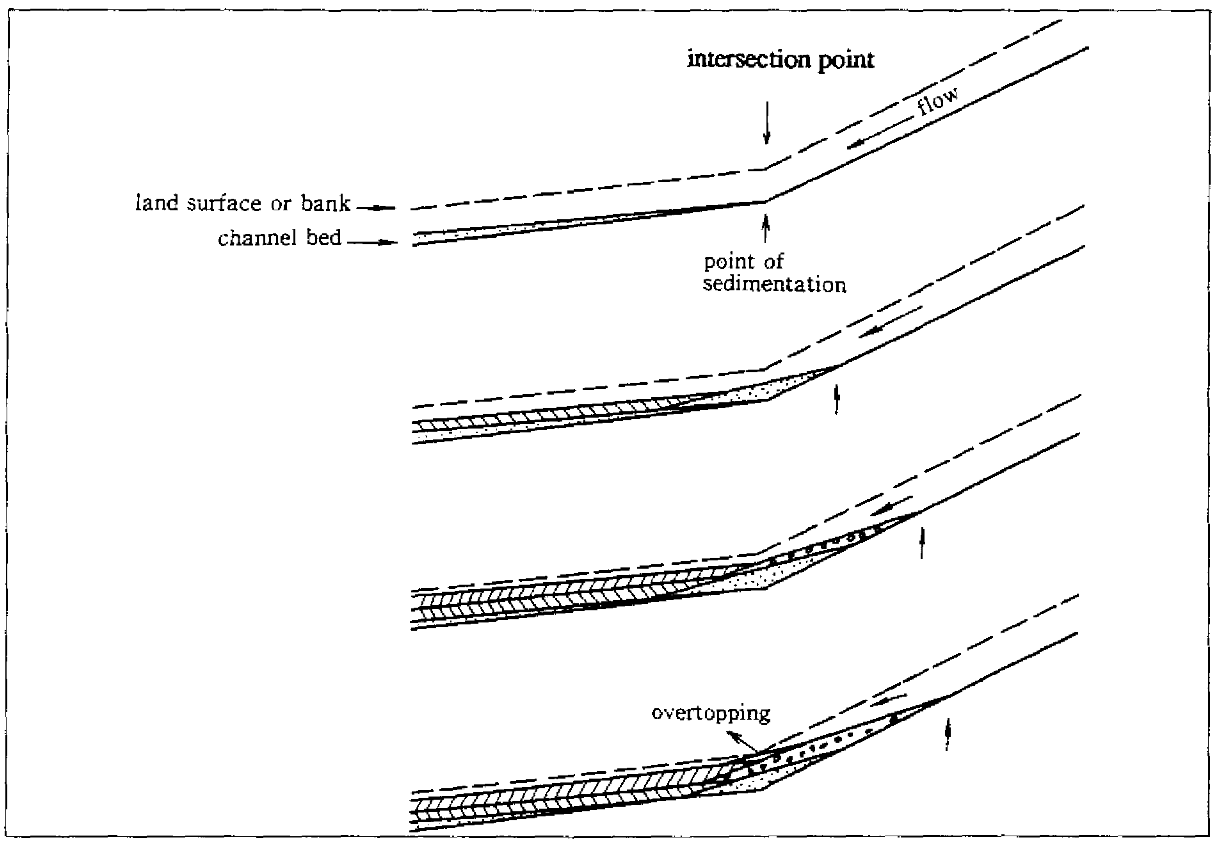

FIGURE 2.8 Mechanism of aggradation at an intersection point, leading to overtopping.

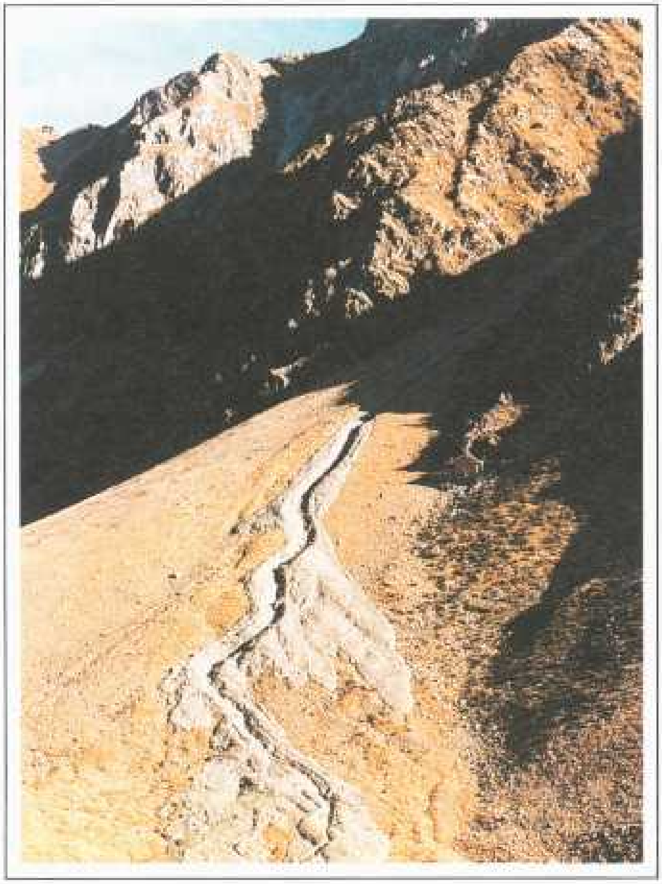

FIGURE 2.9 Development of debris lobes. Successive flows cut into previous lobes and descend further down to form new lobes. Location: Ticino, Switzerland. *Photo: M. Watanabe.*

deposits form natural levees (see *Fig. 2.11*). *Medial deposits* are those within the channel itself.

In *Figure 2.12*, cross-section A shows a snout at maximum lateral extension, corresponding to the maximum wave amplitude. The snout earlier passed the point represented by cross-section B, leaving lateral deposits when the mudflow surface subsided. Further up the channel the lateral deposits formed earlier by the snout are now being overridden by a succeeding wave (cross-section C). The snout and succeeding waves have already passed the point in the channel represented by cross-section D, and each wave caused the mudflow surface to rise and widen locally. Furthermore, each wave was identified by a layer of debris in the lateral deposits shown in cross-section D. If no waves followed those shown in *Figure 2.12*, the final mudflow deposit would appear approximately as shown in cross-section D, except that part of the medial material would have moved further down the channel

A river with relatively little sediment often erodes the deposits from previous mudflows and floods. In other words, they lower the central part of a channel bed. The remaining parts of the channel bed are left at a higher elevation. Repeated mudflow deposition and erosion can thus leave a step-like or terraced topography. Higher terraces may be safe from mudflow and flood discharges unless the magnitude of the discharge exceeds that of the previous flow. A terrace can be identified along a channel both in a gorge and on an alluvial fan. Terraces provide valuable evidence of past mudflows and of their material composition, frequency and magnitude.

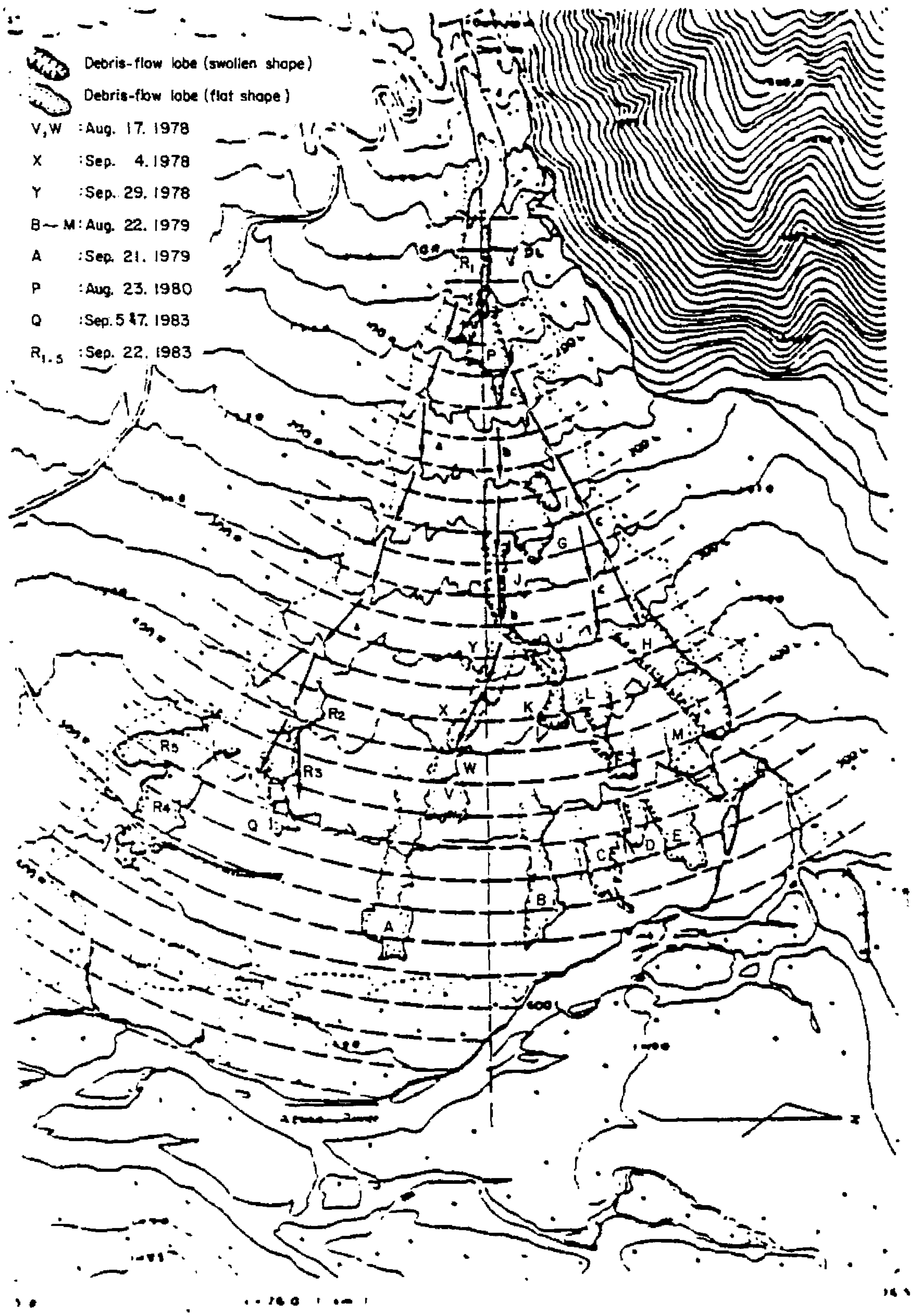

FIGURE 2.10 Distribution of debris flow lobes on an alluvial fan. Flow routes are shown by arrows (Suwa *et al.*, 1985).

FIGURE 2.11 Natural levee formation due to lateral debris deposits. Location: Ticino, Switzerland. *Photo: M. Watanabe.*

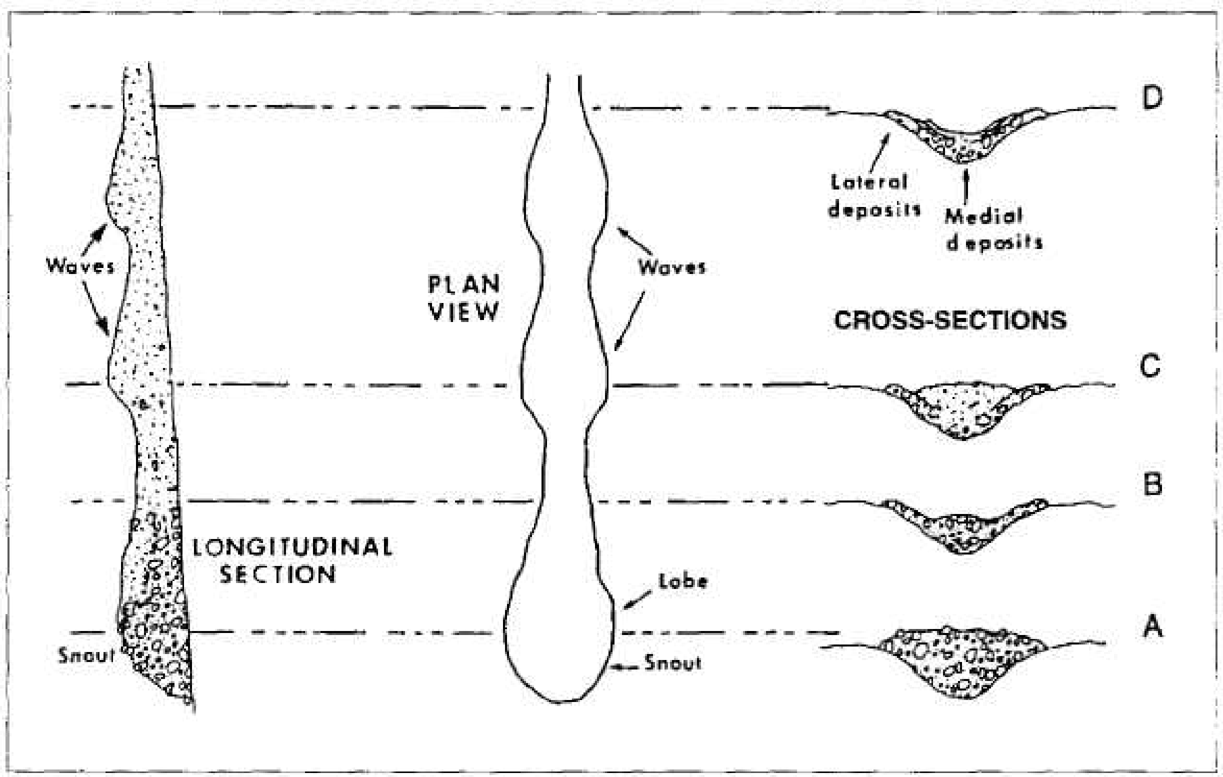

FIGURE 2.12 Idealized representation of a debris flow, showing waves and deposits formed by successive waves of debris (Johnson and Rodine, 1984).