Chapter 5. Bedforms and stratification under unidirectional flows

Introduction

When the flow of water (or air) over a bed of non-cohesive sediment is strong enough to move the particles of the bed material the bed becomes molded into some topographic form with vertical relief ranging from fractions of millimetres to up to several metres. The three-dimensional geometry of the bed topography is governed by the interaction of the fluid and the sediment. Experimental studies in flumes and in modern rivers and in intertidal areas have found that their are a few common forms of bed geometry that develop under unidirectional current and that the overall geometry depends on the flow conditions and the sediment properties. The overall geometry of such a bed is referred to as a **bed configuration** and it is made up of many individual topographic elements termed **bed forms** (or bedforms).

Over the past century much work has been directed at the description of the characteristics and behaviour of bedforms under unidirectional flows. This extensive research has been conducted by both sedimentologists and civil engineers for two very different reasons. Civil engineering is concerned with the effects of fluid flow on mobile sediment beds because this broad field includes the practical problems associated with moving water from one place to another (e.g., for hydroelectric generation, irrigation, or city water supplies). Open canals are often excavated in unconsolidated sediment and the flow of water through the canals may cause sediment to be transported and ultimately results in the formation of bedforms. The most fundamental problem with the development of bedforms in canals is that they alter the character of the flow and affect fundamental flow properties such as discharge. For example, in chapter 4 Eq. 4-15 shows that the velocity of a turbulent flow varies with the inverse logarithm of the size of the roughness elements on the boundary (y_o; i.e., the greater the boundary roughness the slower the velocity). Most bedforms comprise major roughness elements that are many times larger than the roughness due to the grains on the bed. Hence, the presence of bedforms will retard fluid flow and their presence and forms must be predictable in order to design canals on unconsolidated sediments. Sedimentologists have benefited greatly from the work of engineers in their study of bedforms. However, sedimentologists derive their interest from the fact that bedforms are commonly preserved on ancient bedding planes in sedimentary rocks. Bedforms are **primary sedimentary structures**; structures that form at the time of deposition of the sediment in which they occur and they reflect some characteristic(s) of the depositional environment. In addition, bedforms produce a variety of forms of cross-stratification (also primary sedimentary structures) that are very common in the geologic record. Because bedforms and their behaviour are governed by fluid processes, they, and their stratification, provide an unequalled basis for making paleohydraulic interpretations of ancient depositional environments. The work conducted by engineers, that attempted to predict the types of bedforms that develop on the basis of the sediment and flow characteristics, has been particularly useful in contributing to our ability to make paleohydraulic interpretations of bedforms and their associated stratification. The aspects of paleohydraulics that can be inferred from the forms of cross-stratification include the relative flow strength, the direction of the current, the type of current (e.g., upper or lower flow regime).

This chapter will focus on the types of bedforms that develop under unidirectional flows and emphasize aspects of their character and behaviour that are particularly useful in the interpretation of sediments and sedimentary rocks. In addition, the final sections of this chapter will describe the forms of stratification that the various bedforms produce.

Bedforms under unidirectional flows

Terminology

As we will see below, there are a variety of bedforms that are most precisely classified on the basis of their geometry and relationship to the water surface. However, the broadest classification of unidirectional flow bedforms is based on the **flow regime** under which the bedforms develop. This concept is widely used by sedimentologists and was introduced by Simons and Richardson (1961; a pair of civil engineers) who distinguished **lower flow regime** and **upper flow regime**, partly on the basis of the bedforms that are produced under unidirectional flows. Table 5-1 summarizes the main criteria for distinguishing these two flow regimes. Note particularly that the relationship

Table 5-1. Definition of the flow regime concept of Simons and Richardson (1961).

Flow Regime	Bedforms	Characteristics
Lower flow regime	Lower plane bed, Ripples, Dunes	• **F** < 0.84-1.0*; • low rate of sediment transport, dominated by contact load; • bedforms out-of-phase with the water surface.
Upper flow regime	Upper plane bed, In-phase waves, Chutes and pools	• **F** > 0.84 - 1.0*; • high rates of sediment transport, high suspended load; • bedforms in-phase with the water surface.

*Note that Simons and Richardson (1961) set **F** < 1.0 for lower flow regime and **F** > 1.0 for upper flow regime. However, subsequent work indicated that in-phase waves began to develop over the range 0.84 < **F** < 1.0. Because in-phase waves were particularly characteristic of the upper flow regime the limiting value of **F** has been adjusted accordingly here.

between the bedform and the water surface figures prominently in this scheme. The lower flow regime is dominated by bedforms that are out-of-phase with the water surface and the upper flow regime is dominated by bedforms that are in-phase with the water surface.

Figure 5-1 defines the general terms that will be used to describe bedforms in the following section. Terms for asymmetric bedforms (including ripples and dunes; these are out-of-phase with the water surface) are well-established and widely used. The descriptive terms for symmetrical bedforms (the in-phase waves) produced under unidirectional flows are of only limited scope. The in-phase waves have been relatively little studied and a more complete descriptive terminology will certainly follow the current peak in interest in this class of bedform.

The sequence of bedforms

By "sequence" of bedforms we refer to the hypothetical, sequential development of different bedforms on a mobile sediment bed with increasing flow strength (e.g., increasing velocity with constant flow depth). The concept of bedform sequence is useful in describing bedforms because it provides a basis for a qualitative appreciation for the relative flow strengths that are required to form them. As we will see later, not all of the bedforms describe in this sequence will develop on a bed of any one size of sediment; some bedforms are limited to coarse bed material while others are limited to fine bed material. Thus, we can consider the following to be a "hypothetical" sequence but one that provides valuable insight into the interpretation of bedforms and their stratification. Figure 5-2 (A and B) schematically illustrate the variety of bedforms, in sequence, with increasing flow strength. Note that the two sequences in figures 5-2 A and B form a continuum, more-or-less.

We can think of the sequence of bedforms in terms of the changes in bed geometry under a flow that is slowly and incrementally increasing in velocity, at constant depth, over a mobile sediment bed. Each specific bedform represents the equilibrium bed state for a constant flow velocity and depth. On beds of relatively coarse sand, just as the flow strength exceeds the threshold required for sediment movement, the first bed configuration will be a flat, planar surface. Such a bed is termed a **lower plane bed** to distinguish it from another type of plane bed that develops under higher flow strengths. The lower plane bed will only form on beds of sediment coarser than 0.7 mm and is characterized by its planar surface and relatively low rates of sediment transport (limited to contact load). The limitation of lower plane beds to relatively coarse sand indicates that this bed configuration will only form under dynamically rough turbulent boundaries (i.e., relatively large boundary Reynolds numbers). Sand that is deposited on a lower plane bed is characterized by relatively low angles of particle imbrication (both up- and downstream-dipping; see Fig. 5-3).

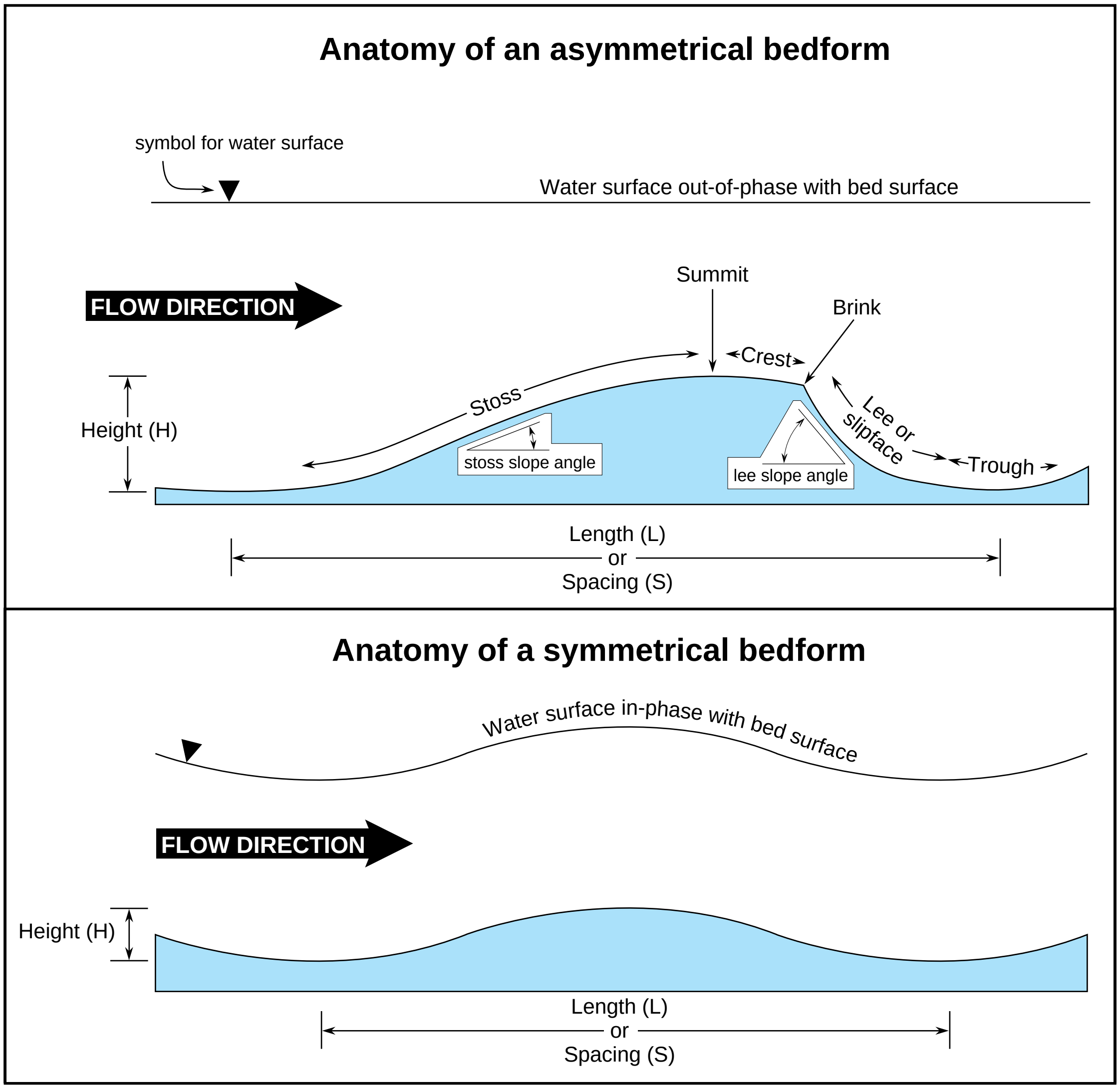


Figure 5-1. Definition of terms used to describe asymmetrical and symmetrical bedforms that develop under unidirectional flows.

On beds of sediment finer than 0.7 mm, just as the current exceeds the flow strength required for sediment movement, the bed will be molded into a series of small, asymmetrical bedforms termed **ripples** (some texts call these bedforms "small ripples"). Figure 5-4 shows the form of a ripple and the flow pattern that it induces near the boundary. Ripples behave like negative steps on a boundary and result in flow separation at the brink of the ripple and attachment just downstream of the trough where fluctuations in boundary shear stress are particularly large and erosion is intense. Ripples migrate downstream, in the direction that the current is flowing, as sediment is eroded from the trough and lower stoss slope and moves up the stoss slope to become temporarily deposited at the crest. As the deposit grows on the crest it eventually becomes unstable and avalanches down the lee slope where it is deposited. The avalanche deposit is subsequently buried by continued avalanching and the entire ripple form migrates downstream. In this way the pattern of flow separation, which is governed by the ripple geometry, also moves downstream, as does the site of erosion at the point of flow attachment. Over time, as the region of scour migrates downstream the sediment deposited previously on the lee slope is eroded from the bed and begins its cycle of transport all over again. Hence, migration results from the pattern of spatial variation of erosion and deposition along the length of the ripple.

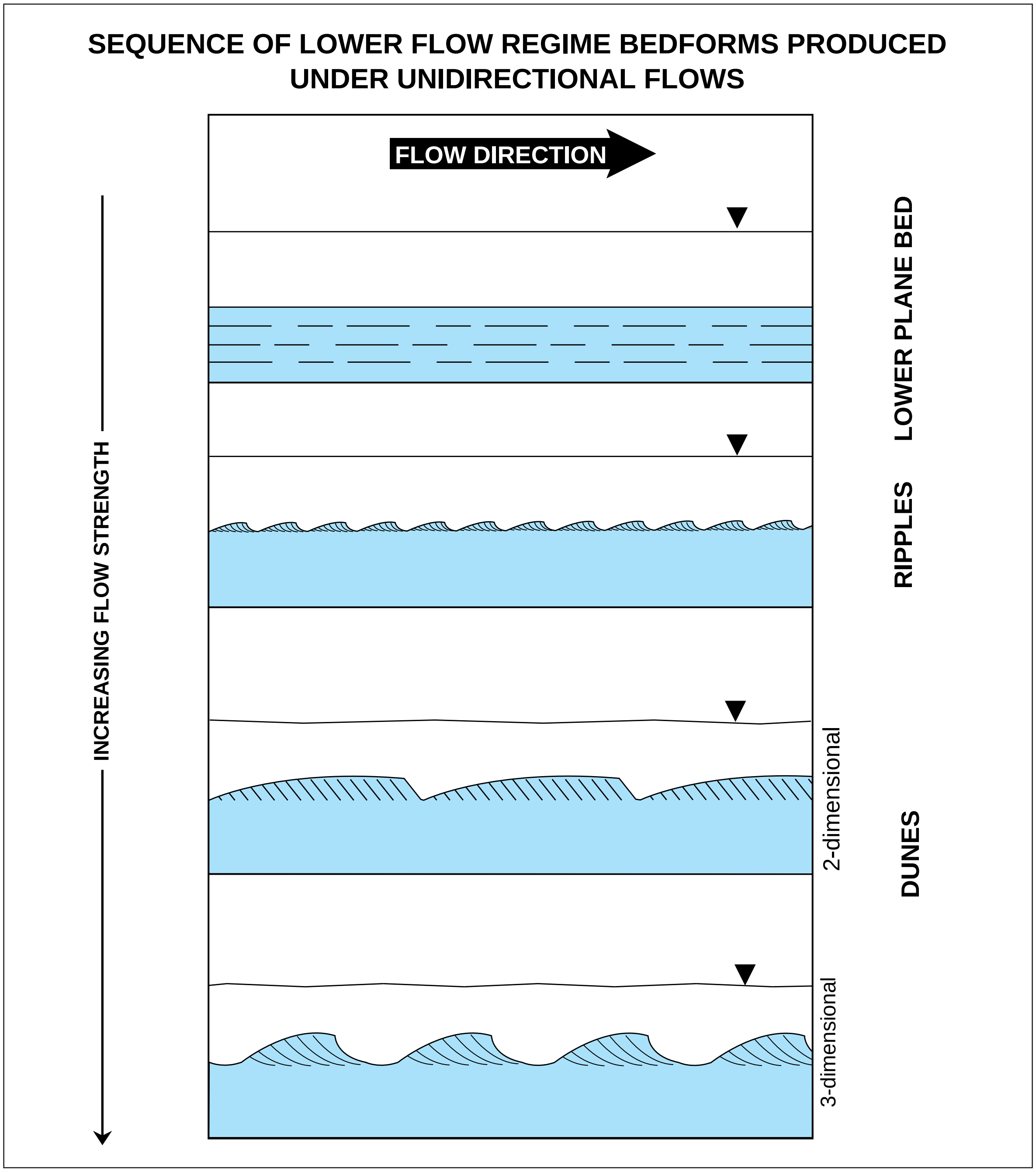


Figure 5-2A. The sequence of bedforms that develop under lower flow regime conditions (after Simons and Richardson, 1961). Note that washed-out dunes (Fig. 2B) are also lower flow regime bedforms.

Ripples range in length from approximately 0.05 m to about 0.6 m and in height from 0.005 m to just less than 0.05 m. The size of a ripple is independent of flow depth but there is a crude correlation between ripple length and grains size; length (L) increases with grain size (d) such that $L \cong 1000d$. The lee slope angle of ripples is close to angle of repose for the sediment, in the range of 25 to 30°. In plan form ripples are highly variable and a terminology has developed to describe these forms (see Fig. 5-5). The first ripples that form when sediment is moved over a bed are straight-crested (2-dimensional) but these do not appear to be a stable bedform because they rather quickly evolve into 3-dimensional forms (with irregular crests). Some experimental evidence suggests that the forms shown

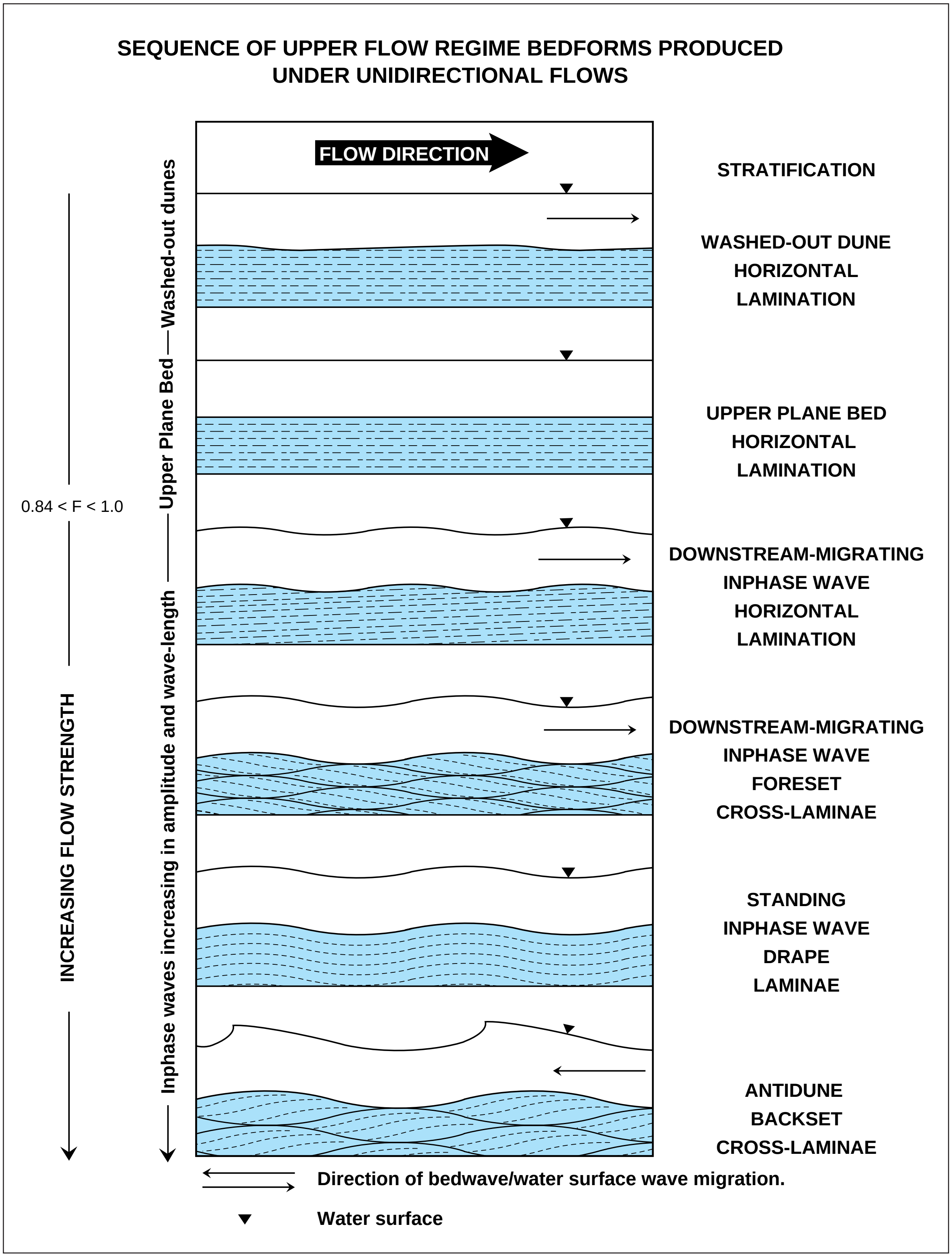


Figure 5-2B. The sequence of bedforms that develop under upper flow regime conditions (after Cheel 1990a). Note that washed-out dunes are actually lower flow regime bedforms.

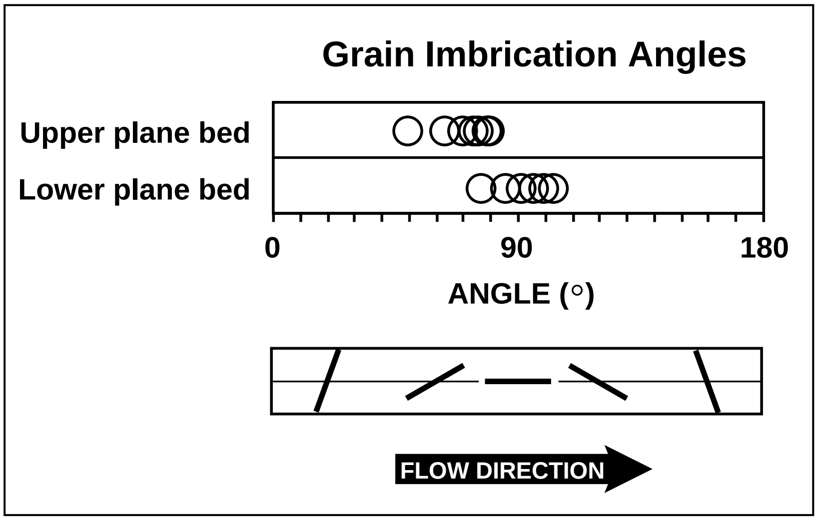


Figure 5-3. Comparison of vector mean imbrication angles, from the plane normal to the depositional surface and parallel to flow, of lower plane bed and upper plane bed deposits. After Gupta et al. (1987).

in sequence in figure 5-5 develop with increasing flow strength. Note that ripples form on beds of sediment finer than 0.7 mm, suggesting that they are typical of dynamically smooth turbulent boundaries. In fact, ripple development seems to require the presence of a well-developed viscous sub-layer. Irregularities on an otherwise planar bed over which sediment is transported may protrude through the viscous sub-layer and cause flow separation, much like that over fully developed ripples (see Fig. 5-4). Once established, the flow character associated with separation and attachment will cause erosion just downstream of the irregularity and deposition even further downstream. Thus, an irregularity on the bed that is high enough to protrude through the viscous sublayer will set up a pattern of erosion and deposition that will propagate laterally and downstream, causing ripples to form.

Dunes are the next bedform to develop with increasing flow strength beyond the upper limit of ripples. They are similar in form to ripples (i.e., asymmetric bedwaves) but are larger than ripples with lengths ranging from greater than 0.75 m to in excess of 100 m and heights ranging from greater than 0.075 m to in excess of 5 m. Dunes tend to be most common on sand beds with a mean size in excess of 0.15 mm. Over the past decade several books and articles have considered dunes to be just large ripples (and they have been termed “large ripples” or “megaripples” in the literature). However, dunes appear to be distinctly different bedforms and there is not a clear continuum in size from ripples to dunes. For example, when we plot bedform length against height of ripples and dunes we see that they are related in the same linear fashion but there is a distinct break between the fields defined by ripples and dunes (Fig. 5-6). This break indicates that asymmetric bedforms with lengths and height over the range of the

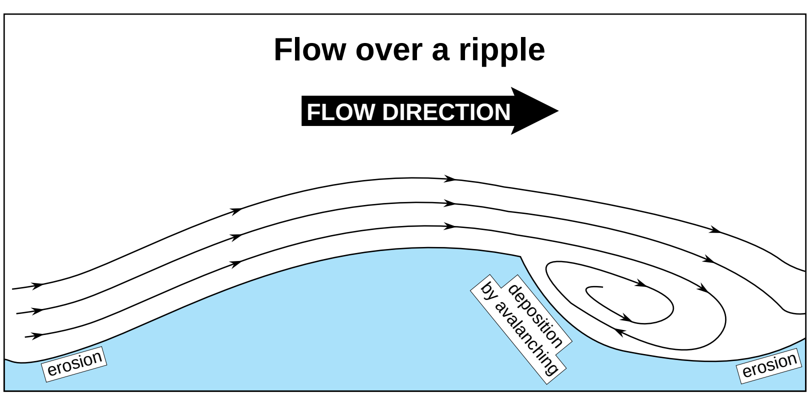


Figure 5-4. Flow separation over the negative step on a boundary due to a ripple.

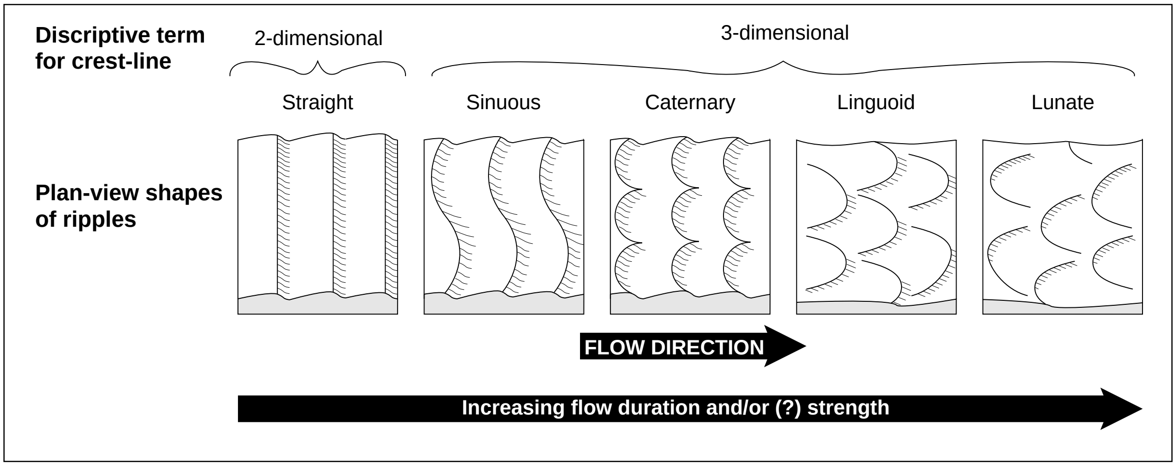


Figure 5-5. Terms used to describe plan forms of asymmetrical bedforms (ripples and dunes). After Allen, 1968 and Blatt, Middleton and Murray, 1980.

break are largely absent in nature. Thus, dunes appear to be distinctly different bedforms than ripples. In fact, under certain hydraulic conditions dunes may have ripples superimposed on their stoss sides (Fig. 5-7). In the lee of dunes large rotating eddies develop (due to the negative step on the bed) and when the upstream velocity of the current generated by the eddy is large it will actually cause the formation of upstream-migrating ripples on the lee slope of the dune (termed "regressive" ripples; Fig. 5-7). Unlike ripples, dune size seems to be **not** related to the grains size of the bed material but **is** related to the flow depth (i.e., mean dune length and mean height increase with mean flow depth). This relationship between dune size and flow depth suggests that the bedforms result from some interaction between large eddies in the flow and the sediment bed. Furthermore, dunes appear to interact with the water surface by producing a "boil" or bulge on the water surface just downstream of the trough due to the ejection of eddies from the trough outwards towards the water surface (it has been suggested that these boils are a form of bursting associated with dunes).

Dunes have been among the most studied bedforms and this has led to considerable confusion of the terminology applied to dunes and to their descriptive characteristics. Finally, in 1989 a symposium was held to arrive at a consensus on many fundamental concerns, beginning with the name to give to these large bedforms (they agreed to call them dunes). Table 5-2 summarizes the main conclusions of this symposium and outlines the descriptive characteristics of dunes that are thought to be important (and these are listed in their order of importance). This

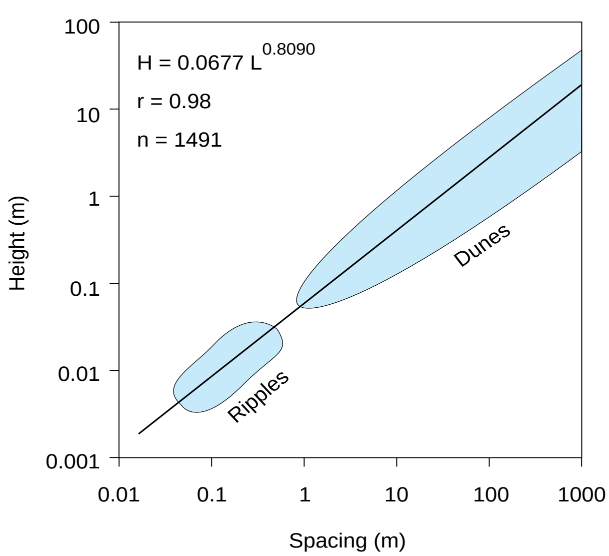


Figure 5-6. Schematic illustration of the field (shaded areas) of ripples and dunes as defined in terms of the lengths (L) and heights (H) of these bedforms. Solid line is based on a linear regression applied to measurements of 1491 bedforms. After Ashley, 1990.

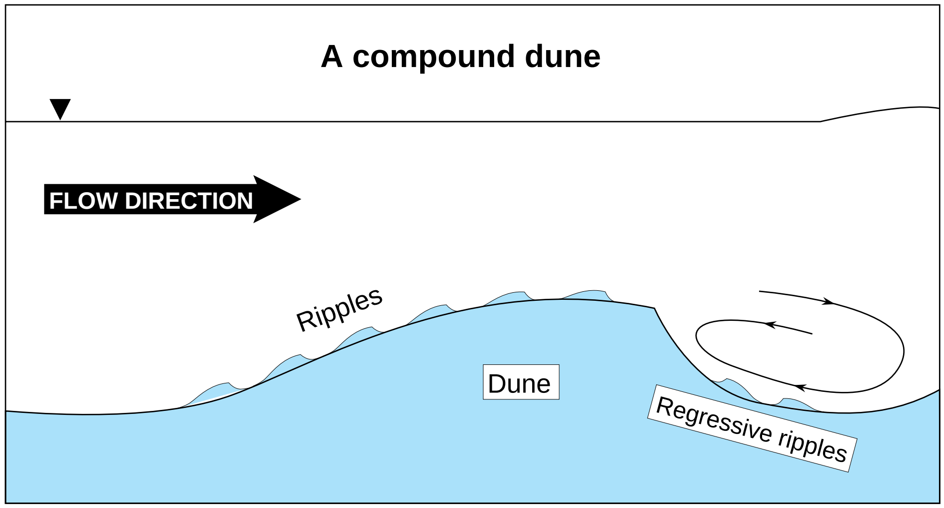


Figure 5-7. Schematic illustration of a dune with ripples migrating up its stoss side (a compound dune). Note that "regressive" ripples (i.e., upstream-migrating ripples) are also shown in the trough and basal lee slope of the dune.

table provides a basis for a consistent terminology regarding dune size and indicates that their are two fundamentally different types of dunes that can be distinguished in terms of their shape: 2-dimensional dunes and 3-dimensional dunes.

2-dimensional and 3-dimensional dunes are so different that many workers have suggested that they are fundamentally different bedforms. However, we now believe that they are part of a continuum of dune forms over the range of flow conditions over which they are stable (the two types of dunes do not form separate fields in plots such as figure 5-6). None-the-less, these two types of dunes differ in many ways and produce distinctive styles of cross-stratification (Fig. 5-2A shows some of the subtle differences between these two types of dunes). 2-dimensional dunes develop under somewhat lower flow strengths than 3-dimensional dunes and they tend to be relatively long and low and straight-crested. The length of the crest is relatively long, in terms of the distance from the summit to the brink. The lee slopes of 2-dimensional dunes are approximately at angle of repose (25-30°) and scour in the troughs is not well-developed. The lee face forms a rather straight, planar surface. Sediment transport may be dominated by contact load and avalanching down the lee slope. Under higher flow velocities 2-dimensional dunes are replaced by 3-dimensional dunes which range from sinuous to lunate in plan form, are shorter and higher, overall, than 2-dimensional dunes and have shorter crests. They have lee slopes that are less than angle of repose and their basal lee faces form curved surfaces extending down into deeply scoured troughs. Sediment transport over 3-dimensional dunes includes contact, saltation and intermittent suspension loads.

With increasing flow strength the 3-dimensional dunes become much longer and lower and their lee slopes undergo a reduction in angle. Such dunes are commonly termed "washed-out" dunes because the strong current seems to wash out the dune form. The heights of such dunes become progressively smaller and their lengths become longer as the flow strength increases and they may form very subtle bedforms as long as several metres with heights on the order of a few millimetres (and may not be considered as dunes as described above). They appear to form a very gradual transition with the next bed form with increasing flow strength: the **upper plane bed**. Note that washed out dunes are shown in figure 5-2B. The reasons for relating this type of dune to plane beds and in-phase waves will become apparent in the section on stratification.

The development of upper plane bed marks the onset of upper flow regime conditions in the scheme of Simons and Richardson (1961). However, this is the one upper flow regime bedform that doesn't really fit into this scheme. While all other upper flow regime bedforms require a free-water surface, upper plane bed does not (unlike in-phase waves, upper plane bed will form in conduits that are completely full, i.e., have no free water surface). In addition, upper plane beds can develop at Froude numbers significantly less than 0.84 (i.e., as low as F = 0.4 under deep flows

Table 5-2. Classification and descriptive characteristics of dunes.

Classification scheme for large-scale, flow transverse bedforms (excluding in-phase waves) recommended by the SEPM Bedforms and Bedding Structures Research Symposium (Austin, Texas, 1987). After Ashley, *et al*. 1990.

General class: **Subaqueous dune**. The modifier "subaqueous" should only be used when a clear distinction between eolian and subaqueous dunes in necessary.

First order descriptors (necessary)

Size:

Spacing	0.6-5m	5-10m	10-100m	>100m
Height[1]	0.075-0.4m	0.4-0.75m	0.75-5m	>5m
Term	**small**	**medium**	**large**	**very large**

[1]Based on the relationship: $H = 0.0677L^{0.8098}$ where H is height, L is spacing

Shape:

2-dimensional. Straight-crested, little or no scour in trough.

3-dimensional. Sinuous to short-crested, deep scour in trough.

Note: no quantitative expression of shape has been agreed upon.

Second order descriptors (important)

Superposition:

Simple. No bedforms superimposed.

Compound. Smaller bedforms superimposed (note size and relative orientation).

Sediment characteristics:

Size

Sorting

Third order descriptors (useful)

Bedform profile: note stoss and lee slope length and angles.

Full-beddedness: fraction of bed covered by bedforms.

Flow structure: time velocity characteristics (e.g., steady flows, tidal flows, etc.)

Relative strengths of operating flows: (e.g., tidal asymmetry)

Dune behaviour-migration history: vertical and horizontal accretion of bed with migration.

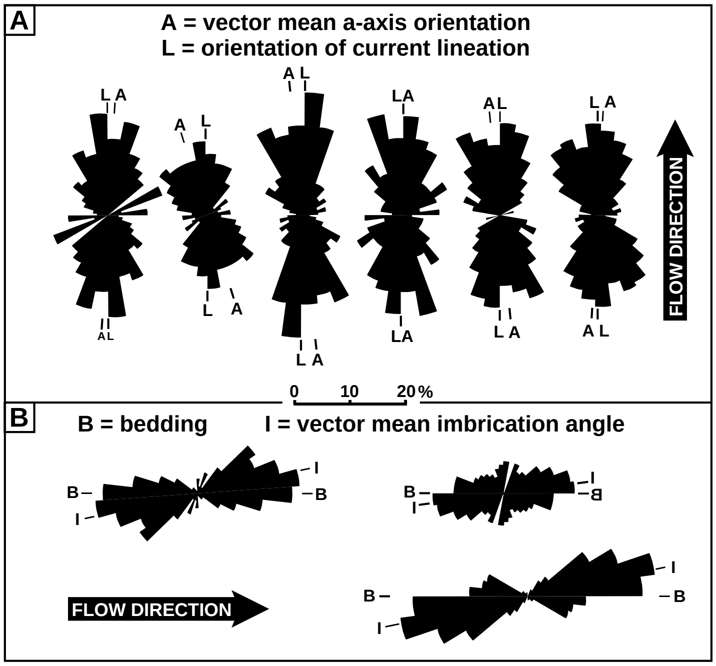


Figure 5-8. Long axes orientations of sand grains deposited on an upper plane bed. A. Long axes orientations, measured on a bedding plane, in comparison to the trend of flow-parallel current lineation. B. Apparent long axes, measured in the plane parallel to flow and perpendicular to bedding. After Allen (1968).

over very fine sand). Upper plane bed seems to have been placed among the upper flow regime bedforms as a matter of convenience to distinguish this bedform from the lower plane bed. Upper plane bed is most common on beds of fine sand and under deep flows and is absent on beds of coarse sand and/or under shallow flows.

Upper plane bed (short for "upper flow regime plane bed"; synonyms include "upper stage plane bed") is widely defined as a flat, planar bed configuration with no significant relief beyond a few grain diameters. The one, widely acknowledged form of regular relief consists of flow-parallel ridges, a few grain diameters high, that are termed **current lineation** that are thought to form due to streaks on the boundary (Weedman and Slingerland, 1985). Current lineations are commonly visible on bedding planes within upper plane bed deposits (and also on the planar stoss slopes of some dunes). In addition to streaks, the bursting process is thought by some to be particularly important to forming horizontal lamination under upper plane bed conditions (more on this below). Several workers have recently suggested that a variety of very low relief bedforms are actually present on upper plane beds (see Bridge and Best, 1988, 1990; Paola et al, 1989; Cheel 1990a,b; Best and Bridge, 1992). The question of the existence of a true plane bed, as defined above, remains and is beyond the scope of these notes.

Sediment transport over upper plane beds is intense as contact, saltation and suspension load and near the bed the concentration of sediment is transport is particularly high. So high, in fact, that a "bedload layer" is commonly prominent that appears as a sheet of moving sediment that been termed a "rheologic layer" by some, a "traction carpet" by others. Grain imbrication is generally well-developed and relatively steep (up to 25° from bedding, dipping consistently into the current; see Figs. 5-3 and 5-8B) and grain a-axes are aligned parallel to the flow (Fig. 5-8A). Note that a-axes alignment on bedding surfaces is bimodal, as seen in plan view with respect to the alignment of current lineation and flow direction; i.e., one mode on either side of the trend of the lineation. This well-developed, flow-parallel alignment of grains is responsible for a structure that is seen in consolidated sandstone termed **parting step lineation**: a structure that occurs on bedding plane exposures of upper plane bed deposits that reflects the tendency for sedimentary rocks that were deposited under upper plane bed conditions to break along vertical planes that are parallel to the direction of flow that deposited the sediment. Note that both current and parting step lineation are useful in determining the "sense" of paleoflow direction. However, because they are lineation they are bidirectional and, therefore, cannot be used alone to determine the absolute paleoflow direction. One common method of determining the absolute paleoflow direction from such lineation is to cut thin sections in the plane perpendicular to bedding and parallel to the lineation. The upstream imbrication direction, seen on that plane, can then be used to infer the absolute flow direction. An easier method of determining absolute

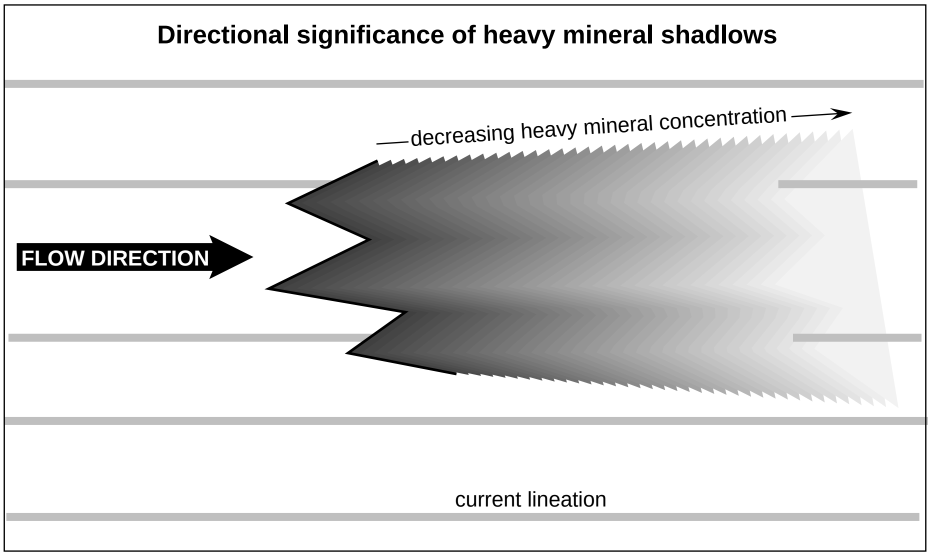


Figure 5-9. Heavy mineral shadows on a current-lineated bedding surfaces on an upper plane bed. Note that the absolute flow direction can be determined as being parallel to the lineation, in the direction of decreasing heavy mineral concentration through the shadow. After Cheel (1984).

paleoflow direction is possible if heavy minerals are abundant on bedding plane exposures of upper plane bed deposits. On upper plane beds heavy minerals commonly form patches on the active bed that display a downstream-decreasing gradient in the concentration of heavy minerals. Such heavy mineral patches are termed **heavy mineral shadows** and may be used, in conjunction with parting and/or current lineation, to determine the absolute paleoflow direction from upper plane bed deposits (Fig. 5-9).

With increasing flow strength beyond the true plane bed very low, asymmetrical bedwaves develop. These bedforms reflect the gradual on-set of the formation of in-phase waves that takes place over a range of Froude numbers from 0.84 to 1.0. Note that on beds of sand coarser than approximately 0.3 mm the transition from dunes to inphase waves may not include a true upper plane bed. Instead, washed-out dunes may pass directly, but gradually, into in-phase waves, with increasing flow strength, and their may be a stage where dune-like forms and in-phase waves co-exist (and this has been reported by Southard and Boguchwal, 1990).

In-phase waves are a class of bedform that are all characterized by their symmetrical form and in-phase relationship with the water surface (although not all in-phase waves are always symmetrical and/or in-phase with the water surface all of the time); hence, in-phase waves are generally described in terms of the water surface wave and the corresponding bedwave and in the description that follows when the bed and water surface wave are behaving in the same fashion these two elements will not be separated. Unlike the other bedforms described above, in-phase waves derive their form and behaviour from the interaction of the water surface and the mobile sediment bed under supercritical (shooting) flows. Cheel (1990a) reviewed literature that indicated that with increasing flow strength, over the range of conditions for in-phase waves, a sequence of different forms of in-phase wave developed (see Fig. 5-2B). Starting with upper plane bed, the first form of in-phase wave is a very low (millimetres high; see Fig. 5-10), downstream-migrating form that increases in height and length with increasing flow velocity. Note that all in-phase waves scale with flow velocity by the relationship $U^2 = \frac{gL}{2\pi}$ (Kennedy, 1963). As flow strength continues to increase the symmetrical bedwaves cease to move downstream and remain stationary on the bed. This type of in-phase wave is termed a "standing" or "stationary" wave. Finally, with a further increase in flow strength the in-phase waves are characterized by upstream migrating forms that Gilbert (1914) first termed "antidunes" (for their upstream migration in contrast to the downstream migration of dunes). Note that all of the forms of in-phase wave may not form on beds of a given sand size or flow depth. Further experimental work is required to establish the fields of hydraulic stability of this class of bedform.

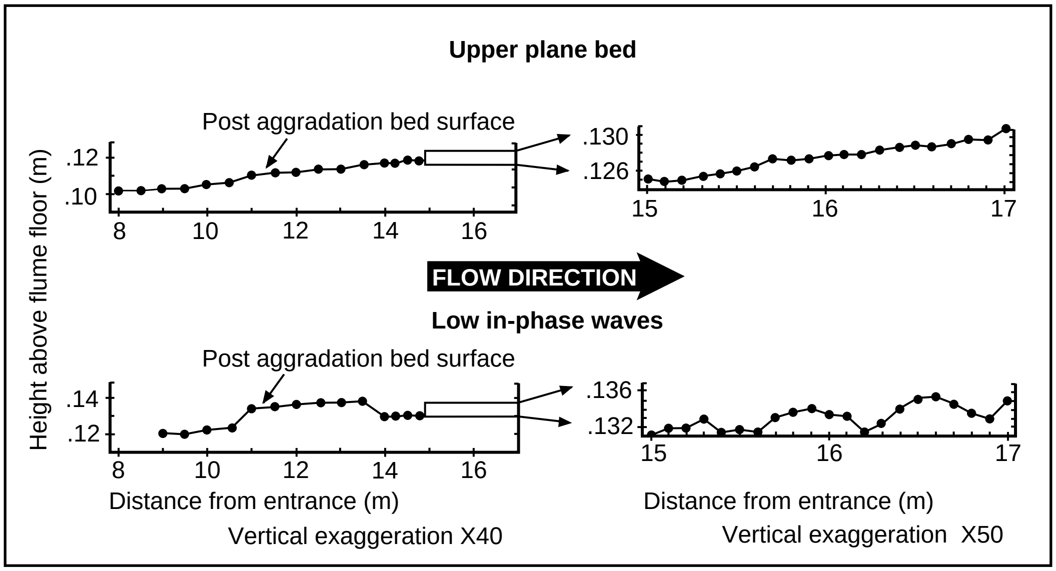


Figure 5-10. Profiles of the sand bed on a flume of upper plane bed and very low in-phase waves. Note the vertical scale. Note that "entrance" refers to the entrance to the flume channel. From Cheel (1990a).

The term antidune is one of the most abused terms used for any bedforms; in many texts it has come to apply to all forms of in-phase waves, a practice that has limited the understanding of these bedforms for almost 30 years. Bed behaviour associated with the antidunes of Gilbert (1914) is very complex; the actual bed state at any one time may be plane bed, downstream-migrating in-phase waves, stationary waves or true antidunes (i.e., upstream migrating). The change in state of the bed associated with antidunes is cyclic, and figure 5-11 shows the sequence of stages of water surface and bed behaviour through a complete cycle of events associated with true antidunes. Note that under flow conditions that produce antidunes all of the stages shown in figure 5-11 may not occur during any particular cycle. For example, the bed and water surface may develop to the stage where stationary in-phase waves form and then the waves may subside back to a plane bed. In addition, all waves on the water surface may not break simultaneously, as shown. In most cases one wave will break and it will disrupt the flow to cause other waves to break subsequently. All of the forms of in-phase waves shown in figure 2B are 2-dimensional (i.e., straight-crested). At flow strengths greater than those required to form 2-dimensional antidunes there are a variety of 3-dimensional in-phase waves (short-crested) of which we know virtually nothing.

Bedform stability fields

In the above description of bedforms we developed some vague idea that certain factors will control which bedforms will develop, depending on flow strength and/or grain size. In fact, many related factors will govern which bedforms will develop on a mobile sediment bed. These factors include those related to the fluid: flow velocity (U), flow depth (D), flow temperature (which controls fluid viscosity and density), and those related to the sediment: grain size (d), grain density (ρ_s), sediment sorting, and particle shape. The effects of particle shape and size sorting on bedforms are not well known and will not be considered here. However, largely because of experimental flume research, carried out by John Southard and his students at the Massachusetts Institute of Technology, we now have a good picture of the fields of hydraulic stability of bedforms on sand beds representing a wide range of grain sizes. Figures 5-12 and 5-13 are "bedform stability" diagrams and show the range of conditions over which the bedforms are stable on diagrams plotting flow velocity versus grain size (each diagram for a given range of flow depths) and flow depth versus flow velocity (each diagram for a given range of grain sizes). Note that the variables on each diagram are scaled to represent conditions where the flow temperature is 10°C. Thus, they are limited to flows of that temperature but they are fairly representative of most natural flows.

All of these figures show that with increasing flow velocity the exact sequence of bedforms that will develop will vary with flow depth and the grain size of the bed material. For example, figure 5-12 shows that with increasing

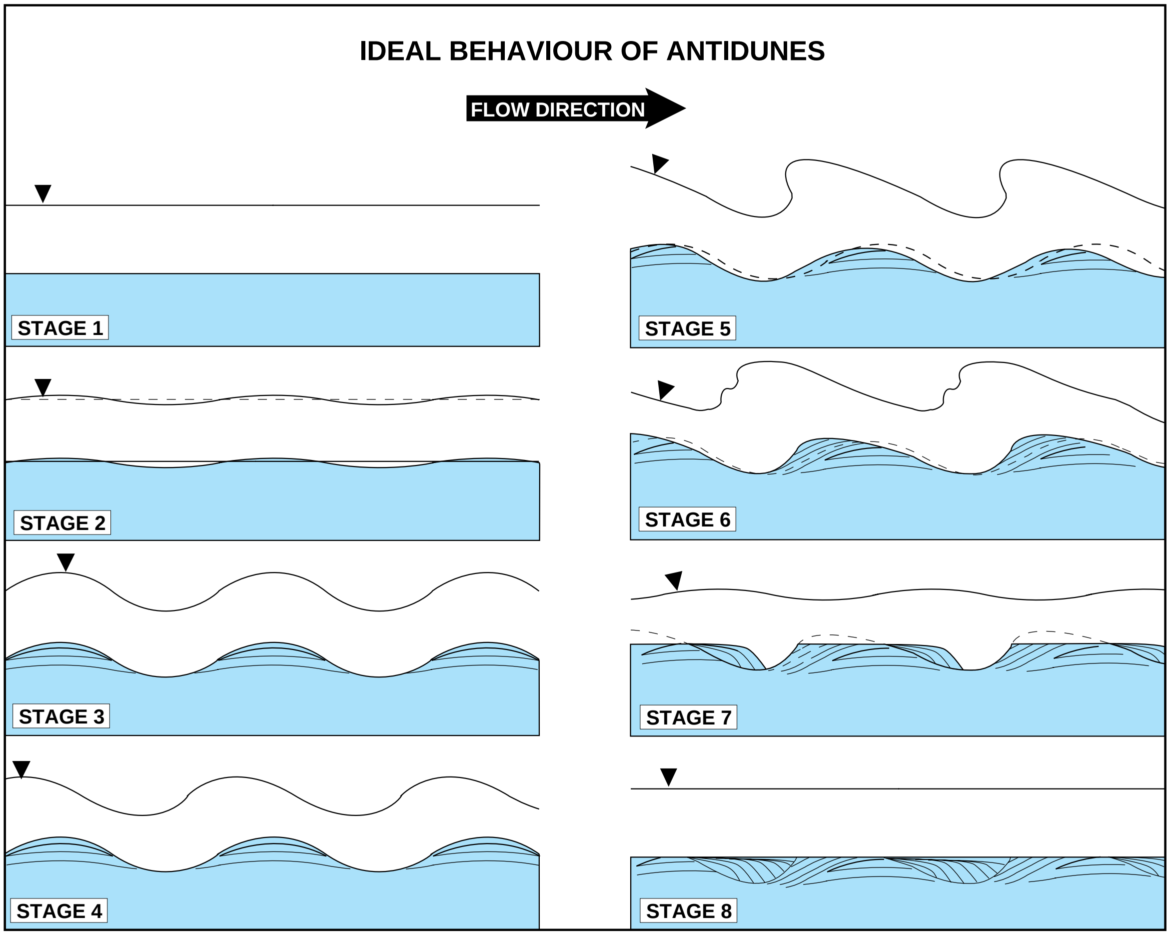


Figure 5-11. A highly schematic illustration showing a complete cycle of the stages of bed and water surface behaviour associated with true antidunes. Note that dashed lines indicate the previous position of bed and/or water surface waves in each stage. The development of cross-stratification is also shown (solid lines within the bed). After Udri (1991). **Stage 1**: under conditions favouring antidune development the bed may remain essentially flat for part of the time. Horizontal lamination may form during this initial stage. **Stages 2 & 3**: stationary in-phase waves develop as sinusoidal water surface waves grow in place and form a sinusoidal bed wave, of lower amplitude. In this stage the bed is molded by erosion under high velocity flow under water surface wave troughs and deposition under the relatively lower flow velocities under the water surface wave crests. In-phase wave drape laminae may develop during this period of *in situ* growth of the bedforms. **Stage 4**: after the water surface wave reaches some critical height and steepness it begins to slowly migrate in the upstream direction. **Stage 5**: the bedwave slowly responds by similarly migrating upstream; the bed and water surface are slightly out-of-phase during this stage. Low-angle backset bedding develops during this stage. **Stage 6**: as the water surface wave continues to migrate upstream and become steeper the bedwave develops what appears to be an asymmetrical bedform on its upstream side; growth of this bedwave is particularly rapid as the water surface wave begins to break by collapsing in the upstream direction. Breaking of the water surface wave results in upstream sediment transport and large quantities of sediment are taken into suspension. Relatively steep (>15°) backset bedding may develop over this stage. **Stage 7**: following collapse of the water surface wave the water surface becomes flatter and the bedwaves are planed off by the very rapid flow. This bed-planing stage involves erosion from the wave crests and deposition in the troughs in the form of a fast-moving, asymmetrical bedform that migrates downstream. Relatively high angle down-stream-dipping cross-strata may develop as this bedform migrates across the trough of pre-existing in-phase wave. **Stage 8**: the water and bed surfaces are planar and the cycle may begin again with deposition of horizontally laminated sand, truncating the underlying cross-stratification produced by the in-phase waves.

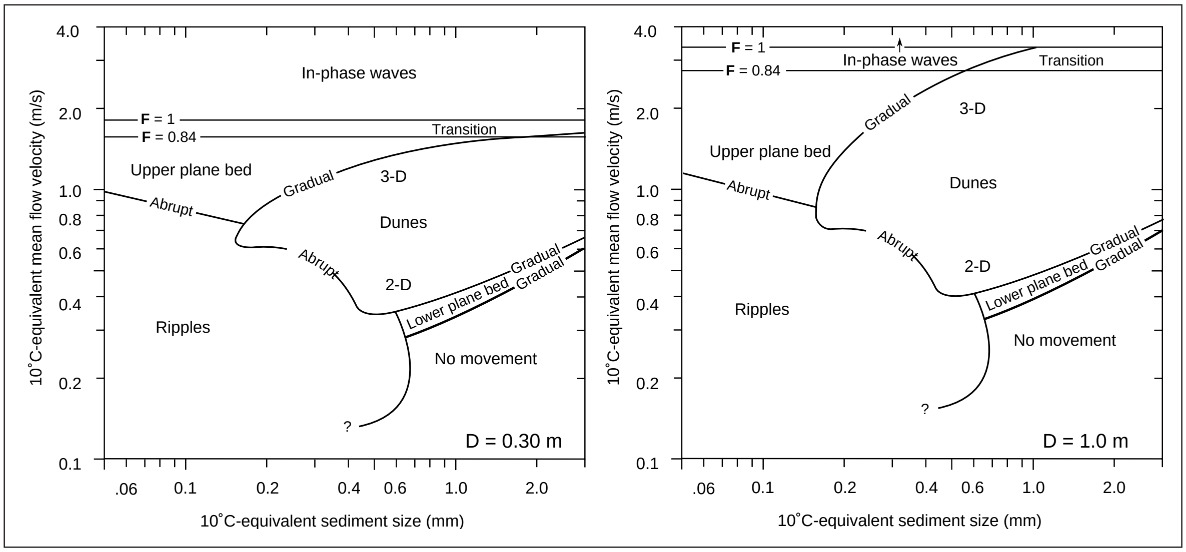
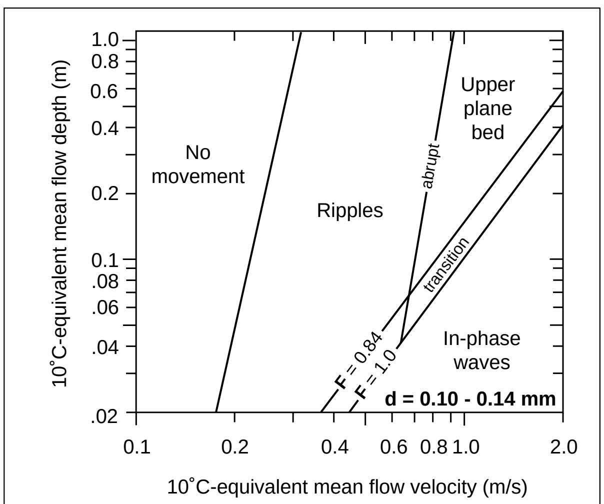


Figure 5-12. Velocity/grain size diagrams showing the fields of bedforms stability for two ranges of flow depth (all variables are scaled to 10°C water temperature). See text for discussion. After Southard and Boguchwal (1990).



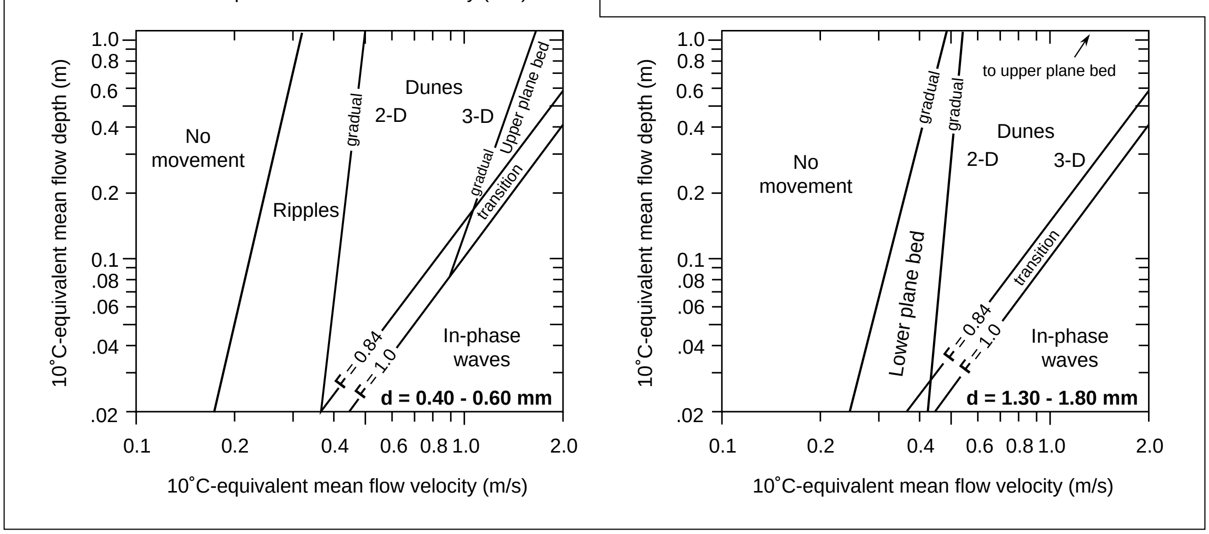


Figure 5-13. Depth/velocity diagrams showing the fields of bedforms stability for three ranges of grain size (all variables are scaled to 10°C water temperature). See text for discussion. After Southard and Boguchwal (1990).

velocity, flow over a bed of 0.1 mm sand will produce the sequence ripples → upper plane bed → in-phase waves, whereas the same flow over a bed of 1 mm sand will produce the sequence lower plane bed → dunes (2-D followed by 3-D) → in-phases waves (with upper plane bed following dunes under deep enough flows). Lower plane bed and dunes are limited to relatively coarse grain sizes whereas ripples and upper plane beds dominate fine grain sizes (particularly for the narrow range of flow depths represented by the figures: 1 to 3 m). Note that the maximum velocity at which dunes and plane beds are stable increases with flow depth and is governed by the flow Froude number. Also, note that the transitions form ripples to dunes and from ripples to plane bed are abrupt whereas all other transitions form one bedform to another are gradual. The effect of flow depth on the sequence of bedforms that develop with increasing flow velocity is apparent in figure 5-13. For example, the depth/velocity diagram for 0.4 mm to 0.60 mm sand shows that at flow depths less than 0.1 m the sequence of bedforms that develops with increasing flow velocity is ripples → dunes (2-D followed by 3-D) → in-phase waves. However, at depths above 0.1 m upper plane bed becomes stable and the sequence of bedforms becomes: ripples → dunes (2-D followed by 3-D) → upper plane bed. → in-phase waves. Figure 5-13 also shows that the range of flow velocities over which ripples are stable decreases with grain size, presumably because the viscous sublayer is destroyed under increasingly lower velocities as the boundary roughness increases. Dunes and upper plane bed are both stable over wider ranges of velocities as flow depth increases (again, because their upper limit is governed by the flow Froude number). In contrast, lower plane bed exists over an increasingly narrower range of velocities as depth increases. Note that the in-phase waves are not subdivided in diagrams like these because most experimental studies to date do not extend through the upper flow regime.

Figure 5-14 shows how the height and lengths of dunes varies as a function of flow depth and velocity and grain size, within the dune stability field. The effect of flow depth on dune size (that was mentioned earlier) is clearly evident in the depth/velocity diagrams: as flow depth increases dunes become both longer and higher. The velocity/grain size diagrams show that dune length increases with flow velocity but also with decreasing grain size: the longest dunes develop on beds of the finest sand. In contrast, dune height is only loosely determined by grain size (fine sand has a greater likelihood of developing higher dunes). For all sand sizes, with increasing flow velocity, dune height first increases and then decreases towards the upper velocity limit of dune stability. This reflects the washing out of dunes described above.

Figures like those shown here are necessary for the interpretation of paleoflow conditions based on the bedforms that are preserved in ancient sediments. However, they continue to allow only qualitative interpretations regarding the relative flow strength. A major limitation of data like that shown in figures 5-12 to 5-14 is that they are based on relatively shallow flows depths, no more than a couple of metres deep whereas natural flows may range up to several tens of metres, beyond the range represented by the experimental data. As our understanding of the controls on bedform stability improve so will our ability to interpret these structures.

Cross-stratification formed by bedforms under unidirectional flows

Cross-stratification is a type of primary structure that occurs in a wide variety of forms and develops in sediments and sedimentary rocks due to temporal and spatial variation in deposition and erosion on a bed, normally in association with the migration of bedforms. In this section we will see how the form of cross-stratification may be used to infer type bedform that produced it and of the bedform behaviour (which may also be used to infer something of the conditions in the depositional environment). However, first the terminology that has been developed for describing cross-stratification must be introduced.

Terminology

Once again, consistent terminology is required so that sedimentologists can communicate with each other. Agreement on definitions of terms is not always easy (for example a symposium had to be held to come to a consensus on the name “dune” for large, asymmetrical, flow transverse bedforms). However, there is a fairly consistent and simple set of terms to describe layered sediments; the general terms will be defined here and more specific terms will be introduced in the following section.

Figure 5-15 shows a hypothetical sequence of layered sediments or sedimentary rocks in order to illustrate

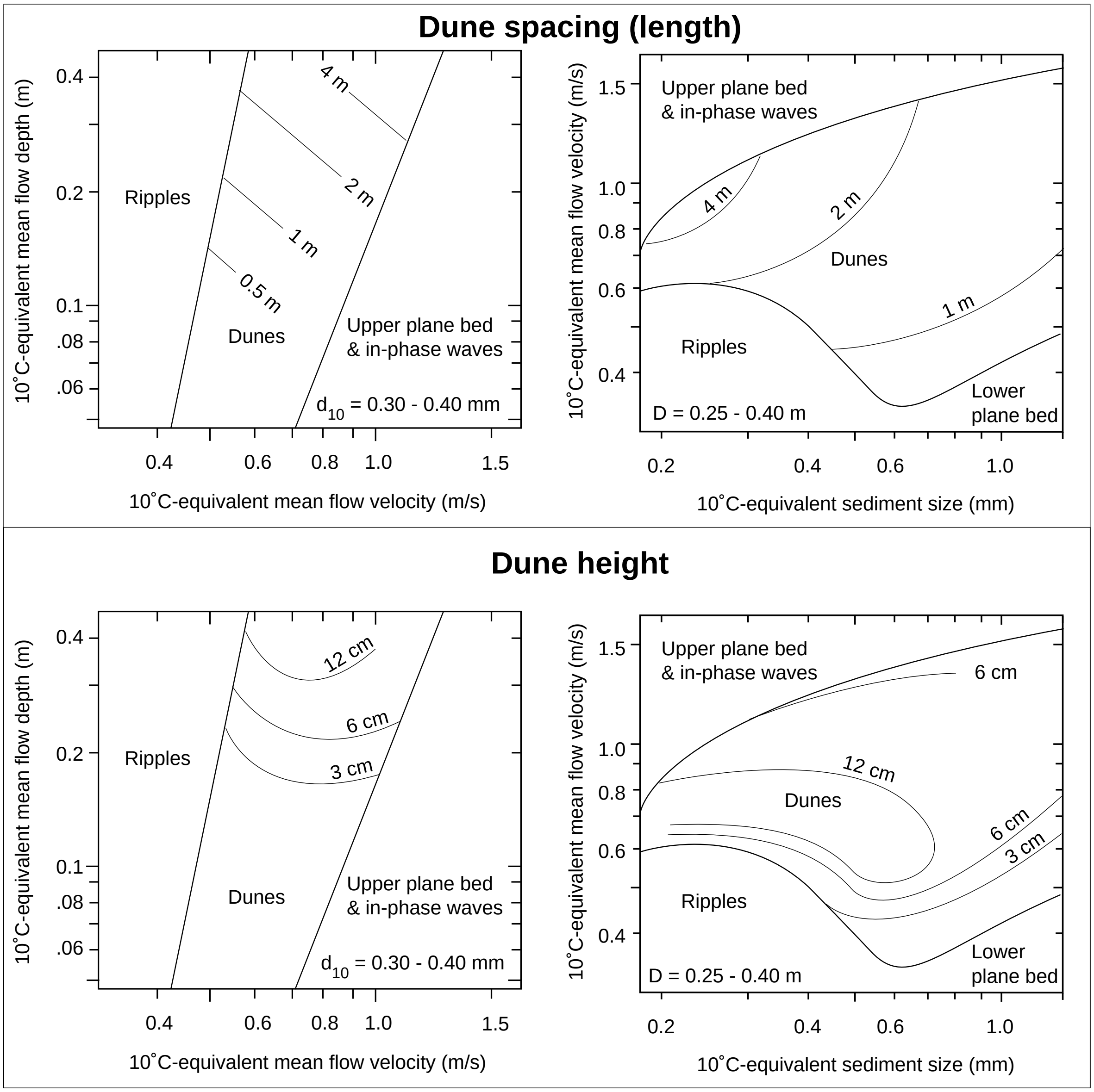


Figure 5-14. Bedform stability diagrams showing variation in dune spacing and dune height as a function of grain size, flow depth and flow velocity (all scaled to 10°C water temperature). See text for discussion. After Southard and Boguchwal (1990).

some of the fundamental aspects of sedimentary layering. The term **stratum** (plural: strata) is a general term applied to any layered rock unit that is clearly distinguishable from units above and below due to some discontinuity in rock type (i.e., composition or texture). Any stratum may be **simple** (i.e., not internally divisible) or **complex** (composed of a number of distinguishable internal units). A **bed** is any stratum that is greater than 1 cm thick and a **lamination** (or lamina, plural is laminae) is any stratum that is less than 1 cm thick. The term **layer** is reserved for a part of a bed which is bounded by some minor, but distinct, discontinuity in texture or composition. The contact between beds is very distinct (e.g., the contact between beds in Fig. 5-15 is reflected by the abrupt vertical change from shale to sandstone). The contact between layers is generally much more subtle and may represent an erosional surface between packages of similar lithology. Such erosional surfaces between layers are termed **amalgamation surfaces**. A **division** is a layer, or part of a layer, that is characterized by a particular association of primary sedimentary structure. A **band** is a laterally *continuous* (on outcrop scale) portion of a layer that is distinguishable on the basis of colour, composition, texture or cementation. A **lens** is a laterally *discontinuous* (on outcrop scale)

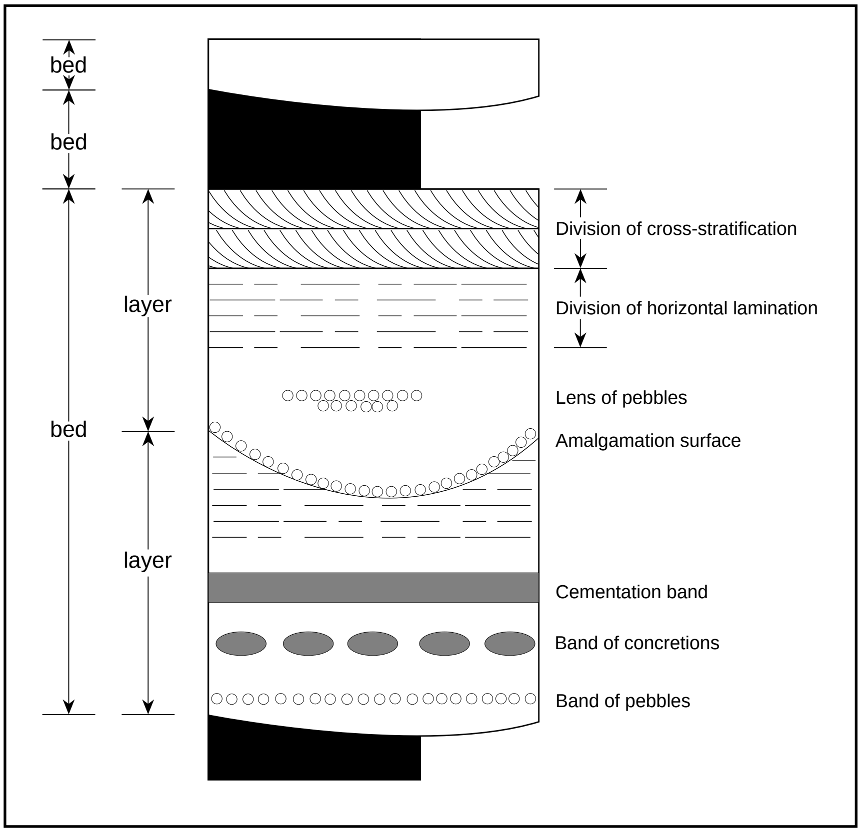


Figure 5-15. Terms used to describe various types of layering in sediments. Note that the lithology that is black is shale and the white lithology is sandstone. After Blatt, Middleton and Murray (1980).

portion of a layer that is distinguishable on the basis of colour, composition, texture or cementation.

All stratification that is inclined due to primary processes (i.e., not tectonic deformation or tilting) is referred to as **cross-stratification**. Cross-stratification typically includes packages of parallel beds or laminae that are bounded by planar surfaces (bedding planes that are termed **bounding surfaces**). The upper bounding surface is normally an erosional surface and truncates underlying internal strata. (Note that this allows us to determine the original direction to top in tectonically deformed sediments.) The strata between bounding surfaces are often termed **internal strata** or **internal stratification**. Internal strata are distinguishable as beds or laminae on the basis of often subtle changes in grain size and/or mineralogical composition. A group of similar internal cross-strata, between bounding surfaces, is referred to as a cross-strata **set**; a group of similar sets of cross-strata is referred to as a **coset**. Any description of cross stratification must include the form of the internal cross-strata (see Fig. 5-16), the thickness of the sets, the direction of dip of internal strata, the form and geometry of the bounding surfaces and the thickness of cosets. Figure 5-17 outlines some of the common forms of cross-stratification and provides terms for describing these structures. Note that the term cross-stratification is a general one; more specific terms are cross-laminae (internal strata are less than 1 cm in thickness) and cross-bedding (internal strata are greater than 1 cm).

Origin of cross-stratification

The occurrence of bedding plane exposure of bedforms is relatively rare in outcrop. We typically can see only the two-dimensional view (in vertical section) of the internal structure produced by bedforms as they move over a bed surface. Because bedforms essentially migrate through each other (see above and further discussion

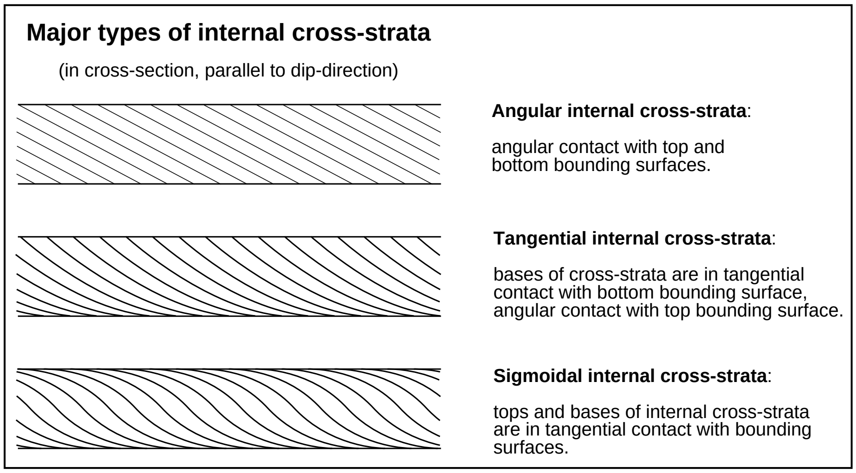


Figure 5-16. Forms of internal cross-strata. Note, parallel horizontal lines are bounding surfaces.

below) we are usually limited to only partial preservation of that internal structure. The internal structure of the bedform that we normally see is some expression of the lee face and trough of the bedform, and only rarely (for anything but ripples) the stoss slope. The expression of these aspects of bedforms form may be visible as layers due to subtle variation in grain size and/or mineralogy (seen as variation in colour) on the surfaces preserved within the deposits. These layers comprise packages of inclined strata (internal strata paralleling depositional surfaces) within the deposits that may be cut by erosional surfaces (bounding surfaces).

The following discussion is directly relevant to the formation of cross-stratification by asymmetric bedforms but also illustrates principles that are important to the formation and preservation of forms of stratification produced by other bedforms. Figure 5-18 shows the hypothetical lateral migration of an inclined surface (like the lee slope of a bedform) due to periodic deposition of sediment on the surface (from times t1 to t_9 forming layers L_1 to L_8). Each

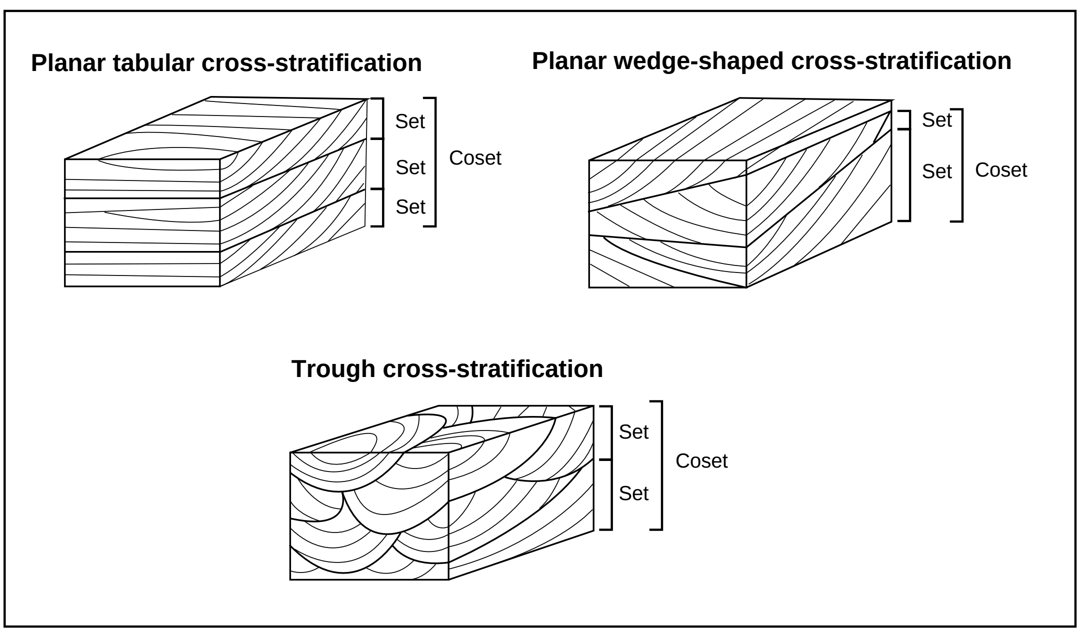


Figure 5-17. Terminology for cross-stratification. Note that planar tabular cross-stratification is characterized by planar, parallel bounding surfaces, wedge-shaped cross-stratification is characterized by planar but not parallel bounding surfaces, and trough cross-stratification is characterized by trough- or scoop-shaped bounding surfaces (also called festoon cross-stratification). After Blatt, Middleton, and Murray, 1980.

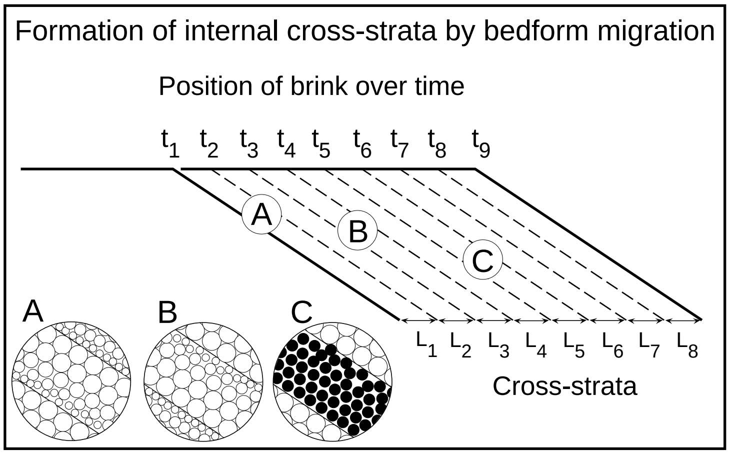


Figure 5-18. Illustration showing the formation of internal stratification with bedform migration by episodic deposition on the lee slope. Insets show internal form of cross-strata formed by avalanching (A; inversely graded), periodic fallout from suspension (B; normally graded) and deposition of heavy mineral-enriched sediment (C). See text for further discussion.

layer, termed a foreset layer or foreset, represents a particular depositional event on the inclined surface. Each individual layer may be visible due to some mineralogical and/or textural differences between layers that might arise from a number of possible sorting mechanisms that act while they are deposited. For example, a particular foreset layer may preserve variations in grain size that developed as the sediment avalanched down the inclined surface; avalanche deposits tend to be inversely graded (i.e., become coarser within cross-strata, in the direction of migration, normal to the lee face; Fig. 5-18A). Avalanching induces shearing within the sliding sediment and the grains exert a "dispersive pressure" (a force per unit area acting in the direction pointing away from the solid surface that arises due to grain collisions) which forces coarse grains upwards relative to fine grains. The largest grains also tend to roll further down the lee slope so that internal strata tend to become finer-grained up slope and sets of such cross-strata are normally graded (i.e., become finer upward) vertically through a cross-strata set. Internal strata formed by avalanching tend to have angular contacts with underlying bounding surfaces (see Fig. 5-16). In contrast to avalanching, sediment that falls periodically and in pulses from suspension tends to be normally graded (i.e., becomes finer, within cross-strata in the direction of migration, normal to the lee face; Fig. 5-18B) because the largest particles tend to reach the depositional surface first, followed by increasingly finer grains (i.e., the grains are sorted according to their settling velocities; internal strata produced by this mechanism tend to be tangential or sigmoidal in form, see Fig. 5-16). Alternatively, the layers that are deposited on the lee surface of the bedform may reflect variation in mineralogy; any particular layer may be formed when a heavy mineral accumulation is swept over the brink of the bedform to deposit as a heavy mineral rich foreset (Fig. 5-18C). In addition, particularly in the case of ripples, micaceous minerals commonly lie more-or-less parallel to the plane of the lee slope and highlight the internal stratification. Note that because internal strata lie on the plane of the lee face of the bedform they dip in the direction of bedform migration and this direction is, on average, also the direction of current flow. Therefore, the dip direction of internal stratification produced by ripples and dunes is a valuable paleocurrent indicator and one that is very easy to measure in the field, in contrast to grain orientation. The majority of paleocurrent data are based on the geometry of cross-stratification.

The above example of deposition on an inclined surface is instructional to visualize how internal stratification develops but it neglects the fact that as bedforms migrate so do the regions of deposition and erosion that they generate by their interaction with the flow. Deposition is largely limited to the lee face of the bedform (as outlined above) and erosion occurs from the trough and continues along the stoss side of the bedform. As the bedform migrates by deposition on the lee face it forces the trough, and its region of scour, further downstream, up the stoss slope of the next-downstream bedform. Thus, as one ripple migrates it consumes the next-downstream ripple, and so on. Figure 5-19A illustrates the condition where there is no net deposition on a bed during ripple migration; as the bedform migrates the volume of sediment deposited on the lee face will equal the volume of sediment removed from its stoss side. Only if there is net deposition on the bed (i.e., more sediment is added to the bedform than is removed with erosion) will all or any part of a bedform survive destruction associated with ripple migration. Figure 5-19B shows the example where there is sufficient deposition on the bed (i.e., the bed undergoes aggradation while the bedform migrates) for preservation of the form of the bedform with downstream migration. In this case, the

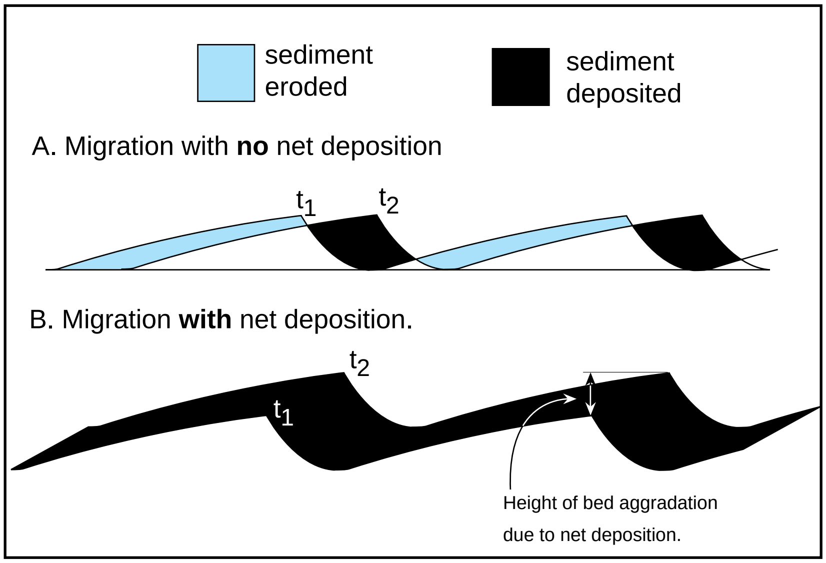


Figure 5-19. Schematic illustration of the pattern of erosion and deposition with ripple migration. A, with no net deposition or erosion on the bed the volume of sediment eroded from the stoss side of the ripple must equal the volume deposited on the lee slope. Thus, with migration by one ripple wavelength, from time t_1 to t_2 all of the sediment contained within ripples at t_1 will have been eroded and deposited within the ripples at t_2. B, with net deposition on the bed no erosion takes place and the entire ripple form is preserved.

volume of sediment added to the bed over the time for migration by one ripple wavelength (for example, sediment deposited by fall-out from suspension) is more than the volume of each bedform so that the complete form of the bedform is preserved. Note that if the volume of sediment added to the bed were less than the volume of the bedform only partial preservation of the bedform is possible.

The extent to which a bedform is preserved depends on the relationship between the rate at which the bed is aggrading and the rate at which the bedform is migrating. Figure 5-20 shows how various proportions of bedforms (and their associated internal stratification) will be preserved as a function of these two variables. Figure 5-20A shows that the path of a migrating bedform on a bed that is aggrading may be defined as a vector that is the sum of the net migration and net bed aggradation over some fixed period of time. If the bed is aggrading the vector describing this path is inclined upward, in the direction of bedform migration, and the bedform is said to "climb" along this path. The angle of climb (β) is determined by the ratio of the aggradation rate to the migration rate. Note that if we consider the migration path with a point of origin in the bedform trough then it delineates a line of erosion as the bedform migrates (or plane of erosion if we consider three dimensions); everything above the migration path is eroded due to scour in the trough and everything below the line of migration is preserved and buried by subsequent deposition. This plane of erosion results in the bounding surfaces that define cross-strata sets and the internal strata are formed directly by deposition on the lee slope of the bedform and preserved between the bounding surfaces. Thus, in figure 5-20B, where there is no bed aggradation, the migration path is simply the path of ripple migration and the line of erosion is horizontal. As in figure 5-19A, all downstream bedforms are planed off by erosion in the trough of the migrating upstream bedform. In this case the only internal stratification that would be preserved would be within any ripple forms that survived after migration (and presumably the current) had stopped. In figure 5-20C bedform migration is accompanied by a moderate rate of bed aggradation compared to the ripple migration rate and the internal deposits of bedforms are partially preserved. However, the path of the migrating ripple trough climbs at an angle that is smaller than the angle of the stoss slope (α) of the bedform. As the ripple migrates the plane of erosion passes beneath the stoss slope of the next downstream bedform and the top portion of the internal strata of the downstream ripple are eroded. Whenever $\alpha > \beta$ only part of the internal strata will be preserved; as a approaches equality with β more and more of the internal stratification will be preserved. Figure 5-20D shows the situation where the aggradation rate is high, relative to the migration rate, such that the angle of climb is larger than the stoss slope angle of the bedform (i.e., $\alpha < \beta$) and all of the internal deposits of the bedforms

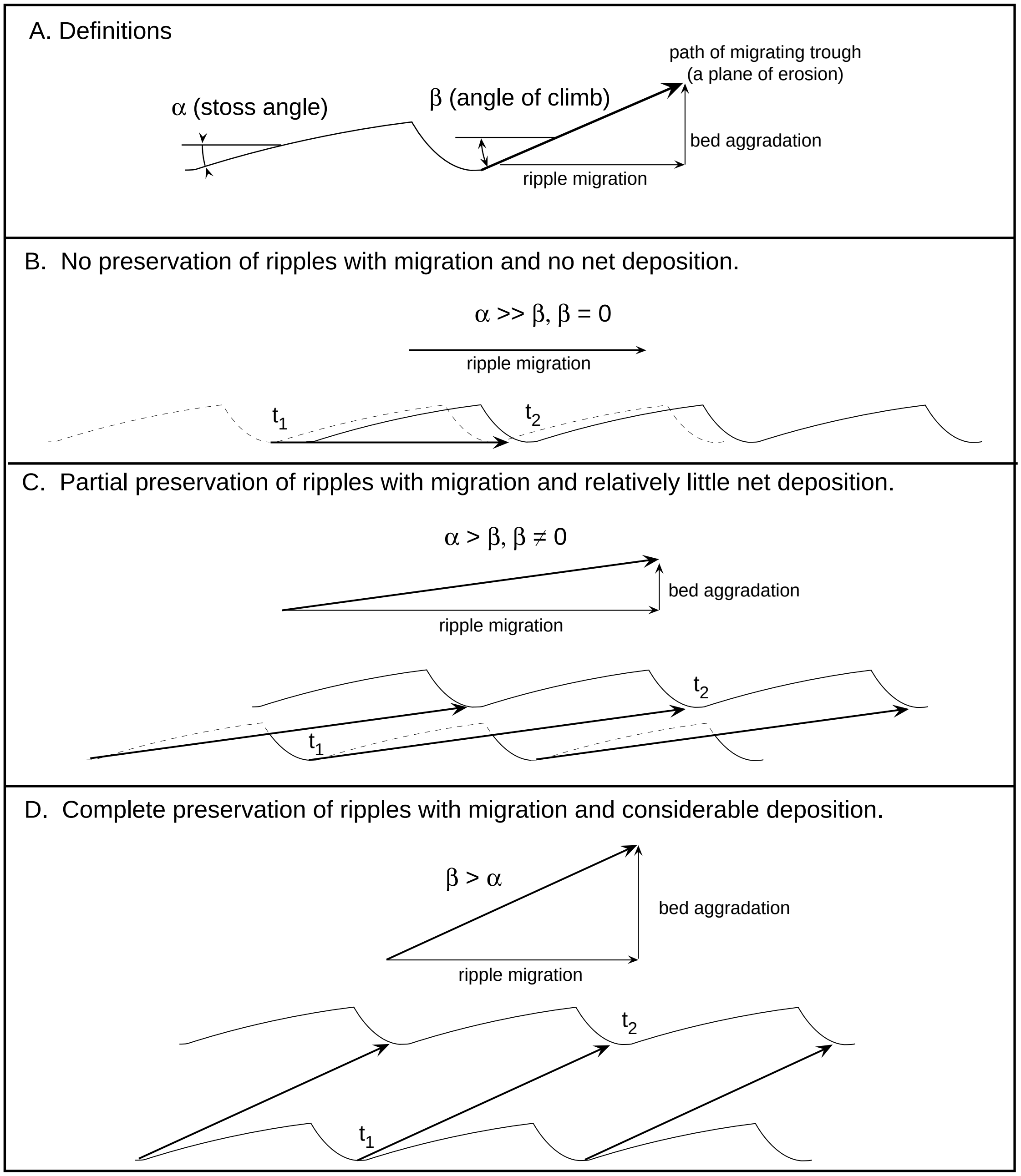


Figure 5-20. Illustration of the effect of the relative aggradation rate and bedform migration rate on the preservation of the internal deposits of asymmetric bedforms. Note that dashed lines indicate the portions of ripples at t_1 that are eroded with migration to positions shown for t_2. See text for discussion.

are preserved as they migrate downstream.

From the above discussion it should be clear that cross-stratification, like that shown in figures 5-16 and 5-17, forms in response to bedform migration and the particular style of cross-stratification that is preserved will depend on: (1) the type of bedform (this controls, among other things the thickness of the cross-strata sets and the thickness of internal strata); (2) the relative rates of bedform migration and bed aggradation; (3) the nature of sedimentation on the bedding surfaces. Figure 5-21 illustrates how the various forms of cross-stratification shown in figure 5-16 might develop. Angular internal strata (Fig. 5-21A) form when deposition on the lee face of the bedform is dominated by avalanching (resulting in the angular basal contact of internal strata with the lower bounding

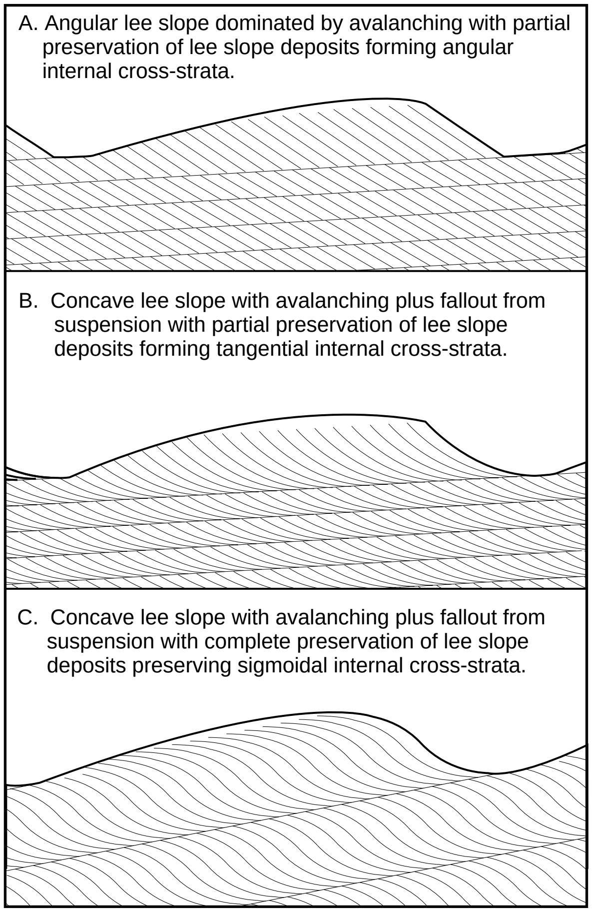


Figure 5-21. Styles of internal stratification shown in terms of their mode of formation and style of preservation (see text for discussion).

surface) and the top portions of internal strata are removed by erosion because the angle of climb of the bedform is less than its stoss slope angle (resulting in an angular contact with the upper bounding surface). Tangential internal strata (Fig. 5-21B) form when deposition on the lee face of the bedform includes an important component of sediment that is falling out of suspension (much of it having gone into suspension after passing over the brink of the bedform). The base of the lee face is gently curved and tangential with the lower bounding surface formed by erosion in the downstream trough. The tops of the internal strata have an angular contact with the upper bounding surface where they are erosionally truncated. Sigmoidal internal strata (Fig. 5-21C) form when bedforms with a concave lee slope climb at relatively high angles such that the erosional bounding surface passes close to the bedform crest. This form of cross-stratification is especially well-developed by bedforms with long crests and relatively short lee faces (e.g., washed out dunes).

Cross-stratification produced by asymmetrical bedforms

There are two major differences in the forms of cross-stratification produced by ripples and dunes. Because of the small size of ripples the internal stratification is typically thin (i.e., they are cross-laminae, < 1 cm in thickness) whereas dunes commonly produce thicker internal strata (commonly cross-bedding, > 1 cm in thickness). Also because of the small size of ripples they tend to form a variety of forms of climbing ripple cross-stratification. Because of the relatively small volume of sediment in a ripple the aggradation rates required to cause these bedforms to climb at relatively steep angles are not difficult to achieve in nature. However, large dunes contain such large volumes of sediment that it is only rare that aggradation rates are large enough to cause the bedforms to climb at angles greater than their stoss slopes. Ripples and dunes bear the common characteristic that their plan form dictates the overall form of cross-stratification. Straight-crested (2-dimensional bedforms) tend to form planar tabular and planar-wedge-shaped cross-stratification (the planes of erosion associated with trough migration are relatively flat). In contrast, 3-dimensional forms produce trough cross-stratification because the planes of erosion are highly irregular,

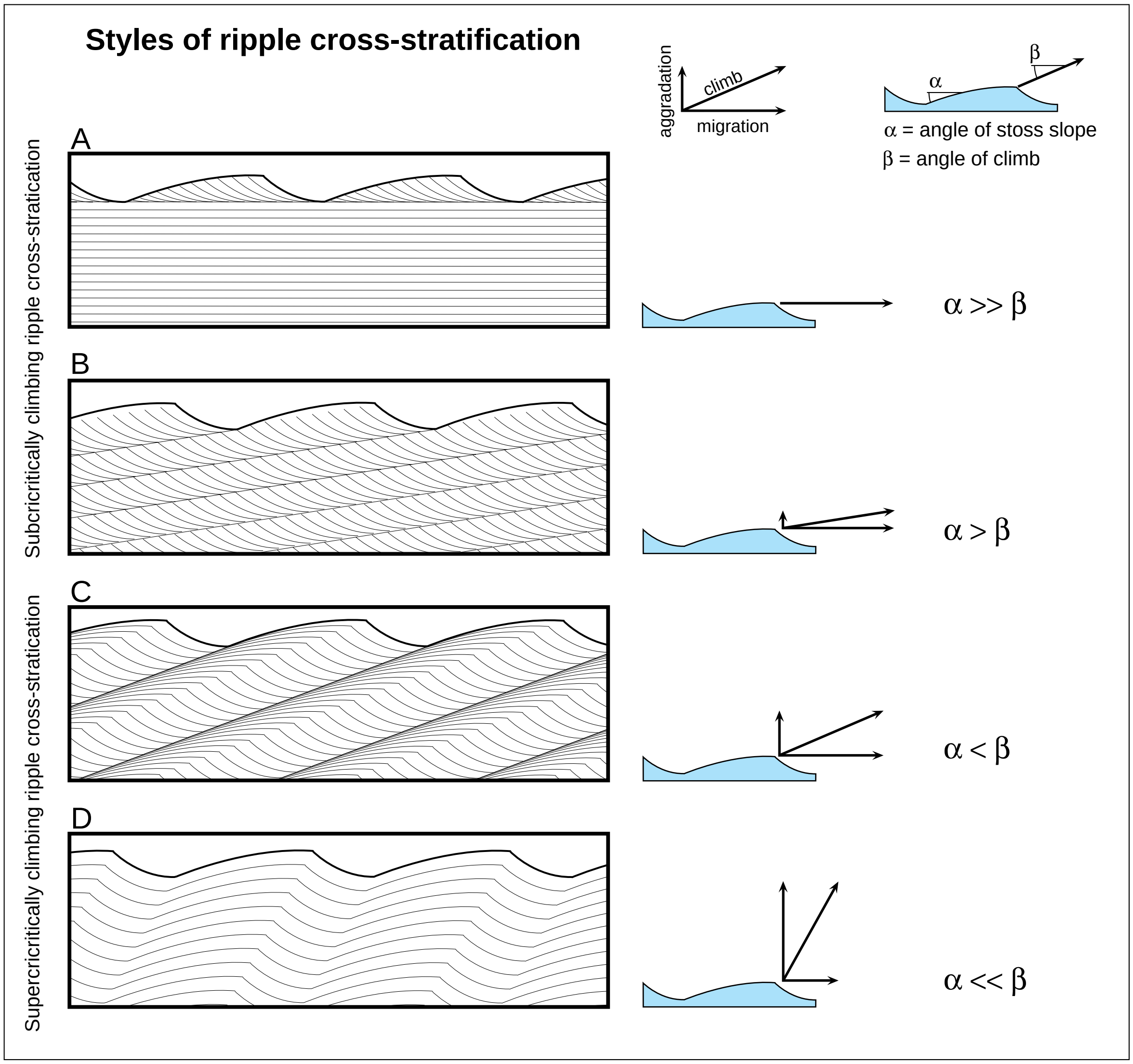


Figure 5-22. Forms of ripple cross-stratification that develop as a function of the angle of climb of the bedforms. See text for discussion of details.

like the troughs and crests of the bedforms, and internal strata are best preserved where the bedforms migrate into a particularly deeply scoured trough.

Ripple cross-lamination

Internal laminae produced by ripples can have a variety of forms: from angular to sigmoidal and planar to trough-shaped in cross-section. The thickness of sets of ripple cross-lamination is limited by the upper limit to ripple height, approximately 0.05 m. Most ripple cross-lamination is classified according to the angle of climb of the bedforms and the resulting forms of internal laminae. Figure 5-22 shows a variety of ripple cross-lamination types. Figure 5-22A is a form of horizontal lamination that is produced when the angle of climb of the ripple is negligible (i.e., much less that the stoss slope of the bedform). Any deposits that are preserved between bounding surfaces as the ripples migrate over the bed surface are too thin to allow the recognition of internal laminae. With increasing angle of climb beyond some critical angle, that will depend on the sorting and mineralogy of the sediment, the deposits between bounding surfaces become thick enough to preserve visible internal lamination (Fig. 5-22B). Because the angle of climb is smaller than the stoss slope angle of the bedform (i.e., $\alpha > \beta$) the foreset laminae are

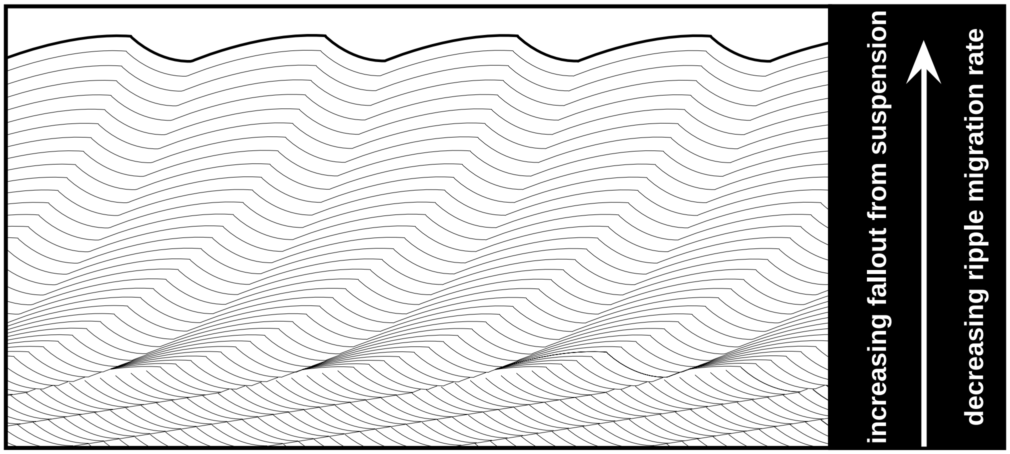


Figure 5-23. A relatively common vertical sequence of climbing ripple cross-stratification that is produced by waning, sediment laden flows. Note that the angle of climb increases continuously upwards as the current wanes, resulting in a temporally increasing rate of fallout from suspension and a decreasing rate of ripple migration.

truncated on top and all of the stoss slope deposits have been eroded. Such ripple cross-stratification is often termed **subcritically climbing ripple cross-stratification**; subcritical means that $\alpha > \beta$. This form of cross-stratification develops under moderately low rates of bed aggradation due to fallout from suspension onto the bed. With increasing angle of climb such that $\alpha < \beta$ the entire ripple form is preserved with ripple migration, including deposits of the stoss slope (Fig. 5-22D; such internal strata are often termed formsets). Such stratification is termed **supercritically climbing ripple cross-stratification**. (Note that when $\alpha = \beta$ the stratification is referred to as **critically climbing ripple cross-stratification** and the entire foresets are preserved.) With increasing angle of climb the thickness of the stoss-slope deposits increases and when the angle of climb reaches 90° the stoss and lee slope deposits are equal in thickness. The forms of ripple cross-stratification that develop with particularly high angles of climb indicate very rapid deposition from suspension while ripples were actively (but relatively slowly, migrating on the bed. Note that the forms of ripple cross-lamination shown in figure 5-22 occur in nature as a continuous range of forms from $0 \leq \beta \geq 90$. Figure 5-23 is a sketch of a common vertical sequence of the forms of ripple cross-stratification and shows the interpretation of this sequence that develops under waning flows, in terms of the rate of ripple migration and the rate of bed aggradation due to fallout from suspension..

The forms of climbing ripple cross-stratification shown in figure 5-22 are often termed **ripple-drift cross-lamination**. In addition to the terminology described above many workers use the classification scheme of Jopling and Walker (1968). In that scheme subcritically climbing ripple cross-stratification (such as that shown in Fig. 5-22B) is termed Type A ripple-drift cross-lamination; critically to slightly supercritically climbing ripple cross-stratification is termed Type B ripple-drift cross-lamination (e.g., Fig. 5-22C); and highly supercritically climbing ripple cross-stratification is termed sinusoidal ripple-drift cross-lamination (Fig. 5-22D).

Cross-stratification formed by dunes

As noted above, dunes produce forms of cross-stratification that are geometrically similar to cross-stratification produced by ripples. Figure 5-24 shows two forms of cross-stratification formed by dunes: planar crossbedding (both tabular and wedge-shaped) and trough cross-bedding. Both of these are distinguished from ripple cross-stratification by their larger scale and the cross-stratification produced by dunes is commonly termed “large-scale cross-stratification” in contrast to the “small-scale cross-stratification” produced by ripples. The geometry of the internal strata will be governed as outline in figure 5-21 and the bounding surfaces depend on the angle of climb of the bedform (although, as noted above, large dunes only rarely climb at high angles). The thickness of cross-strata sets formed by migrating dunes are generally larger than that produced by ripples (because dunes are higher) but set thickness is also governed by the angle of climb of the bedform. As noted above, the thickness

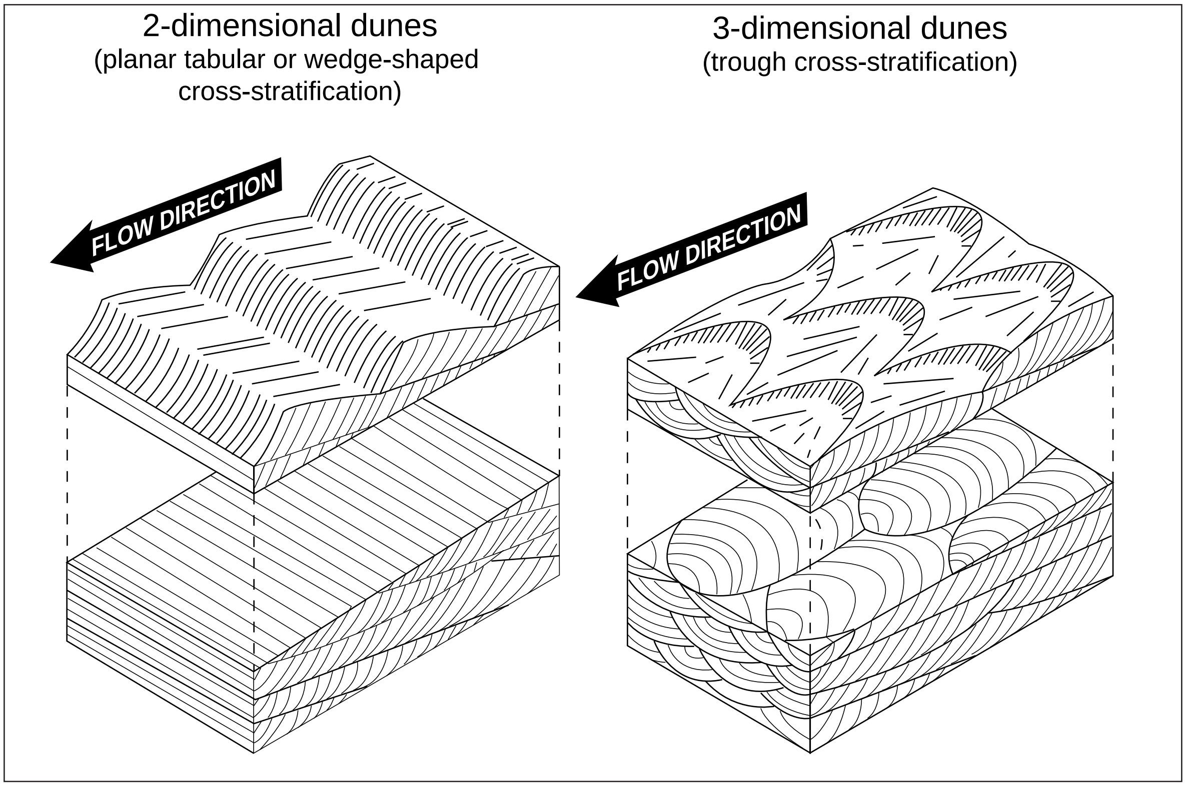


Figure 5-24. Forms of large-scale cross-stratification produced by dunes. See text for details. After Allen, 1970.

of internal strata is also greater than those produced by ripples; their is a general tendency for the thickness of internal strata to increase with the height of the lee face of the bedform.

Large scale planar cross-stratification is produced by 2-dimensional dunes that are dominated by deposition by avalanching on their lee slopes (i.e., internal strata may be inversely graded). Internal strata typically dip at angles of approximately 30° and typically range in thickness from a few decimetres up to several metres. In outcrop caution must be made in interpreting such cross-strata because the form will vary due to the relationship between the orientation of internal stratification and the orientation of the exposure. For example, in figure 5-24, planar cross-stratification looks like a horizontal stratification when viewed on a vertical section aligned normal to the flow direction (the direction of bedform migration; note that the internal strata seen in this view need not parallel the lower bounding surface exactly, as shown). Thus, to avoid mistakenly interpreting such cross-stratification you must try to view it on at least two vertical planes. Note that in plan view, through a planar cross-strata set, the straight, parallel strikes of the planar internal strata are visible, aligned normal to the direction of bedform migration. The internal strata, in plan view, may be traced laterally for up to tens of metres. To determine the paleoflow direction based on planar cross-stratification, we need to find the strike of internal strata and the dip direction: the paleoflow direction will be perpendicular to the strike, into the direction of dip.

Large scale trough cross-stratification is produced by 3-dimensional dunes where deposition on the lee face includes both contact and suspended loads. Dip angles of internal strata vary from 20 to 30°, generally lower than the planar forms, and sets range from a few decimetres to several metres in thickness. The form of such trough cross-stratification varies significantly depending of the view of the exposure. In vertical sections, parallel to the flow direction, sets may appear very similar to planar cross-stratification. However, in the vertical plane, normal to the flow direction, the diagnostic trough-shape is apparent. In plan view the form the stratification appears as inter-cutting, elongate troughs (defined by the bounding surfaces) and internal strata are curved, into the direction of bedform migration, where they terminate against their lower bounding surfaces; thus, their concave surfaces face the direction of migration. Such troughs, in plan view and in vertical section may extend for up to several tens of metres.

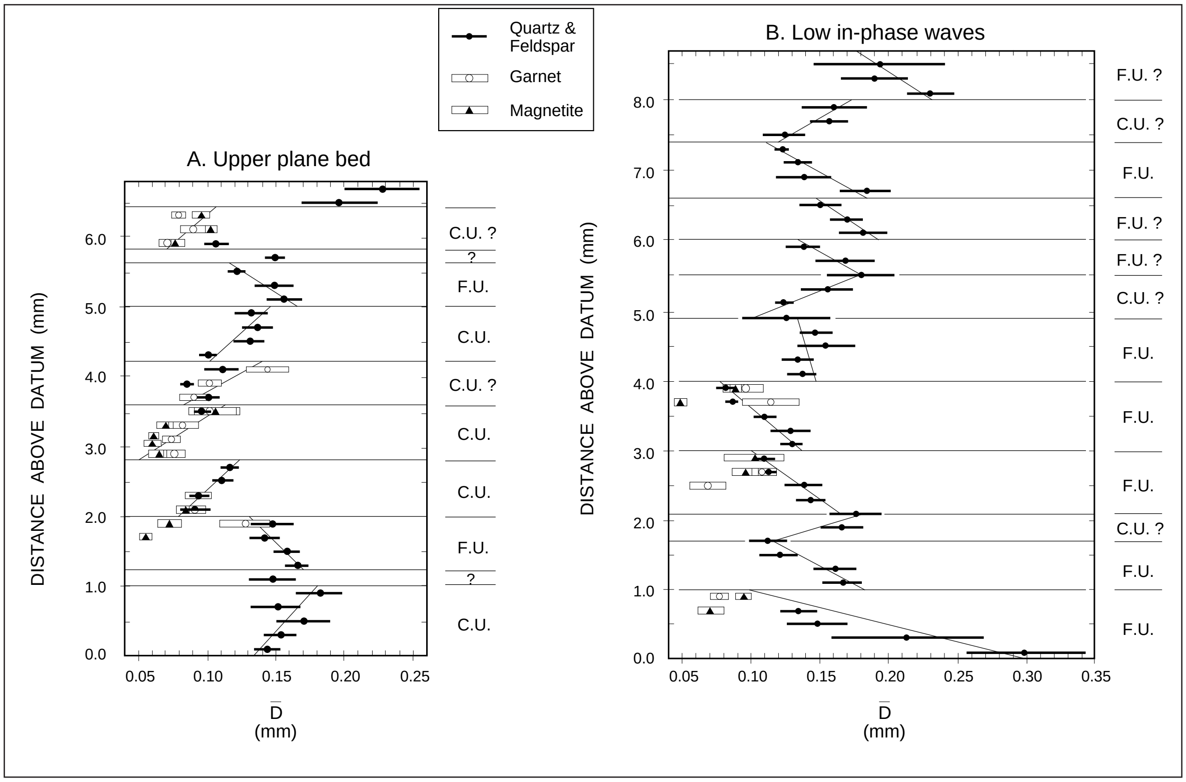


Figure 5-25. Internal grading and the distribution of heavy minerals in horizontally-laminated sediments: (A) deposited on plane beds, and (B), deposited on low, downstream-migrating in-phase waves. All points in each plot indicate the mean grain size in thin, contiguous layers up through the horizontally-laminated deposits. CU indicates coarsening-upwards lamina and FU indicates fining-upwards lamina. Question marks indicate that the textural interpretation is rather uncertain. See text for detailed discussion. After Cheel, 1990a.

Upper plane bed horizontal lamination

Deposition under upper plane bed conditions produces a variety of forms of **horizontal lamination** (sometimes termed "parallel lamination", although parallel laminae need not be horizontal). While one might think that horizontal lamination would be the simplest form of stratification to interpret (i.e., deposition on a planar, horizontal surface), this has actually been one of the most controversial styles of lamination. For example, a recent (not soon-to-be-published) manuscript listed no less than 20 hypotheses for the origin of horizontal lamination formed under upper plane bed conditions. Each such hypothesis suggests various sorting mechanisms that segregate sediment on upper plane beds into "packages" of distinct grains size (and/or sorting) or mineralogy that comprise laminae. It is likely that there are several mechanisms that cause the sorting of the sediment that leads to the formation of lamination on upper plane beds (or nearly plane beds with low bedwaves). The major mechanisms can be listed as: (1) local and small-scale sorting by the bursting process (both bursts and sweeps) on a true plane bed; (2) selective sorting by size, shape and density on a true plane bed; (3) migration of low bedforms over which sediment size, shape and density varies regularly. Each of these will be discussed below.

Bridge (1978) was the first to suggest that horizontal lamination (that is, thin, < 1 mm, horizontal lamination) formed in response to the bursting cycle. He postulated that temporally decreasing boundary shear stress associated with bursting would lead to deposition of increasingly finer sediment over the period of bursting and this sediment would be preserved as a fining-upwards lamination. However, Cheel and Middleton (1986) showed that horizontally-laminated sand and sandstone deposited under upper plane bed conditions were not composed of predominantly fining-upward lamination but consisted of a mixture of fining- and coarsening-upwards laminae of limited lateral extent (several millimetres across the flow direction). Such laminae are shown in figure 5-25A from

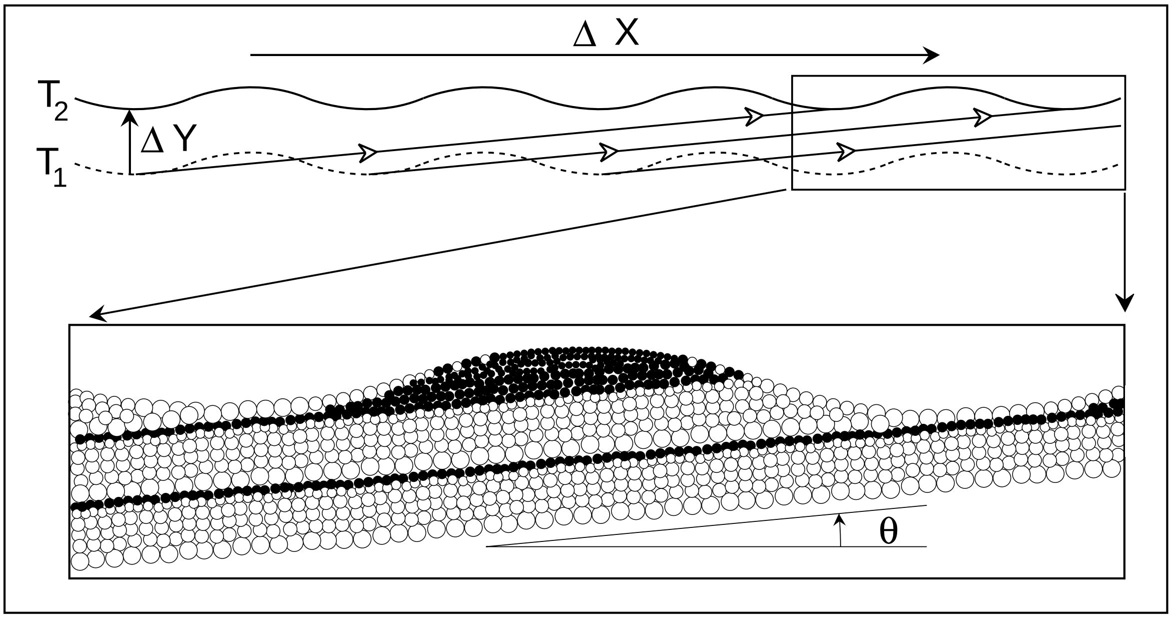


Figure 5-26. Schematic illustration showing the formation of fining-upward laminae with heavy minerals concentrated in the laminae tops due to migration of low in-phase waves with heavy minerals concentrated on their crests. Upper diagram shows the migration path of in-phase waves (lines with open arrows) during bed aggradation of ΔY and bed wave migration of ΔX from time T_1 to T_2. Lower diagram is the enlarged view of boxed area showing the nature of grading and the distribution of heavy minerals on the in-phase wave and within laminae produced with migration. θ is the angle of climb: $\theta = \tan^{-1} (\Delta Y / \Delta X)$. Vertical exaggeration approximately X25. From Cheel, 1990a.

upper plane bed deposits of a flume (the same textural and mineralogical characteristics were found in ancient sandstones and modern inter-tidal deposits). In addition to these "**textural" laminae**, they identified "mineralogical laminae" that were much more extensive than the textural laminae and were made up of very thin sheets of heavy minerals (termed "**heavy mineral sheets**") that seemed to be predominantly associated with coarsening-upwards textural laminae. They attributed the textural laminae to the bursting cycle and the heavy mineral sheets to the migration of heavy mineral accumulations over the active plane bed (e.g., heavy mineral shadows). In their model fining-upward laminae formed by bursting. They suggested that the ejection of fluid away from the boundary carried sand upwards and as the burst decayed this sand fell back to the bed, the largest, heaviest grains first, followed by the smaller, lighter grains (i.e., see Stokes' Law of settling); thus the material deposited on the bed following bursting would form an upward-fining layer. In their model coarsening-upwards laminae formed as an in-coming sweep locally increased the shear stress acting on the granular boundary and the sediment that was taken into transport would re-deposit just downstream as the sweep decayed. Shearing within the material in transport would cause the grains to experience dispersive pressure which pushes coarse grain upwards relative to the finer grains in the moving sediment layer on the bed. Deposition of this coarsening-upwards layer of moving sediment would form the coarsening-upwards lamination. Heavy mineral sheets represent a lag of fine-grained heavy minerals left behind as a heavy mineral accumulation passed over the bed. The fine, high-density grains would easily become deposited, particularly in the spaces between immobile quartz grains on the bed. The association of heavy mineral sheets with coarsening-upwards laminae suggests that the heavy mineral accumulations were responsive to the high bed shear stress due to sweeps but essentially armoured the bed to the effects of bursting. The formation of heavy mineral sheets, described above, is an example of lamination formation by selective sorting by size and density. Extensive horizontal lamination are probably commonly formed by this mechanisms (and low bedforms; see below) whereas the laminae formed by the bursting cycle comprise laterally less extensive lamination in upper plane bed deposits.

The formation of heavy mineral sheets may also represent an example of the formation of horizontal lamination by the migration of low bedforms or bedwaves. A heavy mineral accumulation may be thought of as a mineralogical bedform that migrates over the bed. In fact, the formation of a variety of textural laminae that are of much greater extent that the textural laminae formed by bursting must involve the migration of bedforms over an otherwise plane bed. This raises the question, again, of whether a bed with even the lowest bedwaves is really a plane bed (e.g., see Fig. 5-10). For now, we will extend the definition of upper plane bed to include "true" plane beds and "nominally" plane beds over which low bedforms may migrate. This is probably a reasonable extension of the definition of plane bed, for the purposes of our discussion of horizontal lamination because low bedwaves are definitely responsible for some forms of this structure.

It has long been known that very low bedforms such as washed-out dunes will produce a horizontal lamination if the preserved deposit generally lacks internal stratification (because the lamination is too thin to see the cross-stratification or because the sediment is too well-sorted to undergo significant segregation by size on the lee of the bedform). Indeed, as we saw in the section of bedform migration (see Fig. 5-22A) any asymmetric bedforms with a very low angle of climb might produce a form of horizontal lamination. However, there have been several suggestions of upper plane bed horizontal lamination formed by migration of low bedwaves on a nominally plane bed. For example, Allen (1984a) suggested that the passage of eddies over a bed would produce low-relief, downstream-migrating bedforms that would form extensive horizontal lamination. He described the predicted characteristics of such bedwaves in detail but such waves, with those characteristics, have not been identified in experimental or field studies. However, experimental work by Bridge and Best (1988) and Paola et al. (1989) and Cheel (1990a) have described a variety of low bedwaves that are thought to be responsible for the formation of extensive horizontal lamination under what are essentially upper flow regime plane bed conditions.

Bridge and Best (1988) described low, asymmetrical bedforms (possibly **very** washed out dunes) and noted that they formed extensive laminae in the bed material of their flume. Paola et al. (1989) described very low symmetrical bedwaves (presumably similar to those shown in Fig. 5-10) that migrated downstream. By recording the bed behaviour with high speed video photography they observed very low angle, parallel lamination (essentially horizontal lamination) formed by a combination of small-scale fluctuations of turbulence (bursting), selective sorting, and bedwave migration. Cheel (1990a) described the internal grading and distribution of heavy minerals in low, downstream-migrating symmetrical bedwaves on which grain size became finer towards the tops of the bedwaves and heavy minerals were concentrated in their tops. With migration of such low bedforms predominantly fining-upwards laminae formed with heavy minerals concentrated in the tops of such laminae (compare with the distribution of heavy minerals in true plane bed horizontal lamination, Fig. 5-25). Figure 5-26 illustrates the form and origin of such laminae. Coarsening-upward laminae, and some thin fining-upwards laminae, also associated with this form of horizontal lamination, were attributed to the action of bursts and sweeps, concurrent with bedform migration.

As for a conclusion to the origin of horizontal lamination, it is likely that all of the major mechanisms (and some others not discussed here) will lead to the formation of this structure. Some or all of the three major mechanisms listed above may act together to form horizontal lamination under the same flow conditions and any one deposit of upper plane bed, horizontally laminated sand or sandstone, probably preserve laminae formed by at least two, and possibly all three of these mechanisms. The laterally extensive forms of horizontal lamination certainly involve the migration of coherent sediment structures over an active plane bed (heavy mineral accumulations or full-fledged bedforms) whereas the more subtle laminae are produced by processes associated with near-boundary turbulence. The key to distinguishing the products of these various mechanisms likely lies in detailed studies of the textural characteristics (including grain size, shape and orientation) within individual laminae.

In-phase waves stratification

Descriptions of in-phase wave stratification have been limited and figure 2B shows rather "ideal" forms of internal stratification associated with the variety of forms of in-phase waves. The section on the formation of cross-stratification described conditions that are also necessary for preservation of in-phase wave stratification (i.e., net bed aggradation is require to form thick sequences of in-phase stratification, in conjunction with sorting by size and/or mineralogy that is required to produce visible stratification). However, relatively little is known about the specific origin of internal stratification produced by in-phase waves so that the discussion that follows will be largely limited to its geometry. Note that figure 2B is largely based on descriptions in the literature of such stratification, some of which is reviewed below.

Power (1961) coined the term “backset bed” for upstream-dipping cross-strata (becoming finer-grained towards their tops) formed under antidunes (while Power coined this widely-used term the so-called backset beds that Power described are probably not formed by antidunes at all!). Middleton (1965) characterised in-phase wave stratification by its association with upper plane bed horizontal lamination, low angle (<10°) cross-laminae dipping both upstream and downstream, and by the confinement of cross-laminae to symmetrical lenses related to the form of the in-phase wave. Hand, Wessel & Hayes (1969) described in-phase wave stratification in which cross-laminae dip at angles of up to 24°. Based on flume experiments, McBride *et al*. (1975) documented thin (0.2 to 4 mm thick), laterally extensive, near-horizontal, parallel lamination, characterised by alternating coarse and fine laminae, which formed by downstream migration of low in-phase waves. Allen (1966) noted that in-phase waves will form lenses of backset cross-strata only if the water waves break and/or migrate upstream, whereas undulating parallel laminae draped over the symmetrical bed forms are produced if the water surface wave grows in place during net deposition. Furthermore, downstream dipping cross-strata form if the in-phase waves migrate downstream (Allen, 1966; also see Allen, 1984b, Fig. 10-21). Barwis & Hayes (1985; p. 908) suggested that the occurrence of low angle truncation surfaces in massive or horizontally laminated sands may indicate the presence of in-phase waves. They also provided an excellent description of the variability of in-phase wave stratification on a washover fan in a barrier island complex. They noted that down-fan, in the flow direction, in-phase waves decreased in length (reflecting decreasing flow velocity) and amplitude and passed into plane bed. The form of the in-phase wave cross-strata within the deposits also varied downfan from: (1) lenses of backset cross-laminae, to (2) lenses of laminae subparallel to bounding surfaces, and to (3) lenses of foreset laminae. This sequence was interpreted to reflect downfan variation in the relationship between the water and bed surfaces from: (1) upstream-migrating in-phase waves, to (2) in-phase waves under stationary water surface waves, to (3) downstream-migrating antidunes. Langford & Bracken (1987) described variation in in-phase wave stratification in a fluvial setting, as lenses of backset and foreset cross-laminae of smaller downstream extent than cross-stratification formed under lower flow regime conditions.

Figure 2B shows the ideal forms of stratification produced by the specific behavioural forms of in-phase waves that have been recognized and shows a gradual continuum of stratification styles through the transition from washed-out dunes to antidunes. Washed-out dunes (climbing at a low angle) and upper plane bed, forming horizontal lamination, develops as the upper flow regime threshold is exceeded. Low (on the order of millimetre high) downstream-migrating in-phase waves develop next and form horizontal lamination of the characteristics described here (termed in-phase wave horizontal lamination; see Fig. 5-26). As the in-phase waves grow in height and wavelength they continue to migrate downstream, producing lenses of downstream-dipping cross-laminae (termed in-phase wave foreset cross-laminae). The absence of visible cross-stratification formed by low in-phase waves is likely due to the very small thickness of the deposit and to the relatively poor development of sorting and fabric along the low-angle lee of the bed form. The next bed phase is characterised by standing in-phase waves which produce laminae which approximately parallel the bed forms (termed in-phase wave drape-laminae). With the onset of upstream-migrating and/or breaking wave antidunes, lenses of upstream-dipping cross-laminae form (termed antidune backset cross-laminae). This sequence applies to the 2-dimensional forms of in-phase wave only. The 3-dimensional forms are not well known from flume studies and are not included here. Also not included in figure 2B are associations of the various forms of in-phase wave stratification, due to temporal variation in the form of in-phase waves. Given the behaviour of the in-phase waves bedforms shown in figure 5-11 it is certain that such complex associations are very important and may be diagnostic of stratification formed by in-phase waves.