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A consideration of the dune:antidune transition in fine gravel

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ABSTRACT

Hydraulic data defining the dune:antidune transition in fine gravel are compared with potential flow theory, and information is drawn from published experiments and field-based studies. Attention is given to both transitional bedforms and the development of downstream-migrating antidunes. In the latter case, most data pertain to sand beds and not to gravel. Empirical data provide some weak support for the theoretical notion that the transition occurs at progressively lower Froude numbers at greater relative depths. Although a critical Froude number of 0·84 may reasonably be applied for the beginning of the dune to antidune transformation, lag effects (and a possible depth limitation) ensure that transitional bedforms may persist across a broad range of Froude numbers from 0·5 to 1·8. This latter observation has great relevance for palaeohydraulic estimates derived from outcrop data. Whereas the application of theoretical bedform existence fields, based upon potential flow theory, to fine gravel was previously purely speculative, the addition of experimental and field data to these plots provides a degree of confidence in applying stability theory to practical geological problems. For the first time, laboratory data pertaining to downstream-migrating gravel antidunes are compared with theory. These bedforms have been reported from certain experimental near-critical flows above sand or gravel beds, but have been observed infrequently in natural streams. However, there are no detailed studies from natural rivers and only a few contentious identifications from outcrops. Nevertheless, the limited hydraulic data conform to theoretical expectations.

Keywords Antidunes, dunes, gravel, palaeohydraulics, supercritical flow.

INTRODUCTION

The development of mesoscale bedforms in gravel ($D_{50} > 2$ mm) is poorly understood (Carling, 1999). In particular, the hydraulic conditions associated with the bedform phase transformations from gravel dunes to antidunes or upper-stage plane beds (USPB) are well described from theory, but ill-defined using empirical data. Additional experimental studies in laboratory flumes, as well as field observations, are required, but these need to be developed within the context of existing knowledge. Consequently, this paper assesses published hydraulic data within a theoretical framework.

The data considered here pertain to dunes, antidunes and those bedforms transitional between the two bedform phases. As explained in the discussion on potential flow theory, dunes form in subcritical flows that do not exhibit accentuated standing waves, whereas antidunes form when rapid, shallow flows develop distinct surface standing water waves that interact with a mobile sediment bed (Allen, 1984). Transitional bedforms, as defined here, may exhibit morphological and sedimentological characteristics that pertain to dunes or antidunes or both because, as flows change, there is a time lag before the bedforms can adjust to any new imposed flow regime.

In the case of sand-sized sediments, the internal structures that may be ascribed to antidune deposition are reasonably well-documented from laboratory studies (e.g. Middleton, 1965; Jopling & Richardson, 1966; Yagishita & Taira, 1989; Prave & Duke, 1990; Yagishita, 1994; Yokokawa *et al.*, 1999; Alexander *et al.*, 2001). In contrast, there is a lack of detailed studies of the internal structure of gravel antidunes (Shaw & Kellerhals, 1977; Alexander & Fielding, 1997). Similarly, models linking the hydraulics of standing waves to antidune morphology and sedimentary structures are poorly developed (Hand *et al.*, 1972; Hand, 1974; Normark *et al.*, 1980; Langford & Bracken, 1987; Prave, 1990; Morris *et al.*, 1998; Blair, 1999; Fralick, 1999; Kubo & Yokokawa, 2001). Consequently, the veracity of any interpretations of gravel bedwaves found during study of modern sediments, or the internal structures encountered in the stratigraphic record, is conditional upon an adequate understanding of both the internal structures and the associated hydraulic processes. As will be shown in the next section, further research is required concerning the exact detail of turbulent flow structure associated with bedform initiation and development. The aim of this paper is thus to contribute to a better understanding of the bulk (averaged) properties of flow associated with gravel bedforms by comparing empirical data with summarized, simplified theory.

THEORY

A loose flat granular bed subject to a uniform turbulent shear flow may develop an internal periodic non-uniformity such that the time-average streamlines become deformed and interact with the bed surface. In time, the flat bed must deform into an undulating series of periodic bedforms (Yalin & da Silva, 2001). A range of theories, of varying complexity and completeness, has been proposed to account for the initiation of these bedforms from a plane bed (Smith, 1970; Engelund & Fredsøe, 1982). However, for several decades, approaches to the problem have largely revolved around two separate hypotheses (Raudkivi, 1997).

Potential flow theories

One argument invokes an inherent instability at the sediment–water interface owing to the stress imposed by the moving fluid (e.g. Bagnold, 1956; Kennedy, 1963; Reynolds, 1965; Engelund, 1966), whereby the effect of turbulence may be accounted for only indirectly (e.g. Richards, 1980). The characteristic wavelength and amplitude of the deformed bed is considered to be a function of the bulk flow properties, such as average flow speed or average shear stress. The classic approach considers the character of a potential channel flow above a fixed undular bed (Anderson, 1953; Kennedy, 1963). Recent studies by Huang & Song (1993) and Huang & Chiang (1999, 2001) allowed for bed deformation and a lag effect between sediment discharge and flow velocity. Nevertheless, the resultant bedform discriminatory curves are essentially similar to those first proposed by Kennedy (1963). In summary, stability analyses serve to demonstrate the significance of bulk flow parameters, notably the Froude number, in bedform mechanics (Engelund & Hansen, 1966; Graf, 1971; Colombini *et al.*, 1987), but provide no understanding of the detailed physics of bed deformation (Raudkivi, 1967; Coleman & Fenton, 2000; Gerkema, 2000).

Turbulent flow theories

An alternative view predicates that bedforms are initiated from bed defects as a result of turbulent bursting at the boundary and the development of flow separation (e.g. Grass, 1970; Raudkivi & Witte, 1990; Best, 1992, 1993; Yalin & da Silva, 2001). In time, this approach should lead to a better understanding of bedform development (Dinehart, 1999; Chanson, 2000). However, a general theory for bedform initiation based upon turbulence generation is lacking. In addition, although flow separation plays a significant role in the development of bedforms (Nelson *et al.*, 1995), separation is not necessary to induce an initial instability in the granular bed (Smith, 1970; Coleman & Melville, 1996; Coleman *et al.*, 1999; Yalin & da Silva, 2001). Rather, linear stability theories apply well to the generation of the initial defects (Smith, 1970; Coleman & Melville, 1994, 1996; Qihua *et al.*, 1999).

From the brief argument above, classic approaches to stability theory (Kennedy, 1963; Engelund, 1966; Engelund & Hansen, 1966) should delimit the existence fields in terms of bulk flow parameters for gravel bedforms. However, some bedforms may persist outside, but close to, their respective stability limits, owing to maintenance by turbulence generation, including flow separation. In addition, stability theory does not take into account any effects of the variability in the grain sizes of the bed sediment. These effects, for which there are no general theories, are believed to have only a minor influence on bedform morphology (e.g. Chiew, 1991; Coleman & Melville, 1996). Thus, some data scatter across existence field boundaries might be expected.

Theoretical framework for potential flow theory

In order to describe the results of potential flow analysis, it is necessary to define several parameters. The Froude number (*Fr*) can be defined as;

$$Fr = U/\sqrt{(gh)} \tag{1}$$

where U denotes the depth-averaged velocity, h is the average water depth, and g is the acceleration due to gravity. The wave number of the bed undulations (k) is given by the relationship: $k = (2\pi/L_b)$, where L_b denotes the bedform wavelength. Potential flow analysis delineates three regimes pertaining to bedforms (Fig. 1). These regimes are defined by:

$$Fr_l = \sqrt{\tanh(kh)/kh} \tag{2}$$

and

$$Fr_u = \sqrt{1/(kh)\tanh(kh)} \tag{3}$$

The regime $0 < Fr < Fr_l$ delineates lower flow regime (I in Fig. 1) conditions that are characterized by a water surface that is depressed over the crests of the wavy bed and elevated over the troughs. The undulations in the water surface elevation are thus out-of-phase with those of the bed, and this is the regime in which dunes commonly form. For the case of an individual dune, near-bed flow tends to accelerate from trough to crest and decelerate from crest to trough. Thus, for conditions of bed sediment transport, the flow pattern induces net erosion on the stoss side of the dune and net deposition downstream of the crest. This process induces downstream migration of the bedform. Individual dunes are also often temporally persistent with 'life spans' that are relatively long in contrast to the 'life spans' of antidunes, as noted below. This lower regime $0 < Fr < Fr_l$ corresponds to generalized subcritical conditions for bedforms of long wavelength, i.e. $Fr < 1$. However, and importantly, it is apparent from Fig. 1 that, for sufficiently short bed wavelengths (i.e. large values of kh), such as those pertaining to dunes and antidunes, the Froude number threshold for critical conditions drops noticeably below unity. Thus, although dunes may persist within transitional and supercritical flows as transitional bedforms (developed originally under lower stage flow conditions), there are no conditions under which dunes can develop from plane beds within supercritical flows and for large values of kh.

The regime $Fr > Fr_l$ delineates upper regime (II in Fig. 1), or generalized supercritical conditions, where water surface undulations are high at the crest of the bedform and low at the trough. That is, the water surface undulations are in-phase with the bed undulations. The bedforms for which water surface and bed undulations are in-phase have traditionally been classified as antidunes, regardless of their direction of migration. For sufficiently high Froude numbers, the surface waves tend to grow in amplitude before

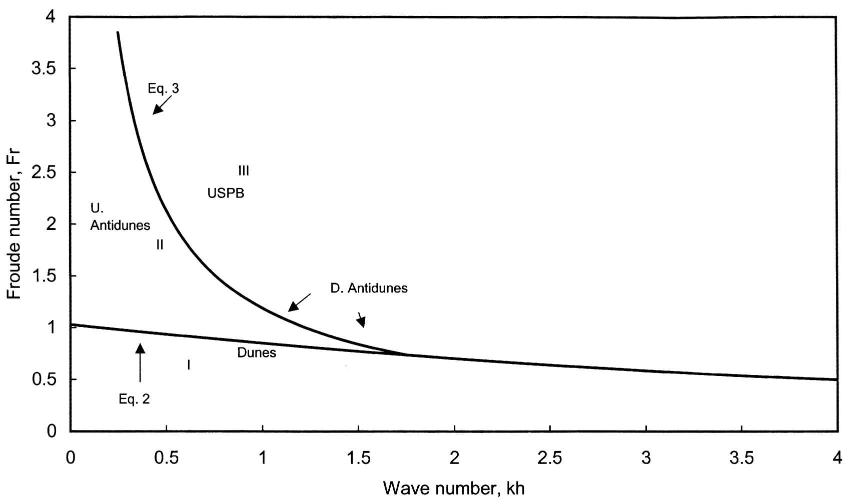

Fig. 1. Plot relating the Froude number and the non-dimensional bedform wavelength. The discriminators, Eqs 2 and 3, separate the bedform existence fields in accordance with the theories of Anderson (1953), Kennedy (1963) and Reynolds (1965).

breaking on their upstream sides, after which the waves collapse and the associated antidunes are washed out by sediment transport in the downstream direction. For a short period of time, in these shallower and higher velocity flows, a USPB may exist before the antidunes reform (Simons *et al.*, 1961; Grachev, 1980). Hence, individual antidunes are often less persistent than dunes. This antidune regime can be divided further into two more regimes, defined in terms of the Froude number Fr_u.

In the regime $Fr_l < Fr < Fr_u$, not only is the water surface in-phase with the bed, but the flow depth is elevated over the crests and depressed over the troughs. This creates near-bed flow deceleration from trough to crest and acceleration from crest to trough. Net deposition will occur upstream of the crest and net erosion downstream of the crest, such that the antidunes tend to migrate upstream. It is this regime that is most commonly reported within natural rivers and within experimental channels.

The regime $Fr > Fr_u$ (III in Fig. 1) cannot be obtained from classic theory of shallow-water waves (Kennedy, 1963; Song, 1983). However, for Froude numbers >1 (e.g. 1–2), the regime is peculiar to short wavelengths (large wave numbers) but, as the wavelengths increase, $kh \rightarrow 0$, for which $Fr_u \rightarrow \infty$, antidunes become muted in amplitude, and USPBs may be reported. Within this regime, and notably for short wavelengths, even though water surface undulations are in-phase with the bed, the amplitudes of the water surface undulations are muted compared with those of the bed. As a result, the flow depth is smaller over the crests than in the troughs. In this regime therefore, the near-bed flow accelerates from trough to crest and decelerates from crest to trough (i.e. the same pattern as for dunes; Song, 1983) and is one in which downstream-migrating antidunes become possible (Reynolds, 1965; Engelund & Hansen, 1966; Kennedy, 1969; Huang & Chiang, 2001).

PREVIOUS OBSERVATIONS OF DOWNSTREAM-MIGRATING ANTIDUNES

Because downstream-migrating antidunes might be confused with dunes, some further comment is warranted. Using the definitions given above, a persistent bed state consisting of downstream-migrating antidunes has been noted to develop in several flume studies of sand beds (ASCE, 1966), in which all flows were just supercritical (Kennedy, 1961; Foley, 1975; Fukuoka *et al.*, 1982). These bedforms tended to possess asymmetric shapes similar to dunes. Fukuoka *et al.* (1982) recorded flow separation in their leesides and noted that, as the flow depth is deeper within the troughs compared with the crests, both flow and bed morphology are superficially similar to the dune phase. In prescribed flow conditions, the downstream-migrating antidune bed state may be unstable, alternating with a dune bed state. Specifically, Kennedy (1961) observed rounded antidunes together with asymmetric downstream-migrating antidunes and dunes in complex temporal and spatial sequences within single flume runs.

In natural streams, downstream-migrating antidunes are most likely to be found on relatively steep slopes with shallow flow and plentiful sediment supply, such as typifies arroyos and beach rills (e.g. Barwis & Hayes, 1985). There are no unequivocal field data pertaining to gravel beds (see Dinehart, 1992a,b).

DATA REDUCTION FROM PRIMARY SOURCES

Data pertaining to gravel ($D_{50} > 2$ mm) were compiled from primary sources and checked for accuracy. Specifically, where possible, bulk flow parameters that were originally calculated from detailed flow measurements have been recalculated from the original tabulated data. Bed shear stresses (τ_0) are based upon Duboy's equation using reported energy slopes (S) or, where reach-scale uniform flow was reported, the slope of the flume bed. Duboy's equation is:

$$\tau_0 = \gamma h S \tag{4}$$

where $\gamma = \rho g$ is the specific gravity of water. The reach-averaged total shear velocity (u_{*tot}) is then given as

$$u_{*tot} = \sqrt{\tau_0/\rho} = \sqrt{ghS} \tag{5}$$

where ρ is the density of water and g is the acceleration due to gravity. In those cases in which the critical near-bed shear stress for incipient sediment motion (τ_c) was derived, then the Shields' non-dimensional ratio (θ) has been used to relate the shear stress to the characteristics of the bed material:

$$\theta_c = \tau_c/[(\rho_s - \rho)gD_i] \tag{6}$$

where ρ_s is the density of the sediment grains and D_i is a reference grain size.

In this paper, particular use is made of the data and observations from a few key Russian studies, and so it is worthwhile to report briefly on the nature of these experiments. Goncharov and colleagues (Goncharov, 1938; see also Kondrat'ev, 1962) used a 5·5 m long flume, 0·20 m wide to study separate size fractions from 0·35 mm to 12 mm. Flow depth varied between 0·07 and 0·10 m and was rarely up to 0·20 m. The basic data considered hydraulic conditions for initial motion, initial development of dunes, dunes of maximum steepness and the transition to USPB. The critical values of Fr for each bed state were found to increase systematically as the bed material was coarsened. Goncharov (1938) considered the transitional phases in 2·16 mm gravel and obtained low-amplitude dunes (L_b = 900 mm) with heights (H) of 10 mm for Froude numbers of $\approx$0·69. These bedforms washed out when $Fr > 0{\cdot}76$ and, for $Fr \approx 1{\cdot}05$, were replaced by well-rounded but asymmetrical downstream-migrating antidunes. USPBs were recorded for Fr > 1·48.

Kopaliani (1972) conducted 100 experimental runs in a 30 m long by 0·5 m wide glass-walled flume. Before every test, the sediment bed (D_{50} = 2·16 mm to 6·5 mm) was graded to the required slope above the horizontal flume base. An automatic sediment feeder at the upstream end of the flume maintained a constant bedload flux during the tests. Flow and bedform parameters were measured when flow and bedform characteristics were steady. Kopaliani's results for a 6·5 mm bedstock are particularly relevant. For subcritical Froude numbers ($Fr < 0{\cdot}76$) and low bed slopes, low-amplitude asymmetrical dunes were observed to form from a mobile flat bed. For separate experimental runs with transitional Froude numbers (Fr = 0·76–1·22, averaging 0·99 for 17 separate runs) and high bed slopes, more symmetrical bedforms developed that were in-phase with standing waves on the water surface. In addition, bedforms consistent with descriptions of downstream-migrating antidunes were reported. For higher Froude numbers, only upstream-migrating antidunes were observed.

RESULTS

Discriminatory Froude number as a function of relative depth

Figure 2 presents experimental data pertaining to dunes, antidunes and transitional bedforms, but field data are also included. Also shown are the rhomboid low-amplitude antidunes reported by Ikeda (1983), which formed in fine gravel in the

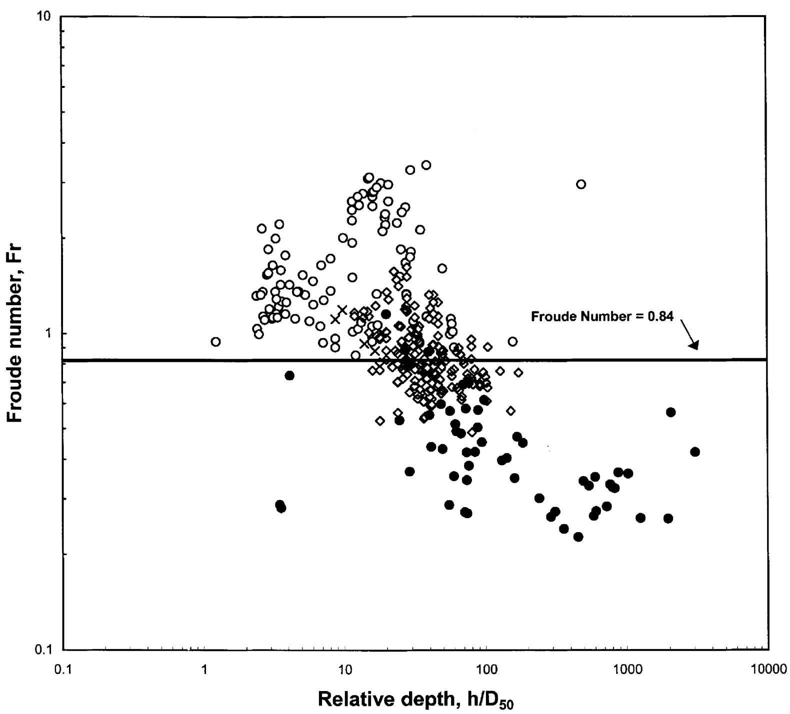

Fig. 2. Plot of Froude number vs. relative depth for 429 mesoscale gravel bedforms. Closed circles, dunes; open diamonds, transitional bedforms; open circles, antidunes; crosses, rhomboid low-amplitude antidunes. Curve shown is for Fr = 0·84, the theoretical discriminator of Kennedy (1963). Data sources: Goncharov (1938); Pushkarev (1948); Martinec (1967); Korchokha (1968a,b); Kopaliani (1972); Graf *et al.* (1983); Kopaliani *et al.* (1985); Bathurst *et al.* (1987); Blair (1987); Anon (1988); Whiting *et al.* (1988); Bennett & Bridge (1995); Alexander & Fielding (1997); Wieprecht (2000) and as cited by Carling (1999).

vicinity of $Fr = 1{\cdot}0$ and seem to be the same bedform as reported in medium sands beneath non-breaking standing waves for similar Froude numbers (Simons *et al.*, 1961). It is probable that these rhomboid antidunes develop in response to the presence of oblique interference waves, as has been shown to occur for experimental flows with Froude numbers slightly >1·0 (Robillard, 1965; Robillard & Kennedy, 1967). A few data sets that are plotted as 'transitional bedforms' may include examples of the 'end-members': dunes and antidunes. For example, Dinehart's (1992a) bedforms are all shown as transitional bedforms because Dinehart's original data do not discriminate between dunes and transitional dunes.

The spread of the 429 data points in Fig. 2 largely reflects the constraints of experimental conditions. Thus, most transitional bedforms have been reported for relative depths (h/D_{50}) between 10 and 100, but these are the relative depths most readily reproduced in laboratory flumes. Although antidunes are well represented, few dunes and no transitional bedforms have been reported for relative depths < 10, even though theory suggests that they should be present (Engelund & Hansen, 1967; Cooper *et al.*, 1972). Consequently, as the transition for small relative depths may occur at small values of shear stress (Fredsøe, 1979), the lack of data may reflect a tendency for very low-amplitude, long-wavelength bedforms in shallow turbulent flow to go unnoticed or unreported (Carling, 1999). There are few data for relative depths > 100, and most of these data are derived from observations in natural streams.

The data viewed *en masse*, or set by set, cannot be used unequivocally to demonstrate the theoretical expectation (Kennedy, 1961; Reynolds, 1965; Vanoni, 1977) that the critical Froude number for the dune:antidune transition declines systematically as the relative depth increases. Only a few sets of data (e.g. Goncharov, 1938; Kopaliani, 1972) show a weak negative trend (Fig. 3) as was noted by Kondrat'ev (1962; fig. 74). In particular, the comprehensive data of Bathurst *et al.* (1987), Graf *et al.* (1983) and Wieprecht (2000) show weak trends in depth dependence. Thus, Figs 4 and 5 indicate that antidunes can develop directly from mobile flat gravel beds, or a dune bed state, at slightly lower Froude numbers as relative depth increases. None of the data sets shows a widening of the USPB phase between dune and antidune bed states, as has been claimed for coarser sediments (Tanaka, 1970). Instead, the transition remains

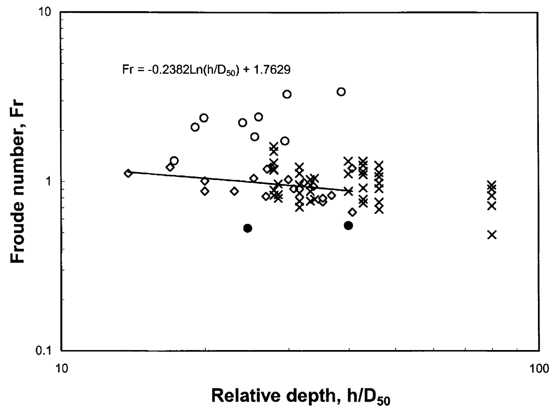

Fig. 3. Plot of Froude number vs. relative depth for data of Kopaliani (1972) and Goncharov (1938) depicting the bedform phase transition from dune (closed circles), through transitional bedforms (crosses), to downstream-migrating asymmetrical antidunes (open diamonds) and downstream-migrating symmetrical antidunes (open circles). An indication of a reduction in the critical value of *Fr* as relative depth increases is provided by the logarithmic least-squares equation and trendline fitted to the downstream-migrating antidune data.

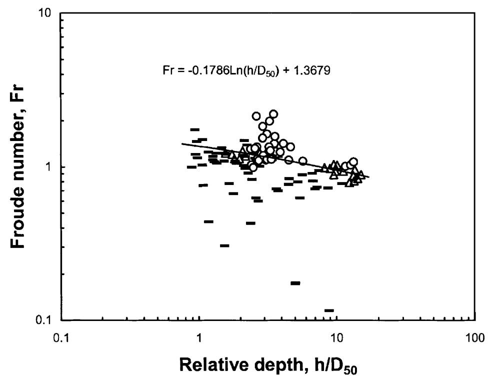

Fig. 4. Plot of Froude number vs. relative depth for data of Bathurst *et al.* (1987) depicting the bedform phase transition from stable flat bed (horizontal bars), through plane mobile bed (open triangles), to antidunes (open circles). An indication of a reduction in the critical value of *Fr* as relative depth increases is provided by the logarithmic least-squares equation and trendline fitted to the plane mobile bed data.

ill-defined (Engelund & Hansen, 1967), and antidunes often directly replace dunes, as has been reported for coarse sand (Simons *et al.*, 1965; Williams, 1967). For practical purposes, the discriminator $Fr = 0{\cdot}84$ is appropriate to separate dunes from antidunes (Kennedy, 1963). Nevertheless, transitional gravel bedforms straddle the discriminator and have been reported for a broad range of Froude number ($Fr = 0{\cdot}5$–$1{\cdot}8$).

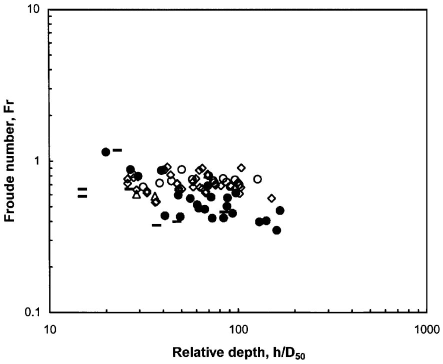

Fig. 5. Plot of Froude number vs. relative depth for data of Wieprecht (2000) depicting the bedform phase transition from stable flat bed (horizontal bars), through dune (closed circles) to transitional bedforms (open diamonds) beneath gentle standing waves, to plane mobile bed (open triangles) or antidunes (open circles). Bedstock consisted of uniform-sized 2-mm or 3-mm gravel.

For example, Kopaliani *et al.* (1985) reported transitional dunes spanning the range $Fr = 0{\cdot}80$–$1{\cdot}22$, which were replaced by symmetrical antidunes for $Fr > 1{\cdot}30$. This is in accordance with most other studies, which demonstrate that transitional bedforms cannot be sustained above $Fr \approx 1{\cdot}3$. Nevertheless, the transition zone is wide, and this may reflect lag effects in bedform adjustment to changes in flow, but also the effect of relative depth noted above (e.g. Kennedy, 1961; Ikeda, 1983). The time lag in bedform adjustment tends to become extended as grain size increases (Engelund & Fredsøe, 1974) and, thus, the persistence of transitional bedforms in gravel is not surprising. In particular, Wieprecht (2000) noted that inertia in the sediment phase meant that dunes developed in 2-mm or 3-mm gravel did not diminish close to the transition. Instead, dunes were replaced, first by symmetrical stationary incipient antidunes beneath non-breaking standing waves and, latterly, at higher Froude numbers, by upstream-migrating antidunes beneath standing waves, which often broke on the upstream sides. No intervening USPB phase was observed between dunes and antidunes.

Criteria for stability in terms of velocity ratio and Froude number

In the present analysis, it seems appropriate initially to use u_* in the restricted sense of Allen (1984) as that pertaining to grain roughness alone at incipient motion (u_{*crit}) on a planar bed (see Fig. 6). Alternatively, as the presence of a bedload and form roughness alters the friction, values of the total shear velocity, u_{*tot}, measured over the bedforms (Eq. 5 and Fig. 7) may be appropriate (Engelund & Hansen, 1966). Engelund and Hansen (1966) estimated expansion losses as follows. The reach-scale energy gradient is expressed as:

$$S = S' + S'' \tag{7}$$

where S' is the component originating from the friction along the upstream side of the bedforms, and S'' is the loss resulting from flow expansion. The estimated expansion loss is:

$$S'' = (H/2hL_b)Fr^2 \tag{8}$$

Engelund & Hansen (1966) found good agreement between theory, flume and field data for both upstream-migrating antidunes where there was no associated flow separation in the leesides and plane beds. Engelund & Hansen (1966) did not include data for downstream-migrating antidunes because they suspected that these bedforms would have expansion losses. This latter condition was defined as occurring where $k_s/D_{65} > 5$. Here, k_s is Nikuradse's equivalent sand-grain roughness determined by Engelund and Hansen (1966) from the velocity profile above a hydraulically rough bed:

$$U/u_* = 6{\cdot}0 + 2{\cdot}5\ln(h/k_s) \tag{9}$$

Consequently, data for downstream-migrating antidunes in Figs 6 and 7 are plotted with the ratio U/u_* corrected for expansion losses using reach-scale energy gradient data reported in the primary sources and Eqs 4, 5, 7 and 8.

Criteria for stability: Froude number and non-dimensional wave number

Figure 8 shows bedform phase discriminators based upon Froude number and the non-dimensional bedform wavelength (kh), according to Anderson (1953), Kennedy (1963) and Reynolds (1965). These discriminators have been validated for sand-sized sediment alone within only limited regions of the different bedform existence fields (Allen, 1984). Three discriminatory curves are asymptotic to Eq. 2 (Kennedy, 1963). Equation 2 separates dunes from antidunes (Allen, 1984). The function:

$$Fr = \sqrt{1/kh} \tag{10}$$

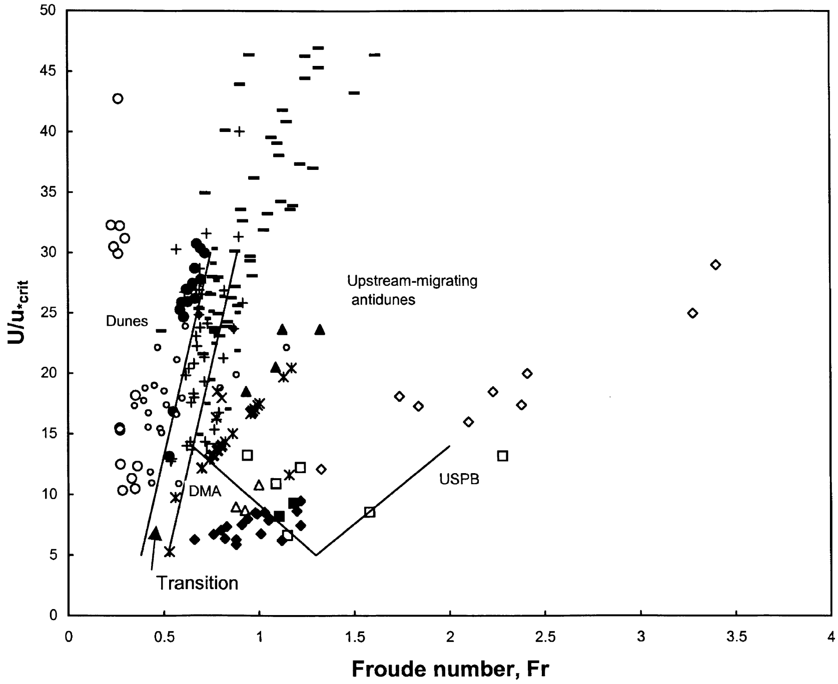

Fig. 6. Plot of the velocity ratio vs. Froude number for a variety of mesoscale gravel bedforms. The ordinate uses the critical shear velocity for initial motion as the denominator (see text). The discriminators separating bedform phase existence fields are in accordance with the theory of Engelund & Hansen (1966). DMA refers to downstream-migrating antidunes. Data sources: long horizontal bars, undifferentiated dunes, downstream-migrating antidunes and transitional bedforms (Goncharov, 1938); crosses, low-amplitude transitional dunes; closed triangles, upstream-migrating steep antidunes; open triangles, upstream-migrating low-amplitude antidunes; closed squares, rhomboid bedforms on USPB (Ikeda, 1983); open circles, River Polmet and River Luznice dunes (Martinec, 1967; Korchokha, 1968a,b); asterisks, downstream-migrating low-amplitude bedforms; closed diamonds, downstream-migrating asymmetrical antidunes; open diamonds, downstream-migrating symmetrical antidunes; open squares, plane mobile bed (Kopaliani, 1972; Kopaliani *et al.*, 1985); closed circles, transitional dunes (Pushkarev, 1948; Kopaliani *et al.*, 1985); open circles, dunes; short horizontal bars, upstream-migrating antidunes; plus signs, transitional bedforms beneath standing waves (Wieprecht, 2000).

defines the upper limit to two-dimensional bed-waves according to Kennedy (1963). Equation 3 defines an absolute upper limit to bed-wave occurrence according to Reynolds (1965), whereas the function:

$$Fr = \sqrt{1/kh(\tanh kh - 2/\sinh 2kh)} \qquad (11)$$

defines the limit according to Anderson (1953).

As was the case for Figs 6 and 7, it is appropriate to adjust the data for downstream-migrating antidunes to account for energy losses associated with flow expansion and separation. In addition, for downstream-migrating antidunes, there is a phase shift between bedform waves and the water surface elevation (Kennedy, 1969), such that the location of maximum velocity and the position of net scour or fill are discordant. Further, because downstream-migrating antidunes form, decay and reform rapidly (Kennedy, 1961), spatially periodic flow cells develop with a length scale that may differ locally from the length scale of the bed undulations (e.g. Colombini *et al.*, 1987; Mel'nikova & Petrov, 1993; Shvidchenko & Pender, 2001). Klaven & Kopaliani (1974) measured the down-channel lengths of these cells using the same experimental conditions as those associated with the development of Kopaliani's (1972) downstream-migrating antidunes. The lengths of cells were determined using a flow visualization method similar to that reported by Shvidchenko & Pender (2001). The data indicate that the length of the turbulence structures (L_t) ranged between 95% and 149% of the average bedform wavelength. However, on average, L_t was typically 22% longer than the average bedform wavelength (L_b). If individual values of L_b for downstream-migrating asymmetric antidunes are used to plot Kopaliani's (1972) data in Fig. 8, the data would fall slightly right of the asymptote. However, if the individual data points are adjusted by increasing each value

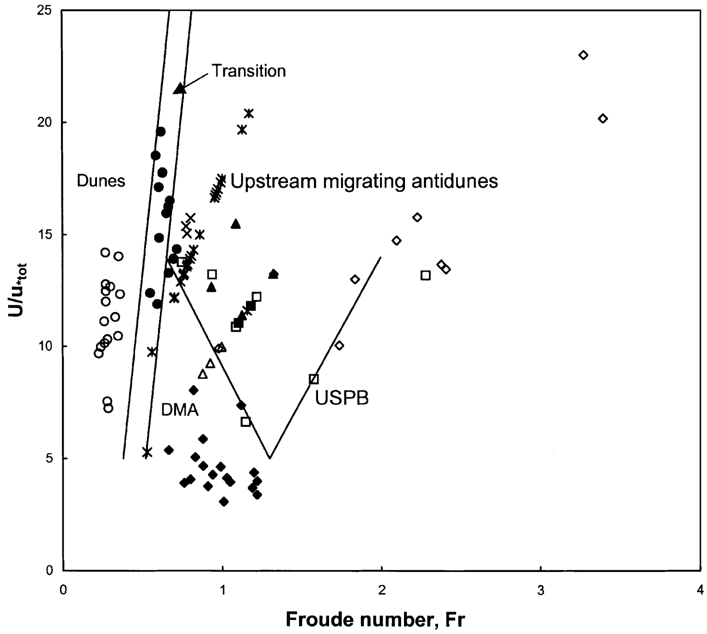

Fig. 7. Plot of the velocity ratio vs. Froude number for a variety of mesoscale gravel bedforms. The ordinate uses the total shear velocity as the denominator (see text). The discriminators separating bedform phase existence fields are in accordance with the theory of Engelund & Hansen (1966). DMA refers to downstream-migrating antidunes. Data sources: crosses, low-amplitude transitional dunes; closed triangles, upstream-migrating steep antidunes; open triangles, upstream-migrating low-amplitude antidunes; closed squares, rhomboid bedforms on USPB (Ikeda, 1983); open circles, River Polomet, River Luznice and River Anuy dunes (Martinec, 1967; Korchokha, 1968a,b; Bashkov from Anon, 1988); asterisks, downstream-migrating low-amplitude bedforms; closed diamonds, downstream-migrating asymmetric antidunes; open diamonds, downstream-migrating symmetrical antidunes; open squares, plane mobile bed (Kopaliani, 1972; Kopaliani *et al.*, 1985); closed circles, transitional dunes (Pushkarev, 1948; Kopaliani *et al.*, 1985).

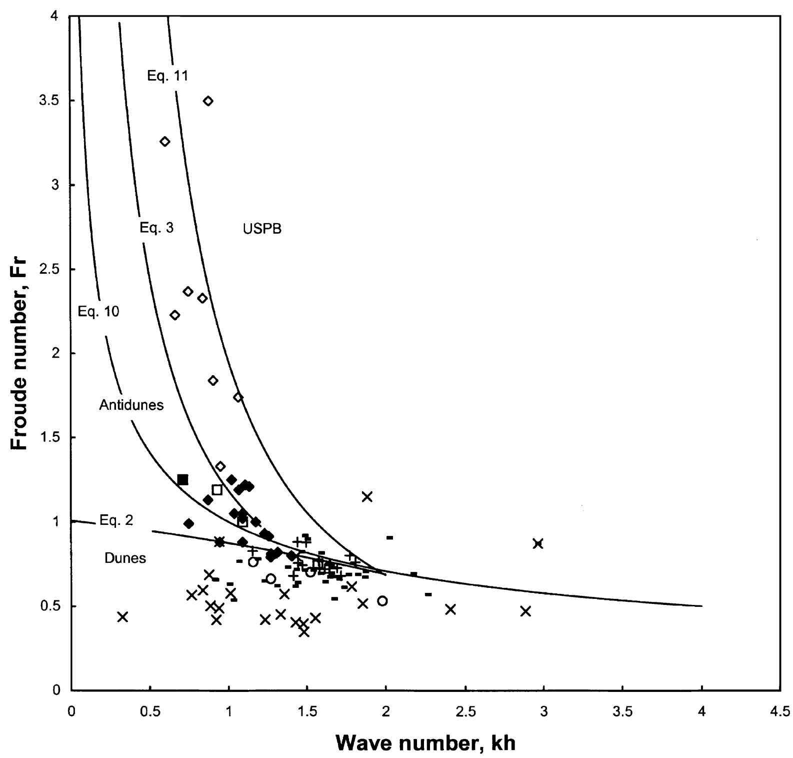

Fig. 8. Plot of empirical data pertaining to the Froude number and the non-dimensional bedform wavelength. The discriminators separate bedform existence fields in accordance with the theories of Anderson (1953), Kennedy (1963) and Reynolds (1965). Data sources: Kopaliani (1972); open circles, dunes to transitional dunes; open diamonds, downstream-migrating symmetrical antidunes; closed diamonds, downstream-migrating asymmetric antidunes. Kennedy (1961) open squares, transitional dunes; closed squares, downstream-migrating antidunes. Wieprecht (2000) crosses, dunes; horizontal bars, transitional bedforms beneath standing waves; plus signs, upstream-migrating antidunes beneath breaking standing waves.

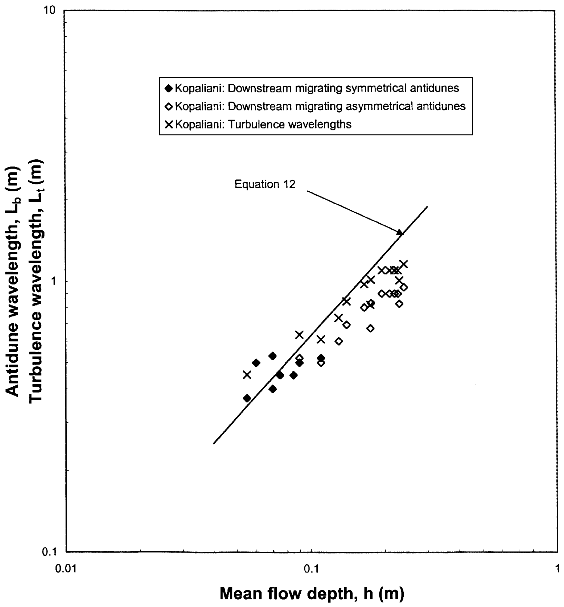

Fig. 9. Group mean wavelength vs. flow depth. Data source: Kopaliani (1972).

by 22%, then the rescaled data (as shown in Fig. 8) plot just to the left of the asymptote. The data of Kopaliani (1972) for symmetrical downstream-migrating antidunes fall between, and parallel to, the curves of Eqs 3 and 11, which define the upper limit for bed waves. However, no adjustment has been made to the plotted position of these data, as there is no information regarding the energy losses associated with these symmetrical gravel bedforms. Generally, it is known that symmetrical sandy antidunes generate negligible expansion losses (Simons *et al.*, 1961).

Allen (1984), among others, deduced from the theory of Kennedy (1963) that a simple relationship should exist relating the formative mean water depth to the group-average length of antidunes, where for freshwater flow of uniform density

$$L_b = 2\pi h \tag{12}$$

Thus, the length of antidunes preserved in outcrops may be used to infer palaeoflow depth. Allen (1984) supported this contention with flume data for antidunes developed in sand beds. The L_b:h antidune data of Kopaliani (1972) for 6·5-mm gravel are shown in Fig. 9 in comparison with Eq. 12 and show good agreement between theory and the empirical data for symmetrical downstream-migrating antidunes. However, the data pertaining to asymmetrical downstream-migrating antidunes plot roughly parallel to, and below, the curve defining Eq. 12. However, values of L_t plot in accordance with the curve. These latter two observations indicate, as noted by Kennedy (1961), that downstream-migrating asymmetric antidunes are an unstable bedform, continually collapsing, washing out and reforming, such that the wavelengths of the individual bedforms, which of course move at a speed less than the flow speed, are not necessarily well-adjusted to the scale of the generating turbulent structures within the flow.

DISCUSSION

Application of stability theory to practical geological applications

When considering the range of relative depths associated with gravel bedforms, the general discriminator $Fr = 0{\cdot}84$ is appropriate to separate gravel dunes from gravel antidunes. However, both theory and empirical data demonstrate that the transformation may occur for Froude numbers significantly below 0·84 (Chanson, 2000) and may be delayed until $\approx Fr = 1{\cdot}0$. Additionally, transitional bedforms rarely persist above $Fr \approx 1{\cdot}3$, whereupon upstream-migrating antidunes or USPBs occur. The broad range of Froude numbers across which transitional bedforms may be recorded largely reflects the lag time required to modify gravel bedforms.

The empirical data are largely consistent with the theoretical discriminators of Engelund & Hansen (1966). These discriminators are appropriate where u_{*crit} is used as the denominator in the ordinate, as in Fig. 6, but data frequently plot in inappropriate positions with respect to the discriminators if u_{*tot} is used (Fig. 7). With reference to Fig. 6, despite reasonable concordance between empirical data and theory, there is little experimental evidence for the occurrence of a narrow USPB phase existence field separating dunes from antidunes. However, a lag effect, whereby there is a propensity for both dunes and antidunes to persist during phase changes, may mask the final equilibrium bed state. Thus, long-period flume studies might be necessary to elucidate gravel bedform dynamics and, in particular, to define the USPB phase. Further, the effects of grain sorting on bedform existence fields (e.g. Chiew, 1991) have not been addressed in the present study. With these provisos in mind, Figs 6 and 7 offer a practical means of ascertaining probable formative flow environments from either outcrop data or direct measurements of inactive antidunes in rivers and in the ocean.

In contrast, Kennedy's (1963) ideal fluid flow calculations do not require knowledge of the bed shear stress. Rather, bulk flow properties, such as mean velocity and water depth, have often been used to estimate antidune wavelengths (e.g. Pantin, 1989; Prave, 1990). Alternatively, measurements of presumed antidune wavelengths and heights (e.g. Hand *et al.*, 1969; Hand *et al.*, 1972; Pickering, 1995; Pickering *et al.*, 2001) have been used to estimate flow properties.

For example, Mel'nikova & Petrov (1989) used flume data for a mobile sand bed to support the theoretical relationship, based on potential flow theory, between bedform wavelength and flow speed for subcritical flows with dunes that also included transitional conditions when Fr $\approx$0·86.

$$L_b = C^2\left(\frac{g}{2\pi}\tanh\frac{2\pi h}{L_b}\right)^{-1} \tag{13}$$

where C is the wave celerity, which may reasonably be assumed to be equivalent to the flow velocity for conditions in which waves begin to spill. Chanson (2000) reviewed potential flow theory and empirical studies and suggested a theoretical relationship for Froude numbers across the transition (0·896 < Fr < 1·589):

$$L_b/h \approx 2\pi Fr^2 \tag{14}$$

Equation 13 provides somewhat lower estimates of bedform wavelength compared with Eq. 14 and the other empirical functions reported by Chanson (2000). Nevertheless, the estimates are broadly consistent. Thus, although little progress has been made in recent years in understanding the controls on antidune development in gravel beds, Eqs 12–14 do allow estimates of bedform wavelength or flow conditions to be derived, such that the wave number (k) can be derived. Figure 8 may then be used in the same manner as has been suggested for Figs 6 and 7.

CONCLUSIONS

Hydraulic data defining the dune:antidune transition in fine gravel compare well with the existence fields for bedforms developed using potential flow theory. There is some empirical evidence to support the theoretical notion that the transition occurs at progressively lower Froude numbers as the relative depth increases. However, for practical applications, a critical Froude number of 0·84 may reasonably be taken as the discriminator between dunes and antidunes. Additionally, lag effects mean that transitional bedforms may persist across a broad range of Froude numbers from 0·5 to 1·8. The existing limited field and laboratory data pertaining to downstream-migrating gravel antidunes compare well with theory.

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NOTATION

$D_{50,65}$ Grain size at 50th or 65th percentile of cumulative grading curve
Fr Froude number
Fr_{crit} Critical value of Froude number
f Darcy–Weisbach friction factor
g Acceleration due to gravity
H Height of bedform
h Water depth
k Wave number
k_s Nikuradse's equivalent sand-grain roughness
L_b Length of bedform
L_t Length of coherent turbulence structure
S Reach-averaged bed slope
S' Form component of bed slope
S'' Expansion loss component of bed slope
U Section-averaged velocity
u^* Shear velocity
u^*_{crit} Critical value for initial motion
u^*_{tot} Total value for bedform reach
θ_c Critical value of Shields non-dimensional shear stress
ρ Density of water
ρ_s Density of sediment
τ_o Bed shear stress
τ_c Critical value of bed shear stress

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