DA1–DA4 files

DA1. GRAVITY EFFECT ON SANDY BEDLOAD

The following section outlines present understanding of how gravity affects bedload transport flux. The initial models based on Bagnold's (1963) work are described first, because they provide a useful framework for understanding the influence on bed morphology. Bagnold's approach is considered valid only for highly developed, dense bedloads (Leeder 1979), so more recent work describing effects of individual saltating particles is also described.

For a simple horizontal sandy bed affected by a strong uniform, steady current, Bagnold's original formula is still considered reasonably accurate (Soulsby 1997):

$$Q_b \text{ (kg/m/s)} \propto \tau_b^{1/2}(\tau_b - \tau_0); \qquad \tau_b > \tau_0 \quad \text{(ES1)}$$

where τ_b is the shear stress imposed by bottom current on the bed and τ_0 is a threshold stress for sediment motion (the symbol "$\propto$" means "is proportional to", i.e., constants are left out for simplicity). Equation DA1 follows a similar relation found empirically in earlier flume experiments (Meyer-Peter and Muller 1948). Because the bed shear stress $\tau_b = C_d\rho_w U^2$, where C_d is a bed friction factor, ρ_w is seawater density, and U is mean flow velocity above the bottom boundary layer, Equation DA1 can also be written $Q_b \propto C_d^{3/2}U(U^2 - U_0^2)$ where U_0 corresponds with τ_0. Thus, for a bed of uniform and unchanging C_d, flow significantly faster than U_0 leads to approximately $Q_b \propto U^3$.

Nielsen (1992) described how Equation DA1 could arise based on Bagnold's original arguments. The mobile bedload imposes a normal stress σ_e on the lower immobile bed equal to the submerged weight of the bedload (allowing for buoyancy):

$$\sigma_e = (\rho_g - \rho_w) g \int_0^\infty C(z)dz \qquad \text{(ES2).}$$

where ρ_g is the sediment grain density, g is the gravitational acceleration, and C is the sediment volumetric density. Assuming that a simple Mohr-Coulomb yield criterion applies to the top of the immobile layer (dashed line in Figure 1A) and that the shear stress at that level equals the shear stress imposed by the current (i.e., the solid phase acquires the current shear stress perfectly through grain to grain collisions (Bagnold 1963)), the flow-imposed shear stress is

$$\tau = \tau_0 + \sigma_e \tan\phi_s \qquad \text{(ES3).}$$

where ϕ_s is the sediment's angle of internal friction. The amount of bedload mobilized then relates to the excess imposed shear stress:

$$\int_0^\infty C(z)dz = \frac{\tau_b - \tau_0}{(\rho_g - \rho_w) g \tan\phi_s} \qquad \text{(ES4).}$$

Thus, the term $(\tau_b - \tau_0)$ in Equation DA1 could arise from friction - a larger imposed stress mobilizes a greater amount of sand, leading to greater bedload flux. Nielsen noted that too few data are available on the velocities of individual grains to then predict the resulting bedload flux, but the fact that various flux measurements follow Equation DA1 (Bagnold 1980; Nielsen 1992) suggests that mean particle velocities scale with flow shear velocity, as found by tracking particles using high-speed film (Fernandez Luque and van Beek 1976).

The effect of sloping beds is illustrated in Figure 1 (shear stress due to the current is omitted for simplicity). For a longitudinal gradient, the normal stress in Equation DA2 is modified by the factor cosγ and the total shear stress acting on the threshold surface includes the component of bedload weight. Equation DA3 then becomes (Bagnold 1963):

$$\tau = \tau_0 + (\tan\phi_s + \tan\gamma)\cos\gamma\left(\rho_g - \rho_w\right)g\int_0^\infty C(z)dz \qquad \text{(ES5).}$$

(γ here is negative for a down-gradient.) The amount of bedload of Equation DA4 should then be modified, with greater amounts mobilized on down-gradients:

$$\int_0^\infty C(z)dz = \frac{\tau_b - \tau_0}{(\tan\phi_s + \tan\gamma)\cos\gamma\left(\rho_g - \rho_w\right)g} \quad \text{(ES6).}$$

Bagnold (1963) derived Equation DA1 by assuming that the power expended by the shearing bedload was a simple proportion of the power expended by the current. Bagnold's energetics argument was extended to arbitrary slopes (Bailard and Inman 1981; Huthnance 1982a, 1982b) by assuming that the flux magnitude is proportional to the current's power expenditure but flux direction is governed by the vectorally combined stresses due to the current and down-gradient component of bedload weight. The bedload flux $\mathbf{Q_b}$ is then:

$$\mathbf{Q_b} = S\,|\mathbf{u}|^2(\mathbf{u}-\lambda|\mathbf{u}|\nabla\mathbf{H})/g \qquad \text{(ES7)}$$

where S is a constant, $\lambda = 1/\tan\phi_s$ and $\nabla\mathbf{H}$ is bed gradient (bold symbols represent vectors and |...| the vector magnitude).

Criticisms have been made concerning Bagnold's approach. In his model, fluid momentum is transferred to moving particles so that the fluid shear stress becomes insignificant at the base of the mobile layer. This only occurs, however, if the bedload is well-developed, otherwise the bedload is better described as isolated saltating particles than as a continuous layer (McEwan et al. 1999; Niño and Garcia 1998; Seminara et al. 2002). Saltation models of varying complexity have been developed (McEwan et al. 1999; Niño and Garcia 1994; Niño and Garcia 1998; Wiberg and Smith 1985, 1989), which variously incorporate particle extraction from the bed, trajectory, rebound or deposition, and dislodgement of bed particles. Trajectories are potentially affected by lift caused by fluid shear or particle rotation (Leeder 1979). Despite their complexity, these models can reproduce the variations in Equation DA1 remarkably well (e.g., McEwan et al. 1999).

Sekine and Parker (1992) summarized models of bedload on transverse slopes. They suggested that the components of flux down-gradient q_n and in the direction of the current q_s can be separated. If there is no current down-gradient, their ratio is

$$q_n/q_s = -B\tan\gamma; \quad B = B_0(\tau_c/\tau_b)^m \qquad \text{(ES8).}$$

Depending on the model, the coefficient B_0 incorporates the sand friction coefficient and other parameters. The different models summarized by Sekine and Parker (Engelund 1974; Hasegawa 1981; Ikeda 1982; Kikkawa et al. 1976; Parker 1984; Struiksma et al. 1985), their own results of numerical simulations of saltation, and a more recent model based on entrainment rates varying with shear stress (Parker et al. 2003) predict $m = 0$ to 1.0. If $q_s \propto \tau_b^{1.5}$ (Equation DA1, omitting the threshold for simplicity), such values of m imply that the down-gradient flux variation lies between $q_n \propto \tau_b^{1.5}$ (i.e., $q_n \propto u^3$) and $q_n \propto \tau_b^{0.5}$ ($q_n \propto u^1$). The wind-tunnel data of Ikeda (1982) and recent model of Parker et al. (2003) are consistent with $q_n \propto \tau_b^{1.0}$ ($q_n \propto u^2$) for large τ_b.

Given the diversity of theoretical predictions, experiments are needed to inform this question, but few are available and most were carried out with longitudinal gradients.

Damgaard and co-workers (Damgaard et al. 2003; Damgaard et al. 1997) used a recirculating flume in which flow rate was held fixed but the longitudinal gradient varied. In the first set, fine sand (median diameter $d_{50} = 208$ μm or $\phi = 2.3$) was injected into the base of the flume with a piston controlled such that the injection rate exactly matched removal as bedload. Bedload fluxes derived from sand pickup rate are shown in Figure 2A for three different sets of experiments made with different flow rates. They show the expected increasing flux with increasing down-slope gradient, with an abrupt increase towards the sand angle of repose.

In their second study at higher flow rates (Damgaard et al. 2003), sediment-trap measurements (representing largely bedload) show a systematic variation with bed gradient. Ripples, however, formed on the bed, significantly affecting suspended sediment fluxes because of sand thrown into suspension at ripple crests. Hence, suspended sediment fluxes (gray-filled circles in Figure 2B) are varied and peak at -5° rather than at maximum gradient. The effect of ripples was complex because different ripple morphologies formed at different bed gradient, leading to varied suspension. Their flow speed of 0.35 m/s measured 13 cm above bed is comparable with maximum speeds measured near the shelf edge (Huthnance et al. 2002). Considering that ripples are observed around the shelf edge (Yorath et al. 1979), varied suspension could be a further complication, but Figure 2B nevertheless shows a general tendency for fluxes to increase with increasing down-gradient.

Further experiments (Fernandez Luque and van Beek 1976; Smart 1984) documented effects of longitudinal gradients. In Smart's experiment, flux increased with $S^{0.6}$ (where S is bed gradient) but included some suspended transport. Fernandez Luque and van Beek's experiments recorded an effect of gradient on the threshold of motion, and bedload flux correlated moderately well with excess stress corrected for the gradient effect.

Japanese experimental results with transverse gradients (Hasegawa 1981; Yamasaka et al. 1987) shown in Sekine and Parker (1992) are reproduced in Figure 2C (those of Hasegawa were carried out in water whereas those of Yamasaka et al. were carried out in air). Based on Equation DA10, the trend in the data should reveal the value of the exponent m. The main group of data were claimed (Sekine and Parker 1992) to be consistent with $m = 0.25$, which implies $q_n \propto t_b^{1.25}$ ($q_n \propto u^{2.5}$).

The theoretical and experimental results therefore suggest that bedload flux should be affected by bed gradient, with a component down-gradient, $Q_b = -K|\nabla\mathbf{H}|$. Although not well constrained, the Japanese data suggest that the dependence of K on current speed u probably lies between $K \propto u^2$ and $K \propto u^3$. If threshold effects are also considered, a variation $K \propto (u-u_o)^2$ could produce morphological results similar to $K \propto u^3$. As the published current meter data do not allow threshold effects to be fully accounted for, we have compared current variations with morphology assuming that K lies between $K \propto u^2$ and $K \propto u^3$, but threshold effects may need to be considered in more accurate interpretations.

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DA2: CURRENT-METER STATION LIST AND VALUES

Station ID	Latitude (degrees N)	Longitude (degrees W)	Depth (m)	Measurement altitude (m)	Duration (days)	Mean speed (cm/s)	Source	Symbol in Fig. 4
USA Atlantic margin								
SEEP-5/1983[1]	39.805	70.9217	1250	10	197	6.1	Csanady et al. (1988)	square (solid)
SEEP-5/1984[1]	39.805	70.9217	1250	10	173	5.6	Csanady et al. (1988)	square (solid)
NASACS-SA[1]	40.08	68.5583	485	7	148	8.	Butman et al. (1988)	diamond (solid)
NASACS-SE[1]	39.8967	70.0617	491	7	141	9.9	Butman et al. (1988)	diamond (solid)
NASACS-SE[1]	39.8967	70.0617	504	7	244	6.2	Butman et al. (1988)	diamond (solid)
NASACS-SF[1]	39.9617	70.015	202	7	98	13.8	Butman et al. (1988)	diamond (solid)
NASACS-SF[1]	39.9617	70.015	204	7	129	11.6	Butman et al. (1988)	diamond (solid)
NASACS-SA[1]	40.08	68.5583	479	6	149	7.9	Csanady et al. (1988)	square (solid)
NASACS-SF[1]	39.9617	70.015	204	7	143	11.6	Csanady et al. (1988)	square (solid)
NASACS-SG[1]	39.8083	70.0833	1150	7	246	8.4	Butman et al. (1988)	diamond (solid)
NASACS-SH[1]	39.842	70.0283	1220	7	171	5.4	Csanady et al. (1988)	square (solid)
MASAR-A[1]	39.704	73.063	225	5	116	12.4	Csanady et al. (1988)	square (solid)
LCI[1]	40.38	67.5517	250	5	148	9.81	Butman et al. (1988)	diamond (solid)
LCI[1]	40.38	67.5517	250	5	146	8.86	Butman et al. (1988)	diamond (solid)
LCI[1]	40.38	67.5517	247	5	125	9.93	Butman et al. (1988)	diamond (solid)
LCI[1]	40.38	67.5517	249	6	157	11.95	Butman et al. (1988)	diamond (solid)
Atlantic data also plotted but not used in the speed-depth regression:								
A[1]	39.48	72.98	59	1.5-2.0	11.92	10.7	McClennen (1973)	plus
C[1]	39.06	74.0767	30	1.5-2.0	8.88	12.	McClennen (1973)	plus
D[1]	38.8467	73.16	74	1.5-2.0	10.79	11.7	McClennen (1973)	plus
B[1]	39.25	72.5667	143	1.5-2.0	10.13	17.8	McClennen (1973)	plus
T[1]	40.1817	69.9717	100	7	117	10.95	Butman et al. (1988)	plus
SEEP-2/1983[1]	40.2417	70.9167	125	5	146	10.3	Csanady et al. (1988)	plus
Aa[1]	39.3917	72.9917	60	-	36	8	Butman (1979)	plus
Ab[1]	39.3917	72.9917	65	-	23	7	Butman	plus

							(1979)	
B[1]	38.7083	73.6333	60	-	68	9	Butman (1979)	plus
Ca[1]	38.5417	73.5083	80	-	36	13	Butman (1979)	plus
Cb[1]	38.5417	73.5083	87	-	14	6	Butman (1979)	plus
MASAR-F[1]	36.836	74.576	1005	100	210	6.5	Csanady et al. (1988)	square (open)
NASACS-SA[1]	40.08	68.5583	475	100	337	10.4	Csanady et al. (1988)	square (open)
NASACS-SD[1]	40.284	67.730	485	100	328	11.4	Csanady et al. (1988)	square (open)
NASACS-SE[1]	39.8967	70.0617	500	100	350	9.5	Csanady et al. (1988)	square (open)
NASACS-SE[1]	39.8967	70.0617	510	100	142	9.9	Csanady et al. (1988)	square (open)
NASACS-SE[1]	39.8967	70.0617	504	100	245	8.8	Csanady et al. (1988)	square (open)
LCI[1]	40.38	67.5517	250	55	148	9.46	Butman et al. (1988)	diamond (open)
LCI[1]	40.38	67.5517	249	50	157	8.97	Butman et al. (1988)	diamond (open)
Iberian Atlantic margin:								
stablecd110[1]	42.6783	9.50833	202	0.21-0.91	18	13	Huthnance et al. (2002)	square (solid)
stablecd114[1]	42.6667	9.50833	200	0.21-0.91	18.5	10	Huthnance et al. (2002)	square (solid)
rcm03192[3]	40.999	9.475	1293	99	730.6	10.62	Malena (WOCE) (2)	star (solid)
rcm02590[3]	42.218	9.509	1338	100	387	3.95	Malena (WOCE) (2)	star (solid)
b0530652[3]	41.315	8.9867	84.0	2.	75.	8.87	OMEX-II(1)	diamond (solid)
b0530720[3]	41.3183	8.9817	84	3.	139.	9.35	OMEX-II(1)	diamond (solid)
-[2]	42.33	9.38	853	0.05-0.4	-	6.6	Thomsen et al. (2002)	circle (solid)
-[2]	42.33	9.38	194	0.05-0.4	-	18.75	Thomsen et al. (2002)	circle (solid)
-[2]	42.33	9.38	97	0.05-0.4	-	27	Thomsen et al. (2002)	circle (solid)
-[2]	43.05	9.52	155	0.05-0.4	-	31.4	Thomsen et al. (2002)	circle (solid)
OTHER W EUROPE ATLANTIC MARGIN								
Hebrides[1]	56.45	9.05	210	0.21-0.91	27	12	Huthnance et al. (2002)	circle (open)
Hebrides[1]	56.46	9.04667	204	0.21-0.91	15	14	Huthnance et al. (2002)	circle (open)
Goban Spur[1]	49.3917	11.6667	879	0.21-0.91	10	11	Huthnance et al. (2002)	circle (open)
Chapelle Bank[1]	47.47	6.54667	388	0.21-0.91	8	17	Huthnance et al. (2002)	circle (open)

"Station ID" is the identifier used in the data sources.

(1) Ocean Margin Exchange (OMEX) Project, OMEX-II Project data set (CD-ROM, British Oceanographic Data Centre, Liverpool, UK (NERC)). Original data attributed to Instituto Hidrografico, Portugal.
(2) Data calculated from current velocity data supplied as part of the WOCE compilation, attributed to ACM27 and ACM28 ("Morena" experiment). Originally collected by scientists of the University of Lisbon and the Spanish Institute of Oceanography. Related publication: Fiúza, A. F. G., Hamann, M., Ambar, I., Díaz del Río, G., González, N., and Cabanas, J. M., 1998, Water masses and their circulation off western Iberia during May 1993: Deep-Sea Res. I., v. 45, p. 1127-1160.

Superscripts in Station ID column refer to the type of measuring instrument:
^{1}Mechanical: rotor(s) with direction vane.
^{2}Acoustic doppler current profiler.
^{3}Unknown.

Notes on accuracy

As the above results include measurements made with different current meters, the relative performance of the different instruments could be a cause for concern. In particular, the mean current measured with mechanical current meters is known to be affected by superimposed oscillating currents, such as from surface waves, because of the finite response time of the rotors and direction vane. The instruments measure the wave current when it adds to the mean current but under-record when the wave-current reverses, leading to a net bias. In one study (Beardsley 1987), when wave currents had root-mean-squared amplitudes equal to half the mean current velocity, the measured current was in error by 10% and greater for larger oscillating current amplitudes. These measurements will also not be particularly representative of the instantaneous current speed (due to both wave and mean current) in such situations. These issues are not expected to affect the arguments in this paper greatly because we are concerned with variations below 150 m where high-frequency oscillating currents tend not to penetrate.

Potential errors in the calibration formulae that have been used at Woods Hole Oceanographic Institution to relate rotor speed to current speed have been noted by Lentz et al. (1995). Their comparison of a mechanical current meter with an acoustic current meter suggested that the error increased linearly to around 2.5 cm/s at a speed of 30 cm/s. If it had affected the results of Csanady et al. (1988) and Butman et al. (1988) (these papers unfortunately lack calibration details), the values in Fig. 5A will have been exaggerated by around 1 cm/s at 10 cm/s mean speed and less at depth. This will have steepened the graph slightly but not sufficiently to affect the arguments in the paper.

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DA3. BOTTOM CURRENTS FROM SHIP ADCP DATA OFF IBERIA

Data were collected with a hull-mounted acoustic Doppler current profiler (ADCP) during a cruise off the Iberian margin shown in Figure DA3 (Huthnance 1997). The ship traversed the margin repeatedly, crossing the 200 m contour at 20 random times with respect to the tidal cycle. Although the 10 day period of the cruise is short compared with some oceanographic variations, these data provide a further indication of how the magnitude of seabed oscillations varies with water depth, in relatively mild June conditions (Huthnance 1997).

The ADCP vector currents were corrected for ship motion and converted to current vector magnitudes. Bathymetry along the ship tracks was derived by interpolating (Smith and Wessel 1990) 50 m contours of the General Bathymetric Chart of the Ocean (GEBCO) above 200 m depth (IOC, IHO and BODC, 2003) along with multibeam bathymetry of the slope (NERC 2001). To derive near-bed current speeds, while allowing for bathymetry inaccuracy due to incomplete coverage, we selected all current data within 60 m of the seabed. The solid line in Figure 5B then represents the median average near-bed current as a function of the local water depth and dotted lines show the inter-quartile range of current speeds.

Because the ADCP averages current data over typically 5 minutes, the trend in Figure 5B excludes effects of long-period surface waves (swell), affecting the shallower depths. Northerly winds during the cruise favored upwelling, and interestingly this should have produced currents decreasing inversely with water depths, similar to those observed. The Ekman surface current is expected to have a volumetric transport flux relative to its underlying water that relates to the wind stress rather than depth, but its squeezing into shallow water leads to stronger currents. If the upwelling flux is Q, the bottom-layer onshore current is u_b, total depth H, and upper water layer thickness h_u, then the absolute upper-layer flux offshore is $Q - u_b h_u$. The lower-layer flux onshore is $u_b(H - h_u)$. These two transport fluxes cancel at the coast, so combining the above suggests $u_b = Q/H$,, i.e. flow in the lower layer inversely proportional to water depth, similar to that observed in Figure 5B. This analysis applies over shelf depths 100 to 200 m, but friction affects flow in shallower depths and upwelling can occur mid-depth in deeper water.

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Figure DA3. Path of RRS Charles Darwin during cruise 105 while operating its hull-mounted ADCP.