Threshold of erosion of submarine bedrock landscapes by tidal currents

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This is a pre-publication version of an article published in Earth Surface Processes and
Landforms in 2012: <http://dx.doi.org/10.1002/esp.3347>

Keywords: abrasion, quarrying, erosion, tidal currents.

Running head: Tidal scouring of bedrock

Abstract.

Since sea level stabilised 7000 y BP, shelf seas experiencing semi-diurnal tides will have been affected by streaming four times per day. If tidal erosion of bedrock were even only marginally efficient, the ~10 million streamings since then should have left geomorphological imprints. We examine high-resolution multibeam sonar data from three areas with extreme tidal currents. The Minas Passage (Bay of Fundy) experiencing 8-knot surface tidal currents was surveyed in 2007 with a multibeam sonar. In an area near to transverse dunes, which are evidence for bedload transport, the data show local overhanging surfaces near to the sediment-rock contact, potentially created by abrasion by saltating particles. However, they are uncommon. In the Straits of Messina, where surface currents reach 10 knots, surveying revealed ridges lying oblique to the flow that are not obviously broken into separate outcrops by erosion. In the Bristol Channel, UK, sonar data collected where currents reach 3.4 knots at 1.5 m above the bed reveal outcrops of limestone with superimposed sand dunes, but only minor rounding of blocks.

Holocene tidal currents have apparently been generally ineffective at eroding bedrock. We examine this issue further by compiling extreme tidal streams around the UK and from them estimate shear stresses, representing a macro-tidal environment where peak surface currents reach 9.7 knots. Those data are compared with shear stresses in mountainous rivers where long-term rates of erosion are comparable with tectonic uplift rates and are thus geomorphologically significant. Whereas river stresses reach 10^2 - 10^3 Pa, the largest tidal stresses are generally 10^1 and only rarely approach 10^2 Pa, too small for quarrying to operate generally. However, the vertical faces in the Minas Passage may

represent the onset of abrasion. Given this limited evidence for abrasion, we explore conditions in the geological past for tides that may have locally eroded bedrock.

Introduction

Little published evidence exists, as far as we are aware, that modern tidal currents are capable of geomorphologically important bedrock erosion. The physiography of continental shelves, for example, is largely explainable as a planation surface created by coastal erosion (Hamblin *et al.* 1992; Pantin 1991; Trenhaile 2001; Wood 1976). Some channels on the NW European continental shelf (a macro-tidal environment) lie suspiciously parallel to strong tidal currents, but most of them can be explained as paleo-river or flood valleys (Dingwall 1975; Gupta *et al.* 2007; Smith 1985) or depressions of tectonic origin maintained by river and tidal scour (Lericolais *et al.* 1996) rather than by bedrock erosion. One exception is St Catherine's Deep, a trough developed immediately south of the Isle of Wight (UK) in Lower Cretaceous Gault and possibly (Upper and Lower) Greensand (Hamblin *et al.* 1992). As the trough is a closed-contour depression, it cannot have been created by fluvial erosion when sea level was lower. The Gault clays are described by Hamblin *et al.* as "soft Lower Cretaceous strata", which may explain their excavation by tidal currents. Elsewhere in the English Channel, the strata have become accentuated either by differential erosion by tides or by surf during transgression (Belderson *et al.* 1972; Coggan *et al.* 2009; Collier *et al.* 2006). Belderson *et al.* (1988) described bed gravel and cobbles bare of fauna for decimetres above the floor of furrows in the English Channel, potentially abraded by saltating particles. Harris *et al.* (2005) documented closed-contour valleys extending to 220 m below modern sea level on the

continental shelf extending between NE Australia and Papua New Guinea. The degree of induration of the incised carbonate substrate is unclear though most likely a limestone.

Tidal modelling showed that bottom water shear stresses in the valleys probably exceeded 10 Pa during transgressions when shelf-lagoons became hydraulically connected with the ocean. Other work reviewed by Harris *et al.* (2005) suggests that tidally scoured depressions are not uncommon around Australia.

In contrast, some other observations argue against erosion of bedrock being effective. For example, a fluvial drainage pattern in the English Channel created when the area was exposed during sea level low-stand conditions (Gupta *et al.* 2007) appears unmodified by later submarine erosion despite strong tidal currents in the area (Pingree and Griffiths 1979). In examples of tidal channels now tectonically uplifted above sea level, they have been reported as having only sedimentary floors. Darwin (1841) recognised moribund tidal channels above sea level in southernmost South America. Colella and D'Alessandro (1988) reported marine sediments, which they interpreted as deposited in a paleo-strait crossing Calabria in the Plio-Pleistocene and subsequently uplifted tectonically. A hiatus in the paleochannel suggests that sediments stopped depositing when the channel was 300 m deep, as a result of tidal current action, but evidently tidal scour was not strong enough to erode even poorly consolidated sediments, *let alone* bedrock, to maintain the channel as a seaway. In contrast with fluvial erosion, tidal erosion occurs in the absence of a base level. Progressive erosion can potentially lead to a negative feedback with the formative currents, whereby widening channels cause less constriction of tidal flows leading to weaker stresses. More limited extents or incidences of tidally eroded channels might therefore be expected compared with fluvial channels.

Multibeam echo-sounders generate “soundings” by calculating the range and angle associated with seabed echoes. When combined with data from motion sensors and positioning data of the vessel or vehicle carrying the sonar, the result is a point cloud of elevations similar to those produced by LiDAR systems. Traditionally, we have worked with these data in simple map form, in which depth is a single-valued function of (x,y). Processing algorithms are therefore typically designed to filter the data on the expectation of only a single depth value within each grid cell (Caress and Chayes, 1996). However, with the higher resolution systems we can now record three-dimensional topology of the seabed (illustrated in the present study by overhanging outcrops) so depth can have multiple values. As with the LiDAR datasets, sonars also present challenges of interpretation because of their potentially large volumes. For example, a 1-m resolution dataset collected over a 100 km coastline implies a grid with a dimension of 10^5 , much larger than the 10^3 -pixel dimension of a computer screen by two orders of magnitude. These issues are mostly avoided in the present study by selective choice of data but they will present difficulties of processing and interpretation in future studies.

In this study, we examine recently acquired multibeam sonar datasets from areas of extreme tidal currents (Figure 1) to see if there is any evidence for erosion of bedrock. A dataset of shear stresses beneath typically strong tidal currents is developed and compared with a dataset for bedrock-eroding rivers. Such rivers are responsible for the long-term denudation of landscapes and many of those selected here exist in tectonically uplifting areas. Although it is unclear whether such rivers are locally eroding at rates that exactly equal rates of tectonic uplift (Whipple 2001), bedrock rivers are anticipated to achieve on average a balance between erosion and uplift over long time-scales (this

assumes that modern rates are not systematically different from Pleistocene averages). If the processes eroding bedrock beneath tidal currents were the same as those in rivers, tidal current stresses would need to be comparable. This leads to a discussion of whether more extreme tidal current stresses could have existed at times in the geological past sufficient to have been geomorphologically important.

Case studies

The following multibeam sonar datasets are from areas of extreme tidal currents, where evidence for erosion is most likely to be found if it occurs. In assessing the morphological evidence, we were guided by the types of erosional geomorphological features observed in desert environments and riverbeds. For example, in deserts, saltating particles abrade rock surfaces up to the height limit of particle motion and can ultimately produce sculpted rock formations (yardangs) with under-mined bases (Anderson 1986). Although the greater viscosity of water compared with air could lead to differences in abrasion between the two environments, Carling (in press.) has described similar yardang-like features developed in soft strata (marls) from the inter-tidal zone in the Severn Estuary. In extreme flows in rivers, the highest energy impacts of particles can occur on the parts of boulders facing downstream, due to detachment of particles in the recirculating flow behind them, leading to indentations there (Hancock *et al.* 1998). In quarrying or plucking, particles or blocks are slid laterally by shear stresses applied by the flow or lifted away from the bed by pressure differences arising from the Bernoulli effect. The geometry of extracted blocks is commonly controlled by the pre-existing

joints in the rock, resulting in sharply defined edges in the remaining bedrock (Hancock *et al.* 1998).

Depth precision and resolution vary between the sonar datasets but largely increases in proportion to water depth (Schmitt *et al.*, 2008). The acoustic resolution is typically determined by the beam widths of the sonars, which are commonly 1° or wider. In shallow marine surveys (10s of metres depth), the position data (GPS, for example) often most limits resolution. However, artifacts due to acoustic noise and motion sensor and water column sound velocity errors can also degrade the data. We mention some of these below where they appear but an interested reader can see examples in Hughes Clark *et al.* (1996).

Minas Passage, Bay of Fundy, Canada

The Bay of Fundy experiences Earth's most extreme tidal ranges as the natural resonance frequency of water in the bay is close to the lunar-forced M_2 tidal frequency (Garrett 1972). The Minas Passage connects the Minas Basin with the broader Bay of Fundy (Figure 2c). As a result of the extreme tidal range and the restricted cross-section of the Passage, surface currents within the Passage reach 3.28 m/s or 6.4 knots (Karsten *et al.* 2008). The local navigation chart shows a peak surface current of 8 knots on the south side of the passage (Canadian Hydrographic Service 2003, Chart 4010). These currents probably became established at around 3400 yBP, when the Minas Basin experienced a rapid increase in tidal range, which is speculated to have been caused by breaching of a barrier of glacial sediments at the Minas Passage (Shaw *et al.*, 2010).

Figure 2b shows a shaded-relief image of bathymetry data collected with a Simrad EM710 multibeam sonar in 2007 within the Passage (located in Figure 2c). Data acquisition is described in Todd *et al.* (2011). The sonar beam widths of $0.5^{\circ} \times 1.0^{\circ}$ suggest a theoretical acoustic resolution of up to 0.9 m at 100 m depth. The image in Figure 2b was created after gridding the data at 0.0001° (11 m in latitude by 8 m in longitude). In its north side, the image shows a series of elongated features trending WNW-ESE crossed by faults. The relatively uniformly layered appearance of the seabed morphology in the northern part of Figure 2b suggests that these may be part of the sequence of lacustrine limestones and siltstones, chert and minor fluvial sandstone overlying basalts that have been identified around the adjacent shores (Olsen and Schlische, 1990; Gibbons *et al.* 1996). Along X-X' in Figure 2c, Olsen and Schlische (1990) showed a cross-section with a syncline of basaltic lava flows beneath the Passage and underlying the sedimentary sequence apparently exposed here. These rocks probably form a moderately hard substrate.

A 20-m thick banner bank lies within the sediment filled base of the passage. It displays a train of large sediment waves, demonstrating that bedload sediment transport has been occurring here. A seabed photograph collected over these bedforms during the short slack tide (Todd *et al.*, 2010) is shown in Figure 2d. It and an adjacent photograph not shown reveal a gravely seabed of diverse lithologies and varied roughness, with grain size varying from a minimum of around 5 mm to a maximum larger than 5 cm (evaluated visually with the aid of a 1-cm grid over the image). The clean surfaces of those particles are consistent with active modern transport. Depth contours are plotted at a close 1 m depth interval in Figure 2a so that bunched contours highlight the steeper rock surfaces

around the edge of the sediment body. Figure 3 shows the sounding data from the five lines marked A-E on Figure 2a (these data have been merged with motion sensor and position data but are otherwise unprocessed and not gridded to preserve original features). Where highlighted by dashed circles, they show nearly vertical or locally overhanging scarps of up to a few metres high immediately adjacent to the sediment-covered floor of the Passage.

The deepest depth measurements along these profiles vary from 134 m (A), 124 m (B), 125 m (C) to 107 m (D, E). They are unlikely to have lain near or above sea level during glacial conditions because of glacio-isostatic depression of this area. The highest high water (marine limit) in the Chignecto Bay immediately north of the Minas Passage was at 48 m above modern mean sea level at 13,500 C¹⁴ yBP (Rampton *et al.* 1984 cited in Amos and Zaitlin 1984-1985). A significant deepening at around 13,000 C¹⁴ years is also suggested by models of relative sea level obtained by combining all data from eastern Canada (Shaw *et al.* 2002). However, erosion by glacial ice or a sub-glacial fluvial (tunnel valley) origin of the scarps in Figure 3 cannot be entirely ruled out. The reconstructed ice palaeogeography suggests ice converged through the Minas Passage as it flowed westwards from central Nova Scotia (Shaw *et al.* 2006) suggesting some role of ice-bed erosion in creating the large-scale morphology. Nevertheless, the location of the scarps in the vicinity of the margins of the modern sediment deposits in the base of the Passage where they are covered with gravel waves suggests strongly that erosion originated from saltating bedload.

Bristol Channel off Nash Point

The tidal range in the Bristol Channel estuary is third only to the Bay of Fundy and Ungava Bay (northern Canada). However, as the estuary narrows landwards, the tidal prism and tidal currents are consequently smaller than those in the Minas Passage. The values on Figure 4 represent tidal current measurements in knots from UK Admiralty Charts, which are generally less than 5 knots. Harris and Collins (1988) recorded tidal currents at “CM2” in Figure 4 with an Aanderaa meter deployed 1.5 m above the bed over a period encompassing one spring tidal phase. The maximum current speed recorded was 1.75 m/s (3.4 knots). The current velocity plot in Figure 5 was formed by reducing those current measurements to 1 m above the bed (Schmitt 2006) as described in the figure caption.

Various parts of the South Wales coast were surveyed in 2002 with a high-resolution Reson 8101 multibeam sonar (Schmitt *et al.* 2008). Figure 6 shows a small portion of those data collected close to Nash Point as located in Figure 4. Bourne and Willemse (2001) described the adjacent rock platform exposed at low tide as containing inter-bedded Liassic limestones and shales, with individual limestone beds rarely thicker than 1 m, dipping southwards at less than 5°. Erosion on the platform occurs by surf removing the thinner shale beds, a process that undermines the harder limestone beds and allows blocks of the latter to collapse (Trenhaile 1972). Based on the loss of segments of Iron Age cliff-top forts, Trenhaile (1972) estimated average cliff recession rates of 24-61 cm/year for the last 2500 years, which are much faster than modern rates. Those rates and the evidence of other sea level terraces from earlier sea level high-stands suggested to Trenhaile that the platforms developed over a longer period than just the present sea level high-stand.

Although the Reson sonar's beam width implies an acoustic resolution in 10 m depths of only 20-30 cm, Schmitt (2006) estimated from crossing swaths of data that the practical resolution is around 1 m due to precision of the differential GPS system used. The image in Figure 6 was derived from a 1-m grid of the data. It shows the edge of the modern coastal rock platform (marked "Platform"), beyond which the seabed drops to below 10 m (the values along the right of Figure 6 are depths below Chart Datum (Lowest Astronomical Tide)). Within that lower area, a small field of dunes can be observed with superimposed ripples, extending across the left of the map. Grain size data of nearby samples suggest a median diameter (d_0) of 0.3 mm (Schmitt 2006). Where not covered by sediment, however, the prominent features of the map are the strata. They form steps typically <1 m high that are remarkably sharply defined.

That sharp definition is consistent with the beds having been quarried, with blocks removed along joints. Aside from some rounded outcrop edges, however, there is relatively little evidence for significant abrasion despite the abundant sand present and its mobility implied by the bedforms. As we have no information on the age of erosion, quarrying may have occurred when this area was affected by surf at earlier times when sea level was lower, such as during the post-glacial transgression. Furthermore, the area may have also been subjected to strong currents associated with storm surges superimposed on the tide, such as during the 1607 AD flooding event. Horsburgh and Horritt's (2006) inundation model of that event shows water level elevated by 2.3 m. If that level is assigned to the approximately 10^3 km^2 of water area east of Nash Point and developed in around 10 hours, it implies an additional discharge of $0.23 \text{ km}^3/\text{hour}$ or an additional current of 0.2 m/s distributed through the 20 km width and typical 20 m depth

of the Bristol Channel south of Nash Point. The additional 0.2 m/s elevates the currents somewhat but, based on the kinds of stress arguments developed later, was probably not sufficiently to explain the bedrock quarrying observed. Quarrying probably instead occurred earlier during the transgression by wave erosion.

To estimate how long the strata have been exposed to potential abrasion after the quarrying, Haslett *et al.* (1998) estimated that sea level has been rising steadily in the eastern Bristol Channel throughout the late Holocene and by 0.41-0.82 mm/yr from 3000 to 1770 y BP. This latter rate implies that the present bedrock depths have not changed by more than a few metres within the past few thousand years. The strata should therefore have been exposed to tidal currents for about this time.

Straits of Messina

The data from this Messina area (Figure 7) are unfortunately less complete compared with the other areas but nevertheless hint at structures that are compatible with the other areas. Figure 8 shows a single swath of bathymetry data collected through the Straits with a Reson 8160 multibeam sonar on *R/V Urania*. The data are shown gridded at 1 m, a smaller interval than the acoustic resolution (beam width of 2.7 m at 100 m range). Continuous bedrock was originally assigned to this area by Selli *et al.* (1979) on the basis of profiler, sidescan sonar data and sampling, although their sonar data showed crossing linear features of unclear origin. The more recently collected multibeam sonar data reveal that bedrock forms isolated outcrops. A sample recovered by dredging was described by Selli as sandstone and cemented conglomerate. In contrast with the surface morphology, seismic reflection data collected across the Straits 10 km to the south show a

shallow dipping acoustic stratigraphy (Argnani *et al.*, 2009; Goswami, 2012). Given their linear morphologies in Selli's sidescan and these multibeam sonar data but flat morphology of strata in the seismic data, we speculate they could be neptunian dykes inverted because of their resistance to erosion. Most likely the intervening sediments are gravels corresponding to the samples of Selli *et al.* (1979). Oblique trends visible in Figure 8 ("B") lying parallel to the tidal currents are most probably gravel ridges and furrows similar to those described by Belderson *et al.* (1988) in the English Channel.

The 2.7 m data spatial resolution is unfortunately insufficient to allow us to observe whether the bases of outcrops are undermined by abrasion. D. Hodgetts (pers. comm., 2010) has analysed the full sounding data with software designed to deal with irregularly spaced three-dimensional outcrops (Hodgetts 2009) revealing no evidence of basal notches. Nevertheless, the 1-m contoured bathymetry in Figure 8 shows that the margins of the outcrops are steep and may be accentuated by tidal scour. However, although some breaks can be observed in Figure 8, scour has not eroded corridors through the outcrops, and the data do not show yardang-like streamlined features aligned with the flow, such as described by Carling (in press).

Bed shear stress estimates

The previous sections showed that the evidence for erosion by tidal currents is generally weak. We examine this further here by estimating shear stresses under tidal currents. In both of the following sections, bed shear stresses τ_0 opposing flow are calculated rather than the skin friction experienced by sediment grains or rock surfaces exposed on the bed. In the case of rivers, researchers typically estimate

$$\tau_o = \rho g R \sin \alpha \quad (1)$$

where ρ is the water density, g is the acceleration due to gravity, R is the hydraulic radius (approximately the flow depth for wide rivers) and α is the bed slope, usually calculated over a reach scale of ~ 100 m. Equation (1) simply represents the stresses required to balance the weight of the flow, neglecting flow acceleration if carried out over the scale of a river's reach (of order 100 m distances)). In the case of tidal current modelling, the flow stress is usually calculated from

$$\tau_o = C_d \rho |u|^2 \quad (2)$$

where C_d is a friction factor and $|u|$ is depth-averaged flow speed (Taylor 1919). By matching the results of tidal models to tide gauge records, C_d has been found to be typically 0.0025 (dimensionless). As the flow is resisted by pressure variations caused by large-scale roughness from the bed, i.e., form drag from bedforms, shear stress computed from (2) is also not a skin friction but a net shear stress opposing flow. The issue of compatibility between the two stresses calculated in equations (1) and (2) is discussed later.

Surface waves can create strong shear stresses near the bed because their short periods compared with other flows result in them developing only a thin turbulent boundary layer. A similar equation to (2) can be used to estimate the maximum bed shear stresses when only waves are present:

$$\tau_o = 0.5 C_w \rho |u_w|^2 \quad (3)$$

where $|u_w|$ is the maximum wave orbital speed near the bed and C_w is a bed friction factor for waves (Grant and Madsen 1982). Experimental results suggest that C_w equals 0.24 (dimensionless) when the wave-induced excursion distance of water near the bed is greater than the characteristic wavelength of bed topography (e.g., ripple spacing) but varies when excursion distances are smaller (Bagnold 1946; Carstens *et al.* 1969; Grant and Madsen 1982). We adopt a value of 0.1 as a lower bound for small excursion distances and 0.24 as an upper bound. Because wave motions influence the effective bed roughness "felt" by a steady current (Vincent 1986), stresses due to waves and tidal currents cannot simply be added, but the non-linear interaction of waves and currents should lead to a net stress that is larger than their sum.

Rivers

Representative shear stresses were compiled for various mountain rivers (Table 1) and summarized as histograms in Figure 9d (sources are given in the figure caption). These are bedrock, alluvial or mixed bedrock-alluvial rivers. Although bedrock may not presently be fully exposed in the alluvial or mixed bedrock-alluvial rivers, which are more common than bedrock rivers, alluvium is expected to be strongly mobilised during rare extreme floods leading to bed erosion in many of them. The South Fork Hoh River data shown in Figure 9d represent reaches separately classified into pool-riffle, cascade and step-pool and bedrock, illustrating the larger shear stresses typical of the latter

(Montgomery and Buffington 1997). Shear stress also can be expected to vary greatly within individual reaches given their variably rugged topography and alternation between sub-critical and super-critical flow. Hence, the reach-averaged values in Figure 9d are smaller than the peak stresses in the streams available for erosion.

Whereas many of these rivers lie in the Mediterranean, four others experience more extreme wet and dry climates. The different rivers also occur in landscapes that are tectonically uplifting by varied mechanisms, for example, related to convergence in the Apennines (Torto, Torrente L'Apa) and Olympic Mountains (South Fork Hoh) and to extension in the Gulf of Corinth (Xerias) and Negev (Nahal Paran), while Sandy Creek gorge lies within a declining orogen (a no-longer tectonically uplifting mountain range). The history and rates of tectonic uplift can be expected to vary amongst these areas and be fairly representative of the range of tectonic forcing that any erosion process would need to overcome to have a significant geomorphological effect.

The calculated shear stresses correspond to various return periods or particular dated floods. Most were obtained from reach-scale gradients and hydraulic radii estimated by surveying evidence of extreme flood levels in the form of trim-lines of vegetation contrasts and scour marks (Cowie *et al.* 2008) and driftwood and slackwater deposits (Wohl *et al.* 1994). In some cases (Jansen 2006; Wohl *et al.* 1994), the along-channel gradient was computed using step-backwater hydraulic calculations (e.g., Webber 1971). Also suggesting that these data are generally representative, further step-backwater hydraulic calculations for rare events (of 100 y return period) suggest shear stresses of 200 and 700 Pa for the Colorado Paleau Cache la Poudre and Rocky Mountain Dry Fork Coyote Narrows streams, respectively (Miller and Cluer 1998) and 140 and 170 Pa for

two reaches of the Rio Chagres, Panama (Wohl and Springer 2005). The histogram for the Zhuokouxi River represents the full distribution of river discharge at one site (hence the generally lower stress values), whereas the bars represent extreme conditions at various points along that river (Barbour *et al.* 2009).

The larger stresses in Figure 9d commonly occur in knickzones or knickpoints where erosion is enhanced (Cowie *et al.* 2008; Jansen 2006; Wohl *et al.* 1994). Even more extreme conditions have been reported, for example, 2000 Pa in the LiWu River, Taiwan during super-typhoons where rock mass loss from abrasion is measurable (Hartshorn *et al.* 2002) and 10^4 Pa in megafloods such as during glacial lake outbursts (Baker and Kale 1998). These are atypical however and appear not required for erosion in most rivers.

Tidal currents

According to C. Jones of the UK Hydrographic Office (pers. comm., 2008), tidal currents reproduced on the UK Admiralty charts were mostly measured before the use of acoustic and other current meters became widespread. The movement of a 5-10 m pole consisting of a spar or tube, weighted to float vertically, would be observed relative to a monitoring ship over 25-50 hours during calm weather of spring tides. Those data would be reduced using a non-harmonic semi-graphical method that removes any effect of residual currents.

Maximum (spring) tidal currents were taken from charts at the locations shown in Figure 10. To provide a database of extreme currents, only the larger current speeds were selected, so the data are deliberately biased towards extreme values. Shear stress opposing flow was calculated using equation (2) using the near-surface currents for the

depth-averaged flow current $|\mathbf{u}|$ (based on our experience, depth-averaged currents are typically 80% of surface currents so this will modestly exaggerate stresses). The grey-filled histogram in Figure 9c was derived from stresses calculated with $C_d=0.0025$.

Howarth and Souza (2005) carried out a study of turbulent stresses in shelf seas and suggested that the parameter C_d should vary according to:

$$C_d = 0.001(0.59 + 0.78M_2) \quad (4)$$

where M_2 is the maximum M_2 tidal current speed (m/s). As the M_2 component dominates the tides around the NW European shelf, C_d was computed using the spring current speeds for M_2 at each site in Figure 10. The resulting histogram of shear stresses is shown by the dashed line in Figure 9c, differing from the histogram computed with a uniform C_d mainly in the lower values, the upper stresses being only weakly modified. The values are broadly in line with more direct estimates of shear stress from turbulence measurements, for example, up to 4 Pa in a narrow tidal channel (Rippeth *et al.* 2002) and up to 5-10 Pa in a tidal stream off North Wales (Howarth and Souza 2005).

Also shown in Figure 9c (solid bars) are shear stress values for the Straits of Messina computed using equation (2) for the maximum current speed in each graph of Mosetti (1988) representing >67 days of measurements dispersed across the Straits. The two rightmost values represent currents of 3 and 5 m/s that reportedly occur in the Straits under extreme conditions (Mosetti 1988; Santoro *et al.* 2002). In Figure 9c, the short open bar at 5 Pa annotated “N” represents the estimated peak shear stress from the largest current recorded off Nash Point in Figure 5 and short open bar at 42 Pa annotated “M”

represents an 8 knot current reportedly occurring in the Minas Passage. All these values were computed using $C_d=0.0025$.

Wave-induced bed stresses

Pantin (1991) estimated near-bed maximum oscillating current speeds under waves generated during a 50-year storm (upper 0.005 time percentile), using statistical model predictions of storm wave period and height from a Department of Energy (1983) report on extreme sea conditions. Bed stresses were computed using equation (3) following the procedure in Soulsby (1997). The shear stress experienced by the bed depends on the scale of bed roughness, which affects the friction factor C_w (" f_w " in Soulsby (1997)). Given a median grain size d_{50} , the Nikuradse roughness $k_s=2.5d_{50}$ (Nielsen 1992) and bed roughness length $z_0=k_s/30$. The friction experienced by the bed then also depends on the maximum excursion distance of water caused by wave motion $A=U_w T/2\pi$.

Soulsby *et al.* (1993) compiled measurements of A and z_0 across a broad range to derive a simple expression for the friction factor C_w :

$$C_w = 1.39 \left(\frac{z_0}{A} \right)^{0.52} \quad (5).$$

In this study, we computed individual values of A for each locality in water depths of 20 to 100 m where there were values of U_w and T (Pantin 1991), recalculated after limiting wave heights to 40% of local water depth to account for wave breaking. Shear stresses were then computed from equation (3) using the derived C_w values. The procedure was carried out for 0.3 and 1.0 mm grain sizes corresponding to medium and coarse-very coarse sand, respectively. The 0.3 mm sand is typical of the Bristol Channel area described earlier (Schmitt, 2006), for example. Natural fracture surfaces vary

enormously in fine-scale relief, for example, 0.1 mm in chert (Poon *et al.*, 1992), but greater than 1 mm observed visually by us in coarse grained rocks such as granite in the field. One mm was chosen as an upper d_{50} value to give the reader a sense of the sensitivity to this parameter. If larger surface roughness is assumed, the calculated shear stresses are correspondingly greater.

The results shown in Figure 9b suggest that a bed with roughness corresponding to sand grades should experience shear stresses above 10 Pa in a 50-year storm. The values are larger than maximum shear stresses derived by Neill *et al.* (2009) for a single year, which are all below 10 Pa. Stresses above 10 Pa should strongly mobilise and most likely suspend sand based on the threshold of motion data in Figure 9a, as observed frequently at sea.

Form drag

Stresses in the two environments (rivers and seas) were calculated somewhat differently so their compatibility should be addressed. The shear resistance experienced by a wide flowing water body can be expressed as the sum of the following components (Soulsby 1997):

$$\tau_0 = \tau_{0s} + \tau_{0f} + \tau_{0t} \quad (6)$$

where τ_{0s} represents the skin friction experienced by the bed, τ_{0f} represents form drag (pressure variations caused by flow over bed topography) and τ_{0t} represents the momentum transfer when entraining sediment grains into the flow and their impacts with

the bed. The shear stresses opposing friction calculated using equations (1) and (2) represent the total shear resistance, whereas the shear stress relevant to plucking of bed particles and sediment and hence abrasion is the skin friction τ_{0s} . Therefore, in order to judge whether differences between the stresses in Figure 9 reflect differences in skin friction, differences in τ_{0f} and τ_0 between the two environments need to be assessed also.

The momentum transfer stress τ_{0t} should be larger where stresses are larger overall because of faster sediment entrainment rates. This would affect flooding rivers most, implying that the difference in skin friction between the two environments could be smaller than implied by the shear stress difference in Figure 9 if form drag differences could be neglected.

Hughes *et al.* (2008) studied current measurements in Torres Strait, Australia, which suggested that C_d locally exceeds 0.007 because of the effect of steep bedforms (sand dunes). They suggested that as much as 65% of the drag coefficient there was due to form drag. However, much lower values for C_d , closer to those required of the larger scale tidal modelling, have been reported in studies of other shelf seas (Huntley *et al.* 1993; Williams 1995) where bedforms are less steep. Reynolds stress measurements in a tidal stream in North Wales suggested smaller friction coefficients of $C_d=0.0016-0.0019$ (Howarth and Souza 2005) and $C_d=0.0025$ (Rippeth *et al.* 2002).

Kean and Smith (2006) developed an analytical solution for the form drag resistance due to a series of Gaussian-shaped roughness elements in the flow boundary. They suggested that the flow resistance should equal:

$$\tau_{0f} = \frac{1}{2} \rho C_d \frac{H}{\lambda} u_{ref}^2 \quad (7)$$

where H is the protruding height of a roughness element, λ is the length-scale of the wake produced by that element and u_{ref} is a reference velocity. The bedform effective friction coefficient C_d was found empirically to equal $1.79\exp(-0.77\sigma/H)$, where σ is the Gaussian width parameter of the roughness. The inverse of the ratio σ/H is analogous to the steepness index of sedimentary bedforms (Flemming 2000). Because of the effect of the exponential (and as λ should increase with H so H/λ varies less), the form drag will increase strongly with increasing steepness of the bed protuberances. To the extent that there is no reason to expect a rocky substrate that has been eroded along joints to have a different roughness in the two environments, the steepness and hence form drag should be comparable. Therefore, broad differences in shear stress calculated using equations (1) and (2) should represent differences in skin friction also, although individual values in either environment would be strongly affected by the non-linear dependence of form drag on bed element steepness.

Bed stress observations

River stresses exceeding 10^2 Pa are common for the floods of longer return periods represented in Figure 9d (aside from the Ro Gllego data, which represent only a 25-year event, and the Zhuokouxi data, which represent the full distribution of discharge). The stresses generated by tidal currents (Figure 9c) are significantly smaller than 10^2 Pa. Some extreme values, such as those computed for the Messina flows, lie within 10^1 and 10^2 Pa, but more typically stresses are less than 10^1 Pa. If a shelf area is exposed to surface waves, the bed can experience typically up to 10^1 Pa of bed shear stress (Figure

9b). The combined stress due to waves and tidal currents is complex but should equal at least their sum (Grant and Madsen 1982) so their combined stresses may be greater than 10^1 Pa under the more extreme conditions.

Threshold shear stress for erosion can be speculated on using the grain size and other data for the Minas Passage site. According to Figure 9a, the minimum grain size of 5 mm observed in Figure 2d in the Minas Passage would be suspended if bed shear stresses exceed roughly 10^2 Pa, a value larger than the 40-50 Pa maximum estimated to occur there (“M” in Figure 9c). The 5 mm grain size and quartz density implies a maximum Shield’s stress of 0.06, whereas the ratio (25) of shear stress of 50 Pa to threshold stress of movement in Figure 9 implies a transport stage (ratio of shear velocity to threshold shear velocity or u_* / u_{*0}) of roughly 5. Abbott and Francis (1977) carried out saltation-suspension experiments on similar sized particles. Extrapolating trends from their Figure 4 suggests that particles could be suspended and reach altitudes of metre-scale at this transport stage. This, nevertheless, would be an extreme occurrence based on the data and the polymict nature of the gravel may inhibit suspension of the finer grains due to their entrapment between larger grains. Undermining of outcrops to several metres may have instead occurred by the migration of the bedforms causing the bed altitude to vary over time.

Discussion

In the modern shelf seas, strong tidal currents tend to coincide with bedload parting zones due to divergence of residual currents at those localities, so they are commonly relatively depleted of sediments (Harris *et al.* 1995), such as illustrated by Figure 7b. Our

sonar datasets demonstrate, however, that these areas are not completely devoid of clastic particles and the data contain evidence for bedload movements. Nevertheless, although some seas may be locally supplied by sediments from adjacent rivers (e.g., the Straits of Messina perhaps), the texture and flux of sediments across the bed will likely differ from those in flooding mountain rivers supplied with sediment by adjacent mass-wasting hillslopes.

Given the common occurrence of $>10^2$ Pa stress in Figure 9d, 10^2 Pa appears to represent the shear stress above which geomorphologically meaningful erosion occurs in rivers, which tends to occur by quarrying and plucking. A stress threshold of 10^2 Pa has also been inferred from detailed observations. According to Snyder *et al.* (2003), a rare flood in Fall Creek, New York visibly moved blocks of up to 0.3 m thickness. The shear stresses within the reach studied were estimated to have been 100-200 Pa, which Snyder *et al.* suggested were just sufficient to cause bedrock plucking. The smaller shear stresses under tidal currents are therefore consistent with the lack of evidence for bedrock quarrying even in the extreme tidal stream of the Minas Passage, aside from the Bristol Channel area probably eroded by surf. Tidal current erosion by quarrying and plucking mechanisms probably only occurs in friable shale or similar incompetent strata, such as the Gault forming St Catherine's Deep (Hamblin *et al.* 1992).

Depending on their components of form drag, some stresses appear sufficient to mobilise particles up to cobble size (Figure 9a), so tidal erosion of bedrock by the abrasion mechanism is not ruled out. This could explain the near-vertical outcrops in the Minas Passage (Figure 3c) and those in limestone mentioned by Harris *et al.* (2005). Coupled with the lack of abrasional sculpting at the other two sites (Bristol Channel,

Straits of Messina), the effective threshold for abrasion probably occurs in the range 10^1 - 10^2 Pa, depending on particle size, bed erodibility and other factors as in rivers (Hancock *et al.* 1998; Whipple *et al.* 2000). Experiments in which rock surfaces were eroded by particles in an abrasion mill (Sklar and Dietrich 2001) have shown that erosion rate is strongly dependent on the bedload particle size. In areas of bare rock superimposed with bedload, the greater tidal shear stresses capable of mobilising larger grain sizes (Figure 9a) should therefore lead to increased erosion rates. In models of stream-bed erosion by bedload (Hancock *et al.* 1998; Lamb *et al.* 2008; Sklar and Dietrich 2004; Whipple *et al.* 2000), erosion rate increases in proportion to the flux of particles impacting the bedrock surface and to the kinetic energy of individual particles, both effects that increase with flow shear stress. All these factors taken together are consistent with a strongly non-linear rate of abrasion with shear stress or an effective threshold.

We speculate, therefore, that, although there is relatively little evidence of on-going erosion of bedrock beneath tidal currents in the modern oceans, the more extreme bed stresses are not greatly less than those needed for abrasion. In the case of the Minas Passage, abrasion may be occurring today. This leads us to question whether conditions might have existed in the geological past locally to elevate stresses to levels at which bedrock erosion occurred.

Egbert *et al.* (2004) developed a numerical model of global tides for conditions during the Last Glacial Maximum, which provides some insight. It suggested that tidal height variations in many parts of the deep water Atlantic Ocean were twice as large then as they are at present, principally because the dissipation of tidal energy that occurs presently on continental shelves would have been reduced with the sea retreated to the

edges of the shelves. In the Pacific Ocean, the effect was much less pronounced due to narrower shelves around that basin. Therefore, during the Pleistocene glacial maxima, tidal height changes driving currents could generally have been larger in some areas, in particular in the Atlantic. Unfortunately, the modelling does not have sufficient resolution to identify sites of elevated shear stress (which can be localised in shelf valleys (Harris *et al.* 2005)), so we instead speculate on how more extreme shear stresses might have arisen in a more general sense (i.e., favourable bathymetry, etc). In the modern oceans, strong tidal flows occur in three types of localities: in estuaries where the water body resonance frequency is close to the M_2 frequency, in areas where topographic constrictions cause the flow to accelerate locally (e.g., around headlands) and in tidal straits where tidal height varies nearly 180° out of phase either side of the straits, which then form amphidromic points. Examples of the latter include the Straits of Messina (Mosetti 1988; Santoro *et al.* 2002), the Torres Strait, northern Australia (Hughes *et al.* 2008) and Cook Strait, New Zealand (Proctor and Carter 1989). Future improvements to the modelling with more accurate paleo-bathymetry datasets and other sea conditions may help to reveal whether similar locations emerged during glacial maxima.

Based on hydraulic relations for channel flow (Bridge 2003; Campbell *et al.* 1998), a doubling of the surface water gradient down a tidal channel resulting from a doubling of tidal range at its ends could lead to a doubling of shear stresses if all other factors were equal. Over the continental shelves, tidal currents could also have been larger. For a shelf of width W , length L and tidal range R , the tidal prism would have a volume WLR . On average that volume needs to be supplied by a peak current U of depth H across the open boundary length l in half a tidal period T . Thus, $U = \pi WLR / (lHT)$. During times of

lowered sea level, H at the shelf break would have been less and, if R and W were not greatly different, the peak current U would have been stronger. If U were twice as large, for example, the shear stress would have been four times as large. Similar arguments can be made if flow became concentrated in tidal channels, as both l and H could have been smaller, leading to even larger U and shear stress. A numerical tidal model of Uehara *et al.* (2006) for the NW European shelf during the post-glacial rise in sea level predicts that shear stresses over the southern Celtic Sea and westerly English Channel exceeded 10 Pa due to a combination of the above effects. Numerical modelling of the Cook Strait tides (Proctor and Carter 1989) suggests that tidal current stresses reach 6.5 Pa in the narrowest area at the present day and 10.5 Pa when sea level was lowered at 17 ka, when the strait was closed, creating an estuary.

Conclusions

Bed shear stresses beneath the strongest tidal currents around the UK and in other tidal streams investigated here are generally an order of magnitude smaller than those in mountain rivers in flood. The tidal shear stresses, which are largely less than 10^1 Pa, are too small to erode bedrock by quarrying or plucking mechanisms, which usually require $>10^2$ Pa in rivers. Nevertheless, the shear stresses may not be so much smaller than those required for abrasion. This would seem consistent with the evidence from the morphological data. Whereas there is little evidence for significant abrasion off Nash Point and in the Straits of Messina, there are candidates for abrasive undermining of outcrops in the Minas Passage. We speculate that greater stresses associated with enhanced deep-water tidal height variations during glacial low-stands, in particular in the

Atlantic Ocean, may have led to sufficiently strong currents for bedrock abrasion to occur locally.

Acknowledgements

Mr C. Jones of the UK Hydrographic Office kindly provided the data on the tidal streams presented on Admiralty charts. Ellen Wohl and Patience Cowie are thanked for supplying river data for Figure 9. Stuart Jones provided further information clarifying aspects of his river studies. The data in Figures 2 and 3 were collected as part of a joint project of the Geological Survey of Canada, Canadian Hydrographic Service and University of New Brunswick (UNB) and provided by Doug Cartwright of the Ocean Mapping Group (UNB). The data are reproduced here with the kind permission of John Hughes Clarke. Andrea Argnani helpfully pointed out the literature on tidal channels in Calabria and is thanked for access to the data in Figure 8. The multibeam data in Figure 5 were collected with funding from the HEFCW, Royal Society, Hanson Aggregates, RMC aggregates and United Marine Dredging. We thank also John Howarth for clarifying aspects of his published turbulent stress measurements. John Shaw made helpful suggestions concerning the palaeogeography and ice history. George Lericolais and Sanjeev Gupta provided details on the English Channel. Colin Stark suggested the uniformity of shear stresses of actively eroding bedrock rivers in flood. We also thank Peter Harris (who pointed out the effect of a lack of base level), Brian MacLean and an anonymous reviewer for helpful comments, which significantly improved this manuscript. Since we submitted this article, Neil Kenyon has mentioned evidence for further tidally eroded outcrops of soft lithologies in the English Channel. Figures were

created with the "GMT" software system (Wessel and Smith 1991). JMH was supported by the Natural Environment Research Council. The tidal data in Figure 5 were provided by the British Oceanographic Data Centre of the UK Natural Environment Research Council. ESS Contribution number 20110298.

Figure 1. Locations of the areas shown in Figures 2, 4 and 8.

Figure 2. Multibeam echo-sounder and other data collected in the Minas Passage, entrance to the Minas Basin, Bay of Fundy, Canada. (a) Image of the multibeam data (pale grey background) overlain with contours derived from two swaths of the multibeam data gridded at 2 m resolution and contoured with a 1-m vertical interval (the 100 m contour is shown in bold). Bunching of contours highlights near-vertical rock outcrops. White lines labelled A-E correspond to the profiles shown in Figure 3. (b) Shaded relief map of the gridded dataset with the artificial sun north of the figure. Circled point on the sediment waves locates the photo in (d). (c) Location of (a) and (b) within the Minas Passage. X-X' is approximate line of cross-section of Olsen and Schlische (1990). (d) Photo taken at location marked in (b). Circled points are laser marks spaced 10 cm apart.

Figure 3. Profiles of soundings along the lines marked in Figure 2. Lower-case 's' marks an acoustic shadow. Dashed circles highlight possible overhangs or vertical outcrops

discussed in the text. Profile A has been plotted with the scale to left. Others are plotted with same vertical scale but each offset vertically in increments of 33.3 m.

Figure 4. The Bristol Channel and Severn Estuary are shown along with tidal stream measurements reported on UK Admiralty charts (peak spring tides). Inset (solid black rectangle) locates the multibeam dataset in Figure 6 against the coastline of South Wales off Nash Point. Also shown in the inset are the locations of current meter CM2 used to collect the data in Figure 5 and the extent of the coastal rock platform exposed at low tide (dashed) from UK Admiralty charts.

Figure 5. Circular scatter plot of tidal current data at 100 cm above bottom (U_{100}) from an Aanderaa Savonius-type rotor current meter deployed near Nash Point collected over one

neap-spring half-cycle from 18/08/1983 to 3/09/83. The original data were collected by University of Wales Swansea scientists (Harris and Collins 1988) and were provided by the British Oceanographic Data Centre. The data were collected 1.5 m above the bed in 10 m water depth. They have been reduced to 1 m above bed assuming a seabed roughness parameter of 0.5 cm and a standard logarithmic law of the wall velocity profile (Schmitt 2006). A critical threshold speed for the local typical 0.3 mm grain size ($U_{cr,100}=0.19 \text{ m.s}^{-1}$) is indicated by the inner dashed circle. This mooring was located at CM2 in the inset to Figure 4.

Figure 6. Example of bedrock exposed in high-resolution bathymetry data collected off Nash Point in the northern Bristol Channel, UK (Schmitt 2006) shown as a shaded relief map (arrow marked "shading" shows the illumination direction). The projection is Universal Transverse Mercator (coordinates are UTM distances in m). Artefacts in the data marked "A" include (lower left) small depth offsets between survey lines due to

errors in corrections for tidal height and (upper right) an across-track crenulation caused by a vessel heave sensor error (Schmitt *et al.* 2008).

Figure 7. (a) Shaded-relief topographic map of the Straits of Messina, Italy based the Shuttle Radar Topography Mission data (Ryan *et al.* 2009). (b) Seabed types based on map of Selli *et al.* (1979) reproduced in Santoro *et al.* (2002). Annotation represents (sG,

G) sandy gravel and gravel, (gS, S) coarse, very coarse and gravely sand, (S) fine and medium sand and (mS) clayey sand and loam. White circle locates the swath in Figure 8.

Figure 8. Bedrock exposures in the Straits of Messina in a single swath of multibeam sonar data gridded at 1 m (located in Figure 7b). Top panel shows a shaded relief image computed with artificial illumination from N315°E (upper-left). Lower panel shows enlarged section of the data contoured at 1 m intervals. Annotation ‘A’ marks scan-parallel artifact and ‘B’ subtle trends crossing the swath at a low angle interpreted as gravel ridges and troughs.

Figure 9. Comparing threshold of motion data with estimates of shear stress below tidal and river currents. (a) Threshold of motion of cohesionless sediment. Curves are approximate boundaries between modes of sediment movement (Bridge 2003). Filled circles are threshold data from a stream (Carling 1983). Plus symbols and unfilled circles are data from Lane (1955) and Neill (1967) in Miller *et al.* (1977). Wentworth grain size descriptors are given along the left.

(b) Shear stresses derived from estimates of near-bed maximum orbital current speed beneath waves generated during 50-year return-period storms (Pantin 1991). Solid and dashed lines represent data calculated using equation (3) with $C_w=0.1$ and 0.24 , respectively.

(c) Shear stresses derived from tidal streams on UK Hydrographic Office charts (Figure 10). The grey filled and dashed histograms show values computed with constant $C_d=0.0025$ and with C_d varying as described in Howarth and Souza (2005), respectively. Also shown by the solid vertical bars are stresses in the Straits of Messina, computed using equation (2) from the maximum current speed in each graph of Mosetti (1988) representing >67 days of measurements dispersed across the straits. The two rightmost values represent currents of 3 and 5 m/s that reportedly occur in the Straits under extreme conditions (Mosetti 1988; Santoro *et al.* 2002). The two short open bars represent 5 Pa (marked “N”) estimated for the maximum current in Figure 5 off Nash Point and 42 Pa (marked “M”) for a reported 8 knot current in the Minas Passage (Canadian Hydrographic Service 2003, Chart 4010).

(d) Reach-scale shear stresses corresponding mostly to peak flow conditions derived from (1) Montgomery and Buffington (1997), (2) Wohl *et al.* (1994), (3) Cowie *et al.* (2008),

(4) Barbour *et al.* (2009), (5) Jones (2009) and (6) Jansen (2006, Fig. 8). The South Fork Hoh data are sub-divided into different reach types. The Zhuokouxi river histogram represents all shear stresses for a single location rather than for peak flow conditions. Also shown over that histogram (vertical bars) are shear stresses during peak flow at various sites along the same river.

Figure 10. Locations of selected tidal current measurements recorded on UK Admiralty charts and converted here to shear stresses (key top-right). The two lines in the English Channel locate deeps mentioned in the text.

Table 1. Sources of river shear stress data.

River (source)	Return period or date (y BP)	Location	Climate
South Fork Hoh (1)		Olympic Mountains, Washington State USA	Wet temperate
Nahal Paran (2)	350	Negev Desert, Israel	Arid
Torto (3)	1-10	Apennines, Italy	Mediterranean
Torrente L'Apa (3)	1-10	Apennines, Italy	Mediterranean
Xerias (3)	1-10	Gulf of Corinth, Greece	Mediterranean
Zhuokouxi (4)	Full distribution	Taiwan	Tropical (typhoon- prone)
R�o G�llego (5)	25	Spain	Mediterranean
Sandy Creek (6)	300	SE central Australia	Arid

Sources of data given in brackets are as in the caption to Figure 9.

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