Seascape evolution on clastic continental shelves and slopes


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ABSTRACT

The morphology of clastic continental margins directly reflects their formative processes. These include interactions between plate movements and isostasy, which establish the characteristic stair-step shape of margins. Other factors are thermal and loading-induced subsidence, compaction and faulting/folding, which create and/or destroy accommodation space for sediment supplied by rivers and glaciers. These processes are primary controls on margin size and shape. Rivers and glaciers can also directly sculpt the margin surface when it is subaerially exposed by sea-level lowstands. Otherwise, they deposit their sediment load at or near the shoreline. Whether this deposition builds a delta depends on sea level and the energy of the ocean waves and currents. Delta formation will be prevented when sea level is rising faster than sediment supply can build the shoreline. Vigorous wave and current activity can slow or even arrest subaerial delta development by moving sediments seaward to form a subaqueous delta. This sediment movement is accomplished in part by wave-supported sediment gravity flows. Over the continental slope, turbidity currents are driven by gravity and, in combination with slides, cut submarine canyons and gullies. However, turbidity currents also deposit sediment across the continental slope. The average angle of continental slopes (~4°) lies near the threshold angle above which turbidity currents will erode the seafloor and below which they will deposit their sediment load. Therefore, turbidity currents may help regulate the dip of the continental slope. Internal waves exert a maximum shear on the continental-slope surface at about the same angle, and may be another controlling factor.

Keywords Seascape, plate tectonics, deltas, clinoforms, submarine canyons, internal waves, turbidity currents.
INTRODUCTION

From bathymetry to seascape evolution

A great deal of what is known about the evolution of continental shelves and slopes is based on bathymetry. Shelf and slope bathymetry provide key information about underlying structures (Pratson & Haxby, 1997), the distributions of sediment facies (Posamentier et al., 1988), past geological events (Shepard, 1934), and processes that have and may continue to affect a continental margin (Driscoll et al., 2000). Advances in seafloor mapping (e.g. global positioning system, satellite altimetry, multibeam bathymetry and sidescan sonar imagery) and extensive surveys (e.g. the US Geological Survey and NOAA surveys of the US Exclusive Economic Zone) achieved over the past several decades are especially noteworthy in this regard. These accomplishments are providing a perspective of seafloor morphology that is finally starting to compare in detail and scope with that which has long been available for subaerial and even some extraterrestrial planetary landscapes (Fig. 1) (Pratson & Edwards, 1996). Correspondingly, marine geologists are increasingly taking a geomorphologist’s viewpoint of the seafloor, trying to understand seafloor evolution in the same terms as landscape evolution by surface processes acting over geological time.

Inherent in the process–product viewpoint of strata formation is the recognition that seascape

Fig. 1 Computer-generated image of the landscape and seascape along the western US continental margin. Bathymetric data of the seafloor are approaching the detail and scope of topographic data for land. The image looks northward over the northern half of California toward Oregon. Elevations are exaggerated fivefold relative to distances, with the straight-line distance between San Francisco and Point Reyes being ~70 km. Land elevations are from the USGS 30-arcsecond digital elevation model (DEM). Seafloor elevations extending from the shoreline out to the black line are from bathymetry collected by the National Oceanic and Atmospheric Administration (NOAA) and gridded by Pratson & Haxby (1996). Bathymetry beyond the black line are from the 2-min bathymetric grid created by Smith & Sandwell (1997). Box (a) is the approximate location of the bathymetry shown in Fig. 16A, while box (b) is the location of the bathymetry shown in Fig. 19A. (Image modified from Pratson & Haxby, 1997.)
do not passively evolve with the build-up of strata, but strongly influence where deposition and erosion occur. Landscapes exert a comparable influence in subaerial settings. They also become shelves when sea level rises, while shelves and even the uppermost slope can become part of the landscape when sea level falls. Thus, in some ways it is not surprising that landscapes and shelf-slope seascapes often look similar. In other ways though, it is surprising, because parts of the shelf and just about the entire slope are purely of marine origin. While the shelf and slope often share the dominantly erosional appearance of landscapes, these provinces contain the thickest accumulations of sediment in the world (Kennett, 1982).

Scope of paper

This paper is about how the overall shapes and dimensions of shelves and slopes are created by the major processes that operate across clastic continental margins in temperate settings, particularly those bordering the continental USA. Not addressed by this paper are the formation of shelves and slopes along carbonate (e.g. Florida, Bahama and north-eastern Australian) or tropical (e.g. Papua New Guinea and Central America) margins, or those that occur in epeiric seas such as the Adriatic. Margins that have been glaciated (e.g. Alaska, Iceland and Antarctica) and those deformed by salt or mud tectonics are also largely omitted (e.g. the USA Gulf of Mexico, Brazilian and Angolan), although impacts of these processes on shelf and slope evolution are considered.

Similarly, this paper does not attempt to review the formation of shelf and slope morphologies across all spatial scales, especially leaving fine scales for other papers in this volume (e.g. the Humboldt ‘slide’ discussed in Lee et al. (this volume, pp. 213–274). The full range of features is too broad and the literature addressing them too substantial to accomplish a complete synthesis here. Instead, the goal of the paper is simply to summarize the basic ways that ubiquitous processes ranging from plate tectonics to internal waves contribute to the large-scale evolution of shelves and slopes over geological time. The impacts are illustrated using conceptual, often geometrically based models that greatly simplify, but also hopefully clarify, the processes that contribute to shelf and slope development.

BACKGROUND

Physiographical definitions

The continental shelf and slope are components of continental margins. Slopes are the transition zones between the continents and ocean basins, and merge with continental rises in a seaward direction (Fig. 2). As with the rise, the shelf and slope are sufficiently distinct from other regions and from one another to be delineated largely on the basis of water depth and average seafloor gradient. The continental shelf extends seaward from the shoreline to the shelf break, where the seafloor gradient begins to increase markedly (Fig. 2). On average, the shelf break occurs ~80 km offshore at a water depth ~130 m (Kennett, 1982), but the range about these averages is significant. The shelf break is as shallow as several tens of metres seaward of the Ganges-Brahmaputra River, and as deep as 350 m at points along the Antarctic margin. Shelf width varies from being non-existent seaward of the Markham River in Papua New Guinea, to more than 300 km wide in the East China Sea. In all cases though, the shelf is a relatively flat surface characterized by a very gentle overall seafloor gradient of ~0.05° (1:1000; Kennett, 1982). As will be discussed, there are distinct features on the shelf with steeper local gradients. The most important of these is the shoreface, the sloping descent from the shoreline to water depths in the order of 10 m that is formed by waves as they impinge upon the shore.

The continental slope extends from the shelf break to water depths of 1.5–3.5 km, although depths are even greater for slopes that descend into oceanic trenches. Seafloor gradients on the slope are similarly variable, ranging from < 1° to > 25°, with the overall average being ~4° (Kennett, 1982). Consequently, the relief and steepness of continental slopes rival those of many mountain ranges.

Historical interest in and importance of the continental shelf and slope

Interest in the bathymetry of shelves and slopes primarily began with concern for charting shoals upon which ships could run aground (Vogt & Tucholke, 1986). Starting at least as early as 85 BC, bathymetric measurements were collected using a weight attached to a measured line. This technique
was used through the 1800s when the submarine telegraph was introduced, and bathymetric surveys were needed to map routes for laying the cables (Vogt & Tucholke, 1986).

Since the advent of the echosounder in the early 1900s, knowledge of the continental shelf and slope has expanded in both breadth and detail. This in turn has led to the incorporation of shelf and slope physiography in the formulation of many revolutionary geological concepts. For example, prior to the mid-1900s, shelf and slope morphology was used to support theories for the existence of continents and ocean basins, and the formation of mountains (Kay, 1951). Before any sediment-transport observations were made, studies of seafloor morphology led to the realization that many features on shelves and slopes had to be shaped by submarine processes and could not be explained solely by subaerial processes acting during periods of lowered sea level (Daly, 1936; Stetson, 1936; Stetson & Smith, 1938). Shelf and slope morphology has gone on to be explained in such unifying theories as plate tectonics (Dewey, 1969; Bird & Dewey, 1970) and the underpinnings of sequence stratigraphy (Vail & Mitchum, 1977). It has also become important for:

1. managing offshore resources, including deepwater (i.e. > 300 m) oil and gas reserves, and marine fisheries (Cook & Carleton, 2000);  
2. identifying regions subject to marine geohazards such as submarine landslides that can trigger tsunamis (Coulter & Migliaccio, 1966) and damage offshore infrastructure (Bea, 1971);  
3. mapping potential pathways and sinks for contaminants introduced offshore (Poppe & Polloni, 1998; Lee et al., 2003);  
4. reconstructing sea-level change over the Holocene, and its implications for future changes resulting from global warming (Fedje & Josenhans, 2000).

**PROCESSES GOVERNING SHELF WIDTH AND SLOPE RELIEF**

**Plate tectonics and the stair-step shape of continental margins**

The ‘stair-step’ shape of continental margins is first and foremost the result of plate tectonics. All margins begin where lithospheric plates (that make up the rigid outer shell of the Earth) interact with one another as they move over the plastically deforming asthenosphere in the upper mantle (Dewey,
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1972; Fig. 3). These interactions have generated the granitic and andesitic crust of the continents, and the thinner, denser, basaltic crust that floors the oceans (Fig. 3). Just by virtue of this difference in thickness and density, granitic and basaltic crust form a step where they meet along the edges of continents. However, the shape of this step and its subsequent evolution are strongly influenced by the regional tectonic environment.

Broadly speaking, plate tectonics leads to two types of continental margins. Passive margins are those that ride ‘passively’ within the interior of a lithospheric plate (Kennett, 1982) (Fig. 3A). They first form when the plate and continental crust are rifted. During this process, the continental crust is stretched and fractured (i–ii, Fig. 3A). Basaltic magma derived from partial melting of the underlying asthenosphere rises to fill the fractures. The magma supply eventually becomes focused along a mid-ocean ridge that follows the axis of the old rift (iii, Fig. 3A). Along this ridge, new basaltic crust forms to floor an expanding and deepening ocean basin. Thus the underlying structure of passive margins, created during the rifting process, is a faulted, several kilometre descent from unextended continental lithosphere down to oceanic lithosphere (iv, Fig. 3A).

Active margins lie along boundaries between plates and are shaken by the plate movements (Kennett, 1982). Active margins form along:

1 convergent plate boundaries where continental lithosphere overrides or subducts oceanic lithosphere (Fig. 3B);
2 transform plate boundaries where continental and oceanic lithosphere slide past one another.

The latter type of margin is relatively rare, so here the focus will be on convergent margins. Beneath these, partial melting of the subducted oceanic lithosphere generates andesitic to granitic magmas that penetrate and add to the overriding continental
lithosphere (Fig. 3B). The convergent margin itself, however, tends to be located seaward of the igneous terrain over an accretionary prism. This is a faulted wedge of sediments that the overriding plate has bulldozed off the downgoing plate (Fig. 3B).

First-order effects of thermal subsidence and tectonic uplift

Sedimentary shelves and slopes along convergent margins are not only built of debris scraped off the downgoing plate, but are partly comprised of sediments eroded off the continent. The continents are also the primary source of clastic sediments to shelves and slopes along passive margins. In general, the greater the accumulation of sediments along a margin, the farther the shelf reaches out into the ocean basin and the greater the water depth to which the slope extends (Jervey, 1988). An equally important control on shelf width and slope relief, however, is the accommodation space available along a continental margin for storing the sediments (Van Wagoner et al., 1988). Abrupt water depth increases with distance offshore cause narrow shelves, and allow for taller slopes to be built from a given volume of sediments (Fig. 4A).

The first-order controls on accommodation space along continental margins are thermal subsidence and tectonic uplift (Hays & Pitman, 1973; Pitman, 1978). Thermal subsidence is geomorphically most important along passive margins, and arises from the cooling of the adjacent oceanic lithosphere following its formation at the mid-ocean ridge (iii–iv, Fig. 3A). As it cools, the oceanic lithosphere thickens and becomes denser, causing it to subside at a rate that tapers off according to $1/\alpha$ (Parsons & Sclater, 1977). The rate decreases from 40–100 m

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**Fig. 4** Crustal controls on continental-margin morphology. (A) The steeper the descent of transitional crust across a continental margin, the more space there is for accommodating sediments eroded from the continent, and thus the narrower the resulting shelf and the greater the relief of the slope (i). Slower subsidence leads to broader shelves and smaller slopes (ii). This relationship holds true to zero subsidence, after which uplift reverses the trend. The faster the uplift, the quicker the slope and shelf are destroyed as they are raised up and incorporated into the continent, leading to narrower shelves and deeper slopes. (Modified from Reynolds et al., 1991.) (B) As a result of uplift, shelf and slope sediments along active margins are recycled, and add to the sediments being eroded from the continent and bulldozed off the subducting plate. (Modified from Kulm & Fowler, 1974.)
Myr$^{-1}$ along a young margin, such as the Gulf of Lions (Burrus, 1989), to $<3$ m Myr$^{-1}$ along an old margin, such as that off New Jersey (Steckler et al., 1999). The rate also decreases landward across the margin to zero at the hinge line or boundary between extended and unextended continental lithosphere at the edge of the initial rift (Pitman, 1978; dashed line in ii, Fig. 3A).

The changing rate of thermal subsidence with time and with distance offshore adds complexity to the evolution of shelf width and slope relief. If the rate of subsidence on the outer shelf exceeds the rate at which sediments are supplied to the margin, more space is created than can be filled by the sediments, and sediment accumulation shifts landward (Reynolds et al., 1991; Fig. 4A). This scenario is most likely to occur along a young, fast-subsiding margin, and when it does occur, the shelf narrows and builds upward or aggrades over time. The relief of the slope increases as well. As a margin ages, however, subsidence rates decrease. When the maximum subsidence rate on the shelf is exceeded by the sediment supply rate, infilling occurs faster than new space is created. Under these conditions, sediment accumulation is forced out into the deeper waters over the slope, which builds seaward or progrades into deeper water. As a result, the shelf widens and the slope lengthens (Reynolds et al., 1991; Fig. 4A).

Along convergent margins, tectonic uplift is the most important control on accommodation space. The uplift stems from the overriding continental lithosphere being thickened by:

1. buckling and folding as it ploughs over the downgoing oceanic lithosphere;
2. intrusions and extrusions of magma generated by the subduction;
3. the build-up of accretionary sediments ‘bulldozed’ off the subducting lithosphere (Allen & Allen, 2005).

Importantly, uplift can also occur along passive margins when the upper lithosphere is stretched less than the lower lithosphere during rifting (Allen & Allen, 2005) (ii, Fig. 3A). The non-uniform stretching exposes the trailing edge of the upper lithosphere to greater heating by the asthenosphere, leading to thermally driven uplift of the margin. This uplift tends to be geologically short lived, however (Fig. 3A), and so it simply sets back the development of accommodation space by thermal subsidence.

The tectonically driven uplift along convergent margins rotates the shelf and slope strata upwards (Kulm & Fowler, 1974; Kulm et al., 1975) (Fig. 4B). The uplift destroys accommodation space on the shelf and forces the greatest sediment accumulation to occur on the slope, which is where tectonic accretion from below is also greatest. The end result is both aggradation and progradation of the margin. However, the uplift also causes the slope strata to eventually become shelf strata and ultimately part of the continental land mass. Here the strata are once again subject to erosion and form a source of sediments for new shelf and slope strata. As a result, there is a continuous recycling of sediments along convergent margins, as well as the addition of materials by subduction and igneous rock formation (Fig. 4B).

**Second-order effects of isostasy, compaction and faulting**

If the New Jersey margin is representative, ~40% of the space that accommodates sediments on mature passive margins is created by thermal subsidence (Steckler et al., 1988), about 40% is produced by isostatic subsidence and 20% by compaction. Consequently, the subsidence generated by isostatic and compaction processes has a significant influence on shelf width and slope relief, but it has a less uniform influence. This is because thermal subsidence is an externally driven process that affects the entire breadth of a continental margin. Isostatic subsidence and compaction are more localized and occur in response to sediment loading.

As sediments accumulate on continental margins, they cause isostatic subsidence of the lithosphere and they compact. **Isostatic subsidence** is the sinking of the lithosphere lower into the asthenosphere to reach a new level of buoyant equilibrium under the sediment load. The amount of subsidence depends on the mass of the load and the rigidity of the lithosphere (Watts & Ryan, 1976). If the lithosphere has no rigidity, it will only subside directly beneath the load to a level at which the buoyancy force produced by the displaced asthenosphere equals the mass of the load. This case is known as **Airy isostasy** (Fig. 5A). However, if the lithosphere...
has rigidity, then it behaves like an elastic beam. The load is distributed over a broader region such that beneath the load the subsidence is less but it extends laterally beyond the load. This is known as flexural isostasy (Fig. 5B & C).

Sediment compaction is similar to Airy isostasy in that it occurs only directly beneath the region of sediment loading (compare Fig. 5A & D). This loading forces the sediment grains to assume tighter packing, resulting in a reduction of the intervening pore space. Compaction is greatest within about the first kilometre of the seabed, and is completed by a burial depth of ~4–5 km (Sclater & Christie, 1980). Over this range, porosities of sand, silt and clay decrease exponentially from ~0.5, 0.6 and 0.6 to ~0.3, 0.2 and 0.2, respectively (Bahr et al., 2001).

Isostatic subsidence commonly lags the loading, due to the sluggish, viscous response of the displaced asthenosphere, and the thermal readjustment of the lithosphere as it moves to its new level (Le Meur & Hindmarsh, 2000; Watts & Zhong, 2000). Measurements of post-glacial rebound indicate that following loading/unloading, subsidence/rebound occurs relatively rapidly for a period of 10 kyr (Watts et al., 1982). Over the next 300–400 kyr, the isostatic readjustment decays exponentially to zero assuming the load remains unchanged.

Compaction will also lag loading, if the loading rate exceeds the rate at which water between the

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**Fig. 5** Isostatic subsidence due to sediment loading and compaction. (A–C) The amount of isostatic subsidence that occurs along a passive margin, as sediments accumulate, depends on the rigidity of the lithosphere; the stronger the lithosphere (i.e. more rigid), the deeper the base of the slope. Even more significantly, rigid lithospheres cause broader shelves, with sediment accumulating landward of the hinge line (compare A–C). Not illustrated here is isostatic subsidence due to sediment loading along active margins, which can be even more significant due to the fact that the subducting plate is not connected to the overriding plate, and so is even less rigid (Allen & Allen, 2005). When the lithosphere has zero rigidity (A), the isostatic subsidence due to sediment loading is referred to as Airy isostasy. (D) Similar subsidence is caused by compaction (compare A and D). (Modified from Reynolds et al., 1991.)
grains can evacuate the pore space (Suppe, 1985). This is because water is incompressible at near-surface temperatures and pressures, and so will resist compaction if it has nowhere to go. This resistance, which can be great enough to bear the complete weight of the overlying sediment load and thus prevent any reduction in pore space, will slow compaction until enough water escapes the pores that the grain framework can again assume the submerged weight of the load. The lower the permeability of the sediments or connectivity between pores, the slower the rate of dewatering and the longer it takes the sediments to compact (Bredehoeft & Hanshaw, 1968).

Being load dependent, isostatic subsidence and compaction generate the greatest accommodation space where the thickest accumulation of sediments occurs, which on most continental margins is beneath the shelf break (Reynolds et al., 1991; Fig. 5). The space created by these processes sets up a positive feedback by promoting additional sediment accumulation here, and so acts to stabilize the position of the shelf break. This influence diminishes, however, when the rate of lithospheric subsidence and compaction falls below the sediment-supply rate. When this happens, the margin can prograde even on lithosphere with zero rigidity. However, the subsidence generated by this loading will then continue long after the input of sediments slows or even stops.

While the impacts of isostatic subsidence and compaction on shelf width and slope relief share major similarities, there are also some important differences. Flexural isostasy affects a broader region than compaction (Reynolds et al., 1991). As the lithosphere flexes beneath a margin, it creates accommodation space to either side of existing strata. When incoming sediments then fill this space, the shelf is extended landward and the slope seaward even if the shelf break does not move (Fig. 5A–C). Thus lithosphere with higher rigidity facilitates the development of a broader margin.

Compaction, on the other hand, can cause smaller-scale subsidence than isostasy, as a result of shorter temporal and smaller spatial variations in sediment accumulation. This subsidence generally develops along the coast and inner shelf. Where rapid sediment supply from a large river has been greatly reduced or even cut off, deflation of the river deposits by compaction can lead to inundation of the coast and landward retreat of the shoreline (Milliman & Haq, 1996; Fig. 6). Where the subsidence becomes great enough, it may ultimately alter coastal topography to the point that river drainage to the area is recaptured and it once again becomes a major shelf depocentre. This process promotes deltaic lobe switching.

**Added effects of faulting**

Many factors (including sediment loading) can cause faulting, which affects far more than just the dimensions of the shelf and slope. Shallow faulting in sedimentary strata can lead to the slumps and slides that scar many continental slopes and even some continental shelves. Deeper faulting, including faults that extend into underlying or adjacent continental crust, can act as important structural controls on the course of rivers and submarine canyons (Kelling & Stanley, 1970; Dolgoff, 1998; McHugh et al., 1998). However, among the most important morphological impacts of faults on continental margins is their effect on accommodation space. Extensional (normal) faulting of continental crust during rifting creates the rugged descent to oceanic crust upon which the sedimentary shelves and slopes of passive margins develop (iv, Fig. 3A). Compressional (high-angle reverse and lower-angle thrust) faulting in accretionary prisms accommodates the addition of new wedges of scraped-off material to the base of the prism and causes the corresponding uplift and back-tilting of debris within the prism (Figs 3B & 4B).

A third important example of faulting that affects accommodation space is growth faulting. **Growth faults** are common in regions where sediment accumulation has been rapid (Suppe, 1985). These are listric or curved faults, in which movement along the fault is gradual rather than catastrophic (Fig. 7). The movement is driven by sediment loading on the basinward side of the fault, which acts somewhat like a spring. As it is loaded with sediment, this side slides downward along the fault, capturing the sediment while rotating the underlying strata (Fig. 7). Early fault movement occurs relatively easily, and a lot of sediment can be trapped. However, the accumulation of strata (toward the base of the fault) slows the faulting and reduces the sediment trapping. Eventually the fault freezes, and no further sediments are trapped.

Where common, growth faults can be significant in sequestering sediments on the shelf and
inhibiting margin progradation (Mitchum et al., 1990; Fig. 7). Large growth faults near the shelf break also can lower the slope gradient (Emery & Uchupi, 1984). In these instances, the seaward sides of the growth faults include parts of the slope, which are lowered when the strata are back-tilted as they are rotated by the curved form of the faults.

**PROCESSES THAT FORM THE SHELF PROFILE**

**Rivers, deltas and growth of the coastal plain**

Rivers deliver some 84% of the total sediment load that reaches the oceans (Milliman & Meade, 1983) and are the dominant means by which clastic sediments are transported to continental margins. Glaciers come in second, supplying 9% of the sediment load to the oceans, or more than twice the contribution of any remaining subaerial transport process (Scholle, 1996). However, temperate glaciers (e.g. south-east Alaska and west side of the New Zealand south island) have the largest natural sediment yields (i.e. mass/area) of any runoff in the world (Milliman & Syvitski, 1992). The sediment supply from glaciers during the last glacial maximum may have been many times larger than it is today. The Pleistocene glaciers bulldozed sediments onto continental shelves and even to the slope. As they retreated, the glaciers left terminal moraines that still rise above the seafloor and even above sea level (e.g. Long Island and Cape Cod; Shor & McClennen, 1988), and they released catastrophic ice-dam breaks (jökulhlaups) that deposited thick sediment wedges on the shelf (e.g. the Hudson Apron; Davies et al., 1992; Milliman et al., 1996; Uchupi et al., 2001).

Even during the glacial epochs of the Pleistocene, however, the majority of the world’s continental margins were more affected by rivers than by glaciers. Prior to the Pleistocene, the only glaciers affecting a continental margin still likely to be in existence today were those covering Antarctica (Kennett, 1982). Thus, while glaciers have had
a major effect on the present-day morphology of continental margins as low as 40° latitude, the effects of rivers have been even greater and so are focused on here.

Rivers have been active in delivering sediments to continental margins since their inception, but individual rivers are ephemeral on geological timescales. Tectonics, glaciations and stream pirating among other processes alter river courses and have created new rivers from old ones. So, while fluvial input to margins is continuous, the cast of rivers responsible for this input is subject to change (Bridge, 2003).

Large individual rivers, such as the Amazon, Indus, Yangtze and Mississippi, often act as a single point source of sediment to a continental margin. Where multiple rivers feed a margin, they can combine to form a line source (Jaeger & Nittouer, 2000). Sediment emanates from the mouth of each river, but on the shelf these contributions are smeared together by waves and currents, making it impossible to define where the input of one river

**Fig. 7** Sediment geometries in a dip section through experimental strata, formed as subsidence and sediment supply were held constant. Growth faults inhibit progradation of the shelf break by trapping sediments that otherwise would be transported onto the upper slope. The deposit varies in thickness because the rate of subsidence varied basinward across the tank in which the strata formed, but these rates did not change over the course of the experiment. A mixture of two sediments was used: quartz sand and coal. The latter served as a visibly distinct proxy for clays because its specific gravity is half that of quartz. The approximate position of the ‘shoreline’ during the experiment is demarcated by the abrupt boundary in sediment type between sand (light) and coal (dark). Note that during the stages in the experiment in which growth faults formed, ‘shoreline’ progradation was significantly impaired, with the sands being captured and rotated downward along the faults. Mass is conserved, so the ‘slope’ must still advance when the shelf break is held stationary. This occurs through the basinward movement of the base of ‘slope’, which is pushed outward by the toes of the growth faults, and results in an overall lowering of the ‘slope’ gradient. The terms ‘shoreline’ and ‘slope’ are in quotes because the features in the experimental strata do not scale up to continental margins. However, large growth faults are found on the Brazil and Niger continental slopes, and show clear indication of having contributed to their markedly low gradients relative to other margins (Emery & Uchupi, 1972). (Figure after Paola et al., 2001.)
ends and that of another begins. Broad lateral dispersal of river sediments is also caused by avulsions or wholesale shifts in the location of a river mouth (Bridge, 2003). These occur when the river channel upstream of its mouth is sufficiently breached during one or more flooding events that the river follows a new route to the ocean and abandons the old one (i.e. lobe switching). Using the Mississippi River as an example, avulsions ultimately can spread deposition across ~10^3 km of coastline, with major shifts in the river mouth location occurring roughly once every 1000 yr (Frazier, 1967; Fig. 6).

In settings where the descent of the river valley continues steeply below sea level, much of the fluvial sediment load may continue to be moved directly toward the deep ocean via slides, debris flows, turbidity currents or some other form of gravity-driven transport (e.g. the Markam River in Papua New Guinea; Krause et al., 1970). In these settings, little if any continental shelf is present.

Along more gradual continental margins, the sediments will come to rest close to shore, and if sea level is constant, the river will begin to build a delta (Elliott, 1979). Deltas are partly subaerial, partly submarine deposits that emanate from the river mouth in a fan shape (Fig. 6). They are characterized by a clinoform geometry in cross-section that can be subdivided into three parts (Fig. 8). The clinoform topset is a gentle seaward-dipping surface that on a delta corresponds to the outer coastal plain leading to the shore. It connects to a steeper inclined foreset (also known as the delta front), which on the delta extends below water level reaching another gradually dipping surface, the bottomset (also known as the prodelta).

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**Fig. 8** Factors controlling clinoform geometry in deltas. (A) Mississippi River delta and (B) Ganges–Brahmaputra subaqueous delta bathymetric profiles are examples of oblique and sigmoidal clinoform shapes (after Swenson et al., 2005), respectively, which are also shown schematically to the right of each profile (after Mitchum et al., 1977). These shapes result from variations in the rate of sediment accumulation across the clinoform foreset. (C) Oblique clinoforms are created when sediment accumulation is greatest at the top of the foreset. (D) Sigmoidal clinoforms result when accumulation is greatest farther down on the foreset. The accumulation rate is tied to the foreset because flow divergence here leads to a loss of velocity and carrying capacity. The faster a flow loses its velocity over the foreset and/or the coarser the sediments it is carrying, the more abrupt the decay in accumulation rates down the foreset and the more oblique the clinoform that is created. Note that on subaerial deltas like the Mississippi, the clinoform rollover lies near sea level. On subaqueous deltas, it is often located tens of metres below sea level.
The clinoform is seen in sedimentary bodies ranging from bedforms to entire continental margins, where the shelf corresponds to the topset, the slope to the foreset and the rise to the bottomset (Thorne, 1995). The clinoform is the reflection of a basinward change in sediment flux, in which sediment accumulation is greatest over the foreset. Consequently, the clinoform is the fundamental shape of prograding strata. However, significant sediment accumulation also can occur across the topset and bottomset regions of a clinoform. In fact, aggradation of the coastal-plain topset is an integral part of delta progradation. As the shoreline moves seaward with the delta, the distance the river must travel to reach the ocean increases. The river responds by depositing sediments along its path across the coastal plain to steepen its profile and give the river enough momentum to carry its remaining sediment load to the ocean (Swenson et al., 2005). This profile is characteristically exponential in form. The curvature results from the increase in river discharge toward the coast as tributary inflow is collected, which in turn lowers the gradient needed to drive the river toward the shore (Ritter, 1986).

Deltas and their coastal plains are important to margin evolution for several reasons. As rivers build deltas toward the shelf break, the width of the shelf is reduced (Fig. 9). When sea level rises, the deltas are drowned and their coastal plains become the foundation for a new, broader shelf (Fig. 9). This then becomes the platform upon which new deltas develop. How this development unfolds during sea-level change is the subject of a later section. The main point here is that fluctuations in sea level over geological time have led to numerous advances and retreats of deltas and their coastal plains. In turn, these have contributed significantly to the upward and outward building of the shelf and slope, with many shelves being the product of stacked delta/coastal-plain sequences. Examples include the ancient Miocene deltas underlying the current New Jersey and Angolan continental shelves (Steckler et al., 1999; Lavier et al., 2001).

**Bedload deposition, sediment plumes and clinoforms**

In general, about 90% of the sediments moved by rivers are carried in the water column as suspended load (Meade et al., 1990). The remainder is dragged and bounced along the riverbed as bedload. Both types of loads contribute to delta formation. On entering the ocean or an estuary, the discharge from a river spreads, loses velocity and thus loses its competence to move the bedload any farther, depositing it at the river mouth. In the absence of any other flow (e.g. tides), the bedload builds up to form a river-mouth bar. The river is then forced to flow around the bar leading to channel bifurcation. Through this process and also through the breaking of levees (i.e. crevasses) during floods, deltas develop distributary-channel networks (Elliott, 1979).

The suspended load of the river largely passes through the distributary-channel network and continues out into the ocean or estuary. Generally the mixture of fine sediments and freshwater is less dense and so spreads out as a surface sediment plume (Fig. 6). Under rare circumstances though, the suspended load may become so concentrated and/or the temperature of the water carrying the sediments may be so cold that the mixture temporarily exceeds the density of seawater and plunges beneath the surface as a hyperpycnal flow (Mulder & Syvitski, 1995; Mulder et al., 2003). Being denser than seawater at all depths, hyperpycnal flows move along the bed, as a gravity-driven flow, a mechanism that will be discussed later. Sediment plumes carry the suspended sediments over the submerged delta foreset. The plumes rarely extend directly offshore. Instead, they are turned parallel to the coast by alongshore winds, currents (e.g. the Amazon shelf; Curtin, 1986; Geyer et al., 1991), or even the Coriolis force if the plume is large enough (Hill et al., this volume, pp. 49–99). Due to lateral spreading and mixing with the ambient seawater, the speed of the plume drops, although ocean currents will continue to move it even after it has lost its river-imparted momentum (Geyer et al., 2000). Sediments suspended within the plume settle. These sediments are typically clays, but under rare circumstances silts and fine sands also have been observed (Syvitski et al., 1985). Consequently, plumes supply sediment to the submarine elements of the delta, principally the foreset and bottomset.

The curvature of the clinoform rollover between the topset and foreset of a delta is dictated by the change in sediment flux across this boundary; the more abruptly the flux drops, the steeper the
break in slope. Rivers carrying a high fraction of their sediment load as bedload, or those that enter a relatively protected margin in which wave and current energy is low, characteristically produce deltas with an abrupt rollover or oblique clinoform shape (Mitchum et al., 1977; Pirmez et al., 1998; Fig. 8A & C). The rapid deposition of sediment near the shore can build a foreset that becomes too steep and fails. As sediments at the top of the foreset cascade to its base, they lower its overall gradient. Accumulation at the top of the foreset then resumes, the process repeats itself and the foreset progrades by avalanching in a manner similar to a ripple or dune.

A different clinoform shape tends to result if the sediment load of a river is kept in suspension by waves and currents in shallow water on the shelf. In this case, most of the sediments bypass

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**Fig. 9** Response of continental-margin sedimentation to sea-level change. The majority of sediments delivered by rivers to the ocean accumulate along the coast. (A) In the absence of any change in sea level, the coastal plain will grow toward the shelf break and the shelf will narrow. (B) During a sea-level rise this pattern is reversed, if the rate of rise exceeds the rate of sediment supply from the rivers. When this occurs, the coastal plain is flooded and river valleys are turned into estuaries with the inundated regions becoming part of a broadening shelf. (C) When sea-level rise slows, and when it falls back below the sediment supply rate, the latter builds the coastal plain once again. (D) If the shoreline reaches the shelf break, the shelf disappears and the coastal plain links directly to the slope. Continental shelves and slopes evolve through the repetition of this cycle over the numerous rises and falls in sea level that occur during their existence. (Modified from Jervey, 1988.)
the topset and are carried to the foreset. Sediment accumulation gradually rises to a maximum over the upper foreset (Fig. 8D) then falls gradually over the remainder of the foreset and bottomset. The broader accumulation produces a low-angle foreset and a gentle clinoform rollover, known as a sigmoidal clinoform shape (Mitchum et al., 1977; Pirmez et al., 1998) (Fig. 8B & D). The enhanced accumulation of sediments below the top of the foreset elevates the bed. Ultimately this region becomes part of the topset, but as it is raised into the more energetic surface waters, deposition is increasingly inhibited and sedimentation is displaced into deeper waters over the evolving foreset (Pirmez et al., 1998). While oblique and sigmoidal clinoforms are different, they represent end members in a continuum of deltaic shapes. As the operative marine processes change, the clinoform shape can evolve toward one end member or the other.

The impacts of waves and currents on the shelf and shoreface profile

While rivers shape the profile of the coastal plain, waves and currents act to modify the profile of the shelf by redistributing the sediments deposited there (Wright, 1995). Waves and currents exert their most obvious influence on continental margins near the shore. A number of excellent texts treat the subject of coastal morphodynamics far more thoroughly than can be covered here (Komar, 1976; Bascom, 1980; Dyer, 1986; Wright, 1995; Short, 1999). The discussion here is limited to the general role of waves and currents in creating the shoreface and to a lesser extent the shelf profile seaward of it.

Neither waves nor currents are very effective at causing sediment movement on their own (see Hill et al., this volume, pp. 49–99). In the case of ocean currents, the bottom shear stresses they generate tend to be too weak to resuspend even the most moveable sediments (Grant & Madsen, 1979). There are exceptions of course. Strong tidal currents on the shelf can generate enough bottom shear in some settings to elongate and segment river deltas (Wright, 1977). The currents also can form flood and ebb tidal deltas at the mouths of inlets that connect estuaries and lagoons to the open ocean (Wright, 1995). Along the western margin of the North Atlantic, geostrophic currents can reach speeds strong enough to erode the continental slope (Poag & Mountain, 1987) and deposit large sediment drifts on the continental rise (Tucholke & Laine, 1982; Mountain & Tucholke, 1985). Generally speaking though, the flow divergence promoted by the large dimensions of the ocean successfully mitigates the generation of bottom currents vigorous enough to cause substantial erosion of the seafloor.

In contrast, shoaling waves can often generate bottom shear stresses strong enough to move sediments, particularly in shallow water (Grant & Madsen, 1979). The term shoaling is used here to refer to those waves that generate water motion that extends all the way to the seafloor (i.e. transitional and shallow-water waves; Fig. 10). The orbital water motion induced by waves is largest near the surface and decreases exponentially with depth (Knauss, 1997). When a water parcel is beneath a wave crest, it moves in the same direction as the waves. When the water parcel is beneath a wave trough, it moves in the opposite direction to the waves. In deep water, wave orbital motion is circular, and becomes negligible before reaching the bed (i.e. deep-water waves; Fig. 10). In shallower water, this orbital motion becomes elliptical due to interaction with the bottom (Knauss, 1997; Fig. 10). As water depth decreases, the bed-parallel diameter of the near-bed orbital motion increases, leading to greater near-bed wave orbital velocities.

Near-bed wave orbital motion generates shear stresses on the seafloor. These stresses are proportional to the square of near-bed wave orbital velocity, and therefore increase with decreasing water depth for a given surface-wave condition. When large enough, these shear stresses are able to move the bottom sediments, transporting them either as bedload or in suspension. The sediments are carried forward with the wave as the crest passes and are moved offshore beneath the following wave trough. If the sediment is relatively coarse, it may stop moving or settle to the bed between the forward and backward motion of the wave. Finer sediment can stay suspended throughout the wave cycle.

During storms, wave-generated orbital motion can affect the seafloor over much or all of the shelf. During fair-weather conditions, the influence of waves on the seabed is limited to shallow water depths. The region persistently affected by waves, the shoreface, typically extends from the outer
surf zone to the shoreward portions of the inner shelf. As waves cross the shoreface, they are modified by frictional interactions with the bed. As a result, onshore flow beneath wave crests becomes faster than the offshore flow beneath wave troughs, and there is a net shoreward movement of sediment transported by the waves (Komar, 1976). Net sediment transport rates are greatest, however, when waves and currents occur simultaneously (Grant & Madsen, 1979). Even if the currents alone are too weak to mobilize the sediments, the currents will still impose direction of transport when particles are resuspended by the waves. Thus the combined action of waves and currents gradually transports sediments across the shelf.

Waves and currents shift sediment grains until they are moved to a water depth where they no longer undergo net transport. This depth is referred to as wave base, or the depth at which ‘wave action ceases to stir the sediments’ (Baker et al., 1966; Fig. 11). Wave base has been equated to the shoreward limit of deep-water waves (Davies, 1980), but here we take the broader view of wave base also being linked to sediment grain size and reflecting a long-term equilibrium profile (Gulliver, 1899). In this context, wave base is a theoretical level more than an easily defined surface for two reasons. The first is that the water depth at which a wave begins to have an impact on the seafloor depends on its deep-ocean height and wavelength, and both vary with time and distance along a margin. For example, waves can become so large during major storms that they touch bottom over the shelf break. It is during such storms that waves should cause the greatest resuspension of sediments. However, waves during less severe but more frequent storms may cause more sediment movement over geological time, so the wave activity that has the greatest influence on the shelf profile is hard to quantify.

Second, in addition to varying with wave properties, wave base varies with sediment grain-size and density, as well as other sediment properties (e.g. cohesion). Larger near-bed shear stresses are required to move coarse sediments, and their transport is generally limited to the relatively shallow depths of the shoreface. Net shoreward transport under shoaling waves tends to accumulate sediment on the shoreface, steepening its slope relative to the remainder of the shelf. The end result then is an equilibrium shoreface profile that...
Seascape evolution on clastic continental shelves and slopes

Shoals, coarsens and steepens landward. Deeper exposures of predominantly sandy sediment are likely to be relict beds, last active during a period of lower sea level.

Finer sediment is more easily resuspended, and will remain in suspension until shear stresses decrease enough to allow it to settle. These conditions are most likely to occur in deeper water, where wave stresses are relatively low. There is a tendency for net seaward transport of fine sediment (Harris & Wiberg, 2002). Consolidation of fine-grained deposits increases their resistance to erosion, so that fine sediment deposition followed by adequate time for consolidation will produce beds with a low probability of remobilization. These conditions are likely to describe fine-grained deposits that form below wave base for typical storm conditions. However, observations on the Eel shelf and elsewhere demonstrate that muddy sediment can also accumulate at depths on the shelf characterized by storm-generated shear stresses that exceed critical values for erosion and deposition. Three key processes appear to be responsible for the development of these deposits:

1. Flocculation of sediment suspended in the water column, which increases particle settling velocity (Hill et al., this volume, pp. 49–99);
2. Consolidation of muddy deposits on time-scales of weeks to months (Wheatcroft et al., this volume, pp. 101–155);
3. Deposition from wave-supported sediment gravity flows of high-concentration, fine-grained suspensions (Parsons et al., this volume, pp. 275–337).

The long-term effect of deposition on the shelf is to fill available accommodation space, provided there is an adequate supply of sediment to the system from river or coastal sources. The shoreface is the region of greatest storm-generated erosion and deposition (Zhang et al., 1999; Wiberg, 2000; Harris & Wiberg, 2002), which has a large effect on the overall shape of the shelf profile. However, recent observations and modelling suggest that deposition from wave-supported sediment gravity flows that can develop in association with a supply of easily eroded sediment (e.g. after a river flood) may play an important role in mid-shelf strata development.
Subaqueous deltas and wave-supported sediment gravity flows

Mid-shelf clinoforms are increasingly being recognized on energetic shelves that also receive significant sediment flux from point or line sources of fluvial sediment. In these regions, sediment plumes deposit sediments across the inner shelf. In rare circumstances, the sediments may also be delivered by hyperpycnal flows. Waves and currents then rework the finer fraction of these deposits, which move into deeper, calmer waters where they accumulate below wave base and build up to it. Although this is the fate of most muddy sediment, some unusual conditions (e.g. river floods) can supply more mud than transport processes can remove from the inner shelf. In general, however, fine sediment moves seaward, and with large systems a submerged clinoform structure develops, referred to as a subaqueous delta (Fig. 12B). Examples are found associated with the Amazon River in Brazil (Nittrouer et al., 1986), the Huanghe (Yellow) River in China (Alexander et al., 1991), the Ganges–Brahmaputra River in Bangladesh (Kuehl et al., 1997), and the Fly River in Papua New Guinea (Walsh et al., 2004). Depositional bulges seaward of smaller rivers with high sediment yields are also suggestive of subaqueous deltas, but are not as well developed (e.g. the inner shelf seaward of the Eel River mapped by Goff et al., 1996). Like their subaerial counterparts, true subaqueous deltas build strata as a result of variable sediment accumulation across their surfaces. The topset aggrades upward, depending on the erosive character of ambient waves and currents. The foreset receives the bulk of sediment and progrades seaward over the small amount of fine sediment deposited as bottomset (Nittrouer et al., 1986).

**Fig. 12** Wave and current energy effects on delta morphology. (A–C) Subaqueous delta formation is linked to subaerial delta formation, as two ends of a spectrum for river-to-shelf profiles that will evolve depending on how fast sediments are delivered to the ocean by a river versus how fast they are then moved into deeper waters by waves and currents. (A) River-shelf profiles with a subaerial topset are at one end of this spectrum. These form where wave and current activity is low, such that the majority of the riverborne sediments accumulate close to shore. (C) River-shelf profiles with only a subaqueous delta are at the other end of the spectrum. In these settings, waves and currents are strong enough to carry all of the river sediments beyond its mouth and into water tens of metres deep. (B) The more common river-shelf profile where subaqueous deltas are found is one that contains both a subaerial and subaqueous delta. In these settings, the sediment supply from the river apparently overwhelms the capacity of the waves and currents to move all sediment farther seaward, but these processes still transport the majority of the sediments onto the shelf.
The combination of processes that move sediments across the topset of subaqueous deltas is more complex than simple wave or current motion. When the supply of freshwater and fine sediments to such a delta is great, convergent estuarine transport can concentrate suspended sediment and form a fluid mud (Kineke, 1993). This is a layer of suspended sediments with concentrations in excess of 10 g L\(^{-1}\). Once developed, fluid muds can move as gravity-driven currents, if the seabed is sloping. More recently, wave-supported fluid muds have been recognized where the suspended sediment is concentrated within a wave boundary layer (Traykovski et al., 2000; Hill et al., this volume, pp. 49–99; Parsons et al., this volume, pp. 275–337).

Turbidity currents are typically thought to form on the continental slope. These are mixtures of sediment and water that entrain additional water, deposit sediment and, if moving fast enough, erode the seabed (Parsons et al., this volume, pp. 275–337). On the slope and rise, the sediment suspensions in turbidity currents are maintained by turbulence that causes erosion as the current moves down an inclined seafloor. In general, the shelf is too flat for turbidity currents to reach erosional speeds. However, waves on the shelf can supply the additional turbulence needed to keep sediments in suspension once the fluid muds begin to flow on a topset (Friedrichs & Wright, 2004). Predominantly wave-supported flows then give way to purely gravity-driven flows when the fluid muds spill below wave base down the relatively steep surface of a subaqueous-delta foreset. Gravity flows have been observed on several subaqueous deltas (Wright et al., 1988; Kineke & Sternberg, 1995; Walsh & Nittrouer, 2004), and these have foresets that steepen downslope at a rate that appears to compensate for the decrease in bottom shear from waves, with increasing water depth (Friedrichs & Wright, 2004).

**Independent movements of the shoreline and shelf break**

As sediments are issued across the shoreline or shelf break, they pass from one process regime into another. The shoreline separates terrestrial sediment transport processes from their marine counterparts. The shelf break spans a more gradual transition from the predominance of waves and currents in pushing sediments across the shelf, to the ascendency of gravity-driven mass movements on the slope. What also commonly changes across these boundaries is the sediment flux. This latter change causes deposition or erosion, which in turn leads to movement of the shoreline or shelf break. For example, if rivers deliver more sediment to the shoreline than waves and currents remove, the shoreline builds seaward.

Changes in sediment flux across the shoreline often differ from those that occur across the shelf break, so these two boundaries commonly move at different rates and at times in different directions (Swenson et al., 2005). Two end-member scenarios exist. One is the case in which all terrestrial sediments are trapped on a delta and none are transported seaward across the shelf, which can occur if waves and currents are minimal (Figs 12A & 13A). The shelf narrows as progradation of the delta moves the shoreline seaward toward the sediment-starved and thus immobile shelf break (Fig. 13). Eventually the two boundaries merge; the shelf is eliminated and the coastal plain leads directly to the slope.

The other end-member scenario is when waves and currents are so energetic and thus so effective in transporting sediments across the shelf that all sediments reaching the shore are transported over the shelf break. Under these circumstances, the shore remains fixed or even retreats landward while the slope and thus shelf break progrades seaward (Fig. 13C). The result is an ever-widening shelf.

The evolution of most continental shelves undoubtedly lies between these two end members. It appears that a significant fraction of the fluvial sediment flux to the shore is trapped there, either in an estuary or in a subaerial delta. Of the remaining sediment, some accumulates on the shelf and the balance is transported beyond the shelf break and accumulates on the slope. As in the formation of clinoforms, the partitioning of sediments among the shore, shelf and slope is a function of sediment grain size and the energy of the shelf processes (Swenson et al., 2005). The coarser the sediment supplied and/or the less energetic the waves/currents, the greater the amount of sediment that accumulates nearshore (Fig. 13A). Likewise, the finer grained the sediments and/or the stronger the seaward transport, the farther the sediments can be transported toward the shelf break (Fig. 13B & C).

Where subaqueous deltas occur, a third boundary becomes important, the clinoform rollover. On
these shelves, most sediments from the river are trapped on the subaqueous delta foreset and do not make it to the shelf break (Fig. 12B & C). This sequestration of sediments leads to progradation of the subaqueous delta, so the position of its clinoform rollover with respect to the shoreline and shelf break can vary depending on the change in sediment flux across all three boundaries (Swenson et al., 2005) (Fig. 13B).

Of the three boundaries, the shelf break will most likely be prograding the slowest. For the shelf break to prograde, the slope must prograde, and, given its large accommodation space, the slope requires far more sediment to prograde than either the shoreface or the foreset of a subaqueous delta. In addition, the slope is likely to receive the least amount of sediment because it is depositionally ‘downstream’ of the shore and subaqueous delta. Thus with enough time, the subaqueous delta and/or the shore will reach and merge with the shelf break (Fig. 13).

The existence of an active subaqueous delta indicates that sediment is escaping to the shelf (Nittroer & Wright, 1994). Additionally, unless there is some change in the sediment flux bypassing the shore or in the wave and current activity...
moving these sediments to the foreset, the subaqueous delta will continue to outpace the shore and should merge with the shelf break first (Fig. 13B). The shore will then follow, as the subaerial delta encroaches on the more slowly prograding slope and shelf break.

**Shelf evolution during sea-level change**

As alluded to in the previous section, the evolution of the shelf and slope is subject to change as environmental conditions change. Possibly the most important of these in terms of direct impacts is change in sea level, which has at least three fundamental and simultaneous effects on shelf and slope morphology. The changes:

1. create or destroy space for accommodating sediments on the margin;
2. move the supply of terrigenous sediments from rivers and coasts away from or toward the shelf break;
3. force a corresponding translation back and forth across continental margins of all processes except those that are driven by plate tectonics.

Sea level is defined in terms of either eustasy or relative sea level. **Eustasy** is global sea level and is measured with respect to a fixed datum such as the centre of the Earth (Van Wagoner et al., 1988) (Fig. 14). The two most important controls on eustasy are tectonics and climate (Pitman, 1978). Tectonic events affect eustasy by changing the volume of the ocean basins. These events include the growth of the continents and, most importantly, changes in seafloor spreading rates. The latter alter the volume of mid-ocean ridges and the average age, temperature, isostatic buoyancy and thus depth of the seafloor. Climate affects eustasy principally through global cooling and warming spells. During global cooling, water is transferred from the oceans to glaciers, resulting in a drawdown of eustasy, while during global warming, the transfer is reversed and eustasy is raised.

When measured with respect to a more local datum, such as depth to basement, sea level is referred to as **relative sea level** (Van Wagoner et al., 1988) (Fig. 14). Relative sea level differs from eustasy in that it also depends on local subsidence/ uplift, which can cause the water depth at a location to deepen or shoal without any change in eustasy (Fig. 14). In this paper, the term sea level is used to mean relative sea level.

The geological record indicates that changes in sea level have rarely exceeded a few hundred metres (see summary in Allen & Allen, 2005).

**Fig. 14** Sea level defined in terms of either eustasy or relative sea level. Eustasy is sea level referenced to a fixed datum such as the centre of the Earth. Relative sea level is referenced to a local datum, often the Earth's crust. Note that both types of sea level differ from water depth in that they pertain to the elevation of the sea surface at a location and not the amount of water there. Eustatic changes in sea level are caused by global variations in the total volume of ocean water and/or the total volume of the ocean basins. Relative changes in sea level also include any local subsidence or uplift. Even if eustatic sea level is constant, relative sea level will rise if the region subsides or it will fall if the region is uplifted. In this paper, the term sea level is used to mean relative sea level. (Modified from Emery & Myers, 1996.)
Exceptions include the Mediterranean (Ryan et al., 1973) and Black Seas (Ryan & Pitman, 2000). During different eustatic falls in the Cenozoic, each was cut off from communication with the open ocean by a bedrock sill located at its mouth. While isolated, these Seas underwent significant evaporation, leaving not only their shelves but also their slopes exposed to subaerial erosion. When sea level eventually rose and topped the rock sills, flooding of the seas was catastrophic (sea level rose ~1900 m in the Mediterranean during the Messinian event, which caused a corresponding drop in eustasy of ~10 m; Ryan et al., 1973). The old shelves and slopes worn by subaerial erosion were drowned and became the slate upon which the current shelves and slopes have since formed. Along other margins, only the largest sea-level falls ever expose the uppermost slope. Consequently, it is the width and depth of the shelf that undergo the most dramatic transformation during sea-level change.

Maximum shelf depth is primarily controlled by the amount of sea-level change. Shelf width is also controlled by the pre-existing surface gradient. The lower the gradient, the greater the increase/decrease in shelf width for a given rise/fall in sea level (respectively), and thus the greater the change in accommodation space on the shelf.

In the absence of any other processes, falling sea level moves the shoreline seaward in a **regression**, while rising sea level moves the shoreline landward as a **transgression** (Fig. 9). Where sediments are simultaneously being supplied to the coast by rivers, movement of the shoreline begins to fall out of phase with sea-level change. This is because the sediment supply acts to continuously build the shoreline seaward through delta progradation (Fig. 13A). Consequently, regressions will occur not only whenever sea level is falling, but also whenever the rate of sediment supply to the shoreface exceeds the rate of sea-level rise. Transgressions will only occur when the sediment-supply rate to the shoreface falls below the rate of a sea-level rise.

Consider a cycle in which sea level falls and then returns to its original elevation (Fig. 9). During the fall, delta progradation will accelerate the shoreline regression (Posamentier et al., 1992). In general, the delta will continue prograding as long as the surface of the continental shelf being exposed by falling sea level has a lower gradient than the coastal plain created by deposition from the lengthening river feeding the delta (Blum & Tornqvist, 2000). However, where the sea-level fall exposes a surface with a greater dip than the coastal plain, the river will accelerate and potentially begin to incise the surface. Such incision appears to be common when a large fall in sea level exposes the steeper upper slope, or when a rapid fall exposes the shoreface or delta foreset (Vail et al., 1991). When it occurs, the incision propagates upstream along the river channel, smoothing the gradient increase while deepening and steepening its profile (Posamentier & Vail, 1988).

As sea level then rises, shoreline regression slows and eventually reverses into a transgression, when the rate of rise exceeds the sediment supply (Jervey, 1988; Fig. 9). The rise drowns the delta and coastal plain, and floods the river valleys, transforming their reaches in the vicinity of the new shoreline into semi-enclosed estuaries (Fig. 9). The estuaries develop until the supply of sediments from the river once again exceeds the rate of sea-level rise or sea level stabilizes. At this point, the rivers develop deltas, shoreline regression is renewed, and the cycle is repeated (Fig. 9).

The degree to which waves and currents have an impact on shelf morphology during sea-level change depends on how quickly they can reshape the shelf profile. If sea-level change is too rapid and/or the shelf sediments are too hard to erode, waves and currents will have little effect on the profile. Deltas, beaches, islands and the coastal plains will be drowned by the rise and become relict features on the new shelf. For relatively slow changes in sea level across shelves with erodible sediments, waves and currents will continuously be reshaping the profile toward one that is in equilibrium with and thus follows wave base (Fig. 15).

Waves and currents probably cause their greatest erosion during a sea-level fall, because it also lowers wave base (Fig. 15A). Shelf depths in the zone between the shoreline and wave base are increasingly eroded by waves. Where they are eroding, the waves cut down into finer sediments initially deposited in deeper water under calmer conditions. These sediments are moved seaward, increasing accumulation rates farther seaward on the shelf or even the slope (Fig. 15A). At the same
time, the asymmetric wave motions saltate coarser-grained sediments landward, helping build the beach seaward and promoting shoreline regression (Fig. 15A).

During a sea-level rise, the motions are reversed. Waves erode the flooding coastal plain, with erosion being greatest when a portion of the coastal plain is first submerged. The erosion diminishes as water depth increases and wave base rises across the new shelf surface. If wave base eventually rises above the seafloor at a location, erosion may switch to deposition as fine-grained sediments are swept offshore by wave and current activity. In areas away from rivers and thus receiving no sedimentation, the waves erode and rework coastal-plain sediments laid down during or prior to the sea-level fall (Fig. 15B). In areas where river deposition is occurring as sea level rises, the waves erode these more recent deposits and older coastal-plain sediments are preserved.

Coastal-plain preservation is enhanced even more where waves build coarse-grained beaches and barrier islands along the shoreline (Fig. 15C & D). As sea level rises, the waves have to rework these deposits before they can erode into coastal-plain sediments. In the process, the waves move the beaches or barrier islands landward. In fact, where these deposits have been built up to significant
enough relief, all of the reworking caused by waves can be expended in back-rolling the beaches/barrier islands such that there is no erosion of the former coastal-plain deposits (Cowell et al., 1995, 1999; Fig. 15C).

Again, the above scenarios assume wave reworking happens faster than sea-level change. This may occur on some shelves but probably not all. In fact, variations in the degree to which shelves are reworked to be in equilibrium with the wave climate may be one of the reasons shelf profiles are so varied.

PROCESSES THAT ACT TO LIMIT THE SLOPE OF THE CONTINENTAL SLOPE

Seafloor failure and submarine groundwater flow

Most continental slopes exhibit one of three types of profiles: (i) an oblique clinoform profile, (ii) a sigmoidal clinoform profile or (iii) a simple linear profile (Adams & Schlager, 2000). All three have been interpreted to result in part (the sigmoidal profile) or wholly (the oblique and linear profiles) from sediment mass movements caused by seafloor oversteepening and failure. If correct, then continental slopes should dip at an angle that approaches the threshold for sediment failure. Slope sediments are predominantly muds, which have a maximum angle of repose ~25–35° when dry. When saturated, the buoyancy force exerted by the pore waters in these sediments reduces this maximum angle by about half to 12.5–17.5° (Allen, 1985). This range is still about three to five times steeper than the ~4° dips that characterize most continental slopes, so factors other than just gravity must be at play in limiting slope gradients.

One of these factors may be earthquakes (Pratson & Haxby, 1996). Earthquakes are geologically frequent even on passive margins. For example along the USA east coast, the recurrence time of a magnitude 8 earthquake on the Richter scale is only 2000 yr (Seeber & Armbruster, 1988). The seismic shaking that results from an earthquake can cause a slope to fail at much lower angles than its sediments would support in a static environment. The threshold for failure may be lowered even further if the seismic shaking induces a cyclic loading on the sediments, which can increase sediment pore pressures and thus decrease the resistance of slope sediments to shearing under the downslope pull of gravity (Hampton et al., 1996; Lee et al., this volume, pp. 213–274). Anecdotal support for the impact of earthquakes is seen in the average slope gradients along USA active margins, which are roughly half those of USA passive margins (Pratson & Haxby, 1996). The one exception to this is the passive margin in the USA Gulf of Mexico. Ongoing deformation and flow of deeply buried salt have so disrupted the slope that its average gradient is <0.5°. The slopes offshore of Brazil and Nigeria have been similarly lowered by deformation of deeply buried muds (O’Grady et al., 2000).

The force driving the deformation of substrates along continental margins is the weight of rapidly deposited sediments. In fact, a correlation exists globally between gentle slopes, high sediment input and unstable substrates, making rapid sediment loading a second potentially important cause for the low average angle of continental slopes (O’Grady et al., 2000). Rapid sediment loading can destabilize slopes in a number of ways, the most important being the generation of pore pressure in excess of hydrostatic pressure, or that due to the weight of the overlying water column. Excess pore pressure results when sediments accumulate faster than they dewater (Suppe, 1985). Hydrostatic pressure reduces the maximum angle of repose for sediments by about half. Excess pore pressure can reduce the angle for failure even more, dropping it all the way to 0° if the pressure reaches lithostatic levels, i.e. it completely supports the weight of the overlying sediment column.

Under static conditions, failure of a mud slope at 2–4° requires a pore pressure approaching ~90% of lithostatic pressure (Prior & Suhayda, 1982). This may occur on a local basis and it may be an important cause for certain seafloor failures, but most continental-slope sediments have much lower pore pressures. Clear evidence of this are the dips of 10° to >15° on the walls of submarine canyons, which could never be achieved if the slope strata were so overpressured.

If pore waters are flowing, then pore pressures do not need to be significantly greater than hydrostatic to cause low seafloor gradients. This is because flowing pore waters exert a drag or seepage force on the sediments, and if a component of this force is directed downslope it adds to the pull of gravity,
lowering the threshold angle for seafloor failure (Iverson & Major, 1986). Seepage force appears to be an important factor in causing submarine slides and slumps along active margins (Orange & Breen, 1992; Fig. 16A–C). The force results from the expulsion of pore waters being driven out of the slope sediments as they are tectonically accreted to the margin. Along passive margins, seepage forces arise from differential sediment loading. Borehole measurements of anomalously high porosities in strata from the New Jersey (Dugan & Flemings, 2000; Fig. 16D) and Louisiana (Gordon & Flemings, 1998) continental slopes suggest pore waters are moving out of the lower slope in response to excessive sediment loading on the upper slope (Fig. 16E). In the Gulf of Mexico, permeability barriers to this

Fig. 16 Seafloor failures along continental slopes associated with submarine groundwater flow. Fluid flow through sediments exerts a drag or seepage force. This force can combine with the downslope pull of gravity and any reduction in sediment shear strength (due to elevated pore pressures) to cause the sediments to fail at significantly lower gradients than they would if the pore waters were not flowing. (A–C) Along active margins, the compression, folding and faulting of sediments in the accretionary prism expel pore waters, inducing a surfaceward flow that in turn can trigger failure. (A & B) This process appears to have played an important role in causing the numerous failures found along the seaward flanks of folded strata on the Oregon slope. (C) The process also may be promoting additional failures at these sites, because the pore-pressure field within the strata should be directed into the chutes and canyons excavated by the failures. (D & E) Fluid flow can occur beneath continental slopes along passive margins. (D) The driving force in these settings is enhanced sediment accumulation on the upper slope, which increases loading and compaction, and (E) drives pore waters toward the surface of the lower slope. Modelling suggests that even without considering the destabilizing effect of this fluid flow, the process can raise pore pressures to near lithostatic levels. (A, modified from Pratson & Haxby, 1996; B, modified from Orange & Breen, 1992; C, modified from Orange et al., 1994; D & E, modified from Dugan & Flemings, 2000.)
flow have produced lithostatic pore pressures. Oil companies have suffered considerable financial setbacks when drilling these regions, because the boreholes create an outlet for the overpressured pore waters, which cause the borehole walls to collapse.

**Bottom shear from internal waves**

Another important mechanism that may be limiting the steepness of the continental slope is internal waves (Fig. 17). These are waves that form within the ocean interior due to perturbations along density interfaces. Triggering mechanisms include the convergence and then divergence of ocean currents and tides as they flow over rugged seafloor topography, and the rise and collapse of sea-surface surges caused by the passage of low-pressure atmospheric storms (Knauss, 1997).

Internal waves form across small changes in density, so the gravitational restoring force damp-...
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Turbidity-current erosion and deposition

Turbidity currents also may be contributing to the characteristically gentle gradient of continental slopes. In addition to other formative mechanisms, turbidity currents can evolve from seafloor failures (Lee et al., this volume, pp. 213–274; Parsons et al., this volume, pp. 275–337). In fact, this is the way that a number of the best-known turbidity-current events have been generated (Heezen & Ewing, 1952; Malinverno et al., 1988). There is a strong association between eroded continental slopes scarred by numerous seafloor failures, and vast accumulations of turbidite deposits extending from the base of these slopes onto the continental rise, commonly in the form of submarine fans (Ericson et al., 1961; Walker, 1978). This combination has contributed to the mistaken viewpoint that the slope is a region of erosion and sediment bypassing, largely caused by turbidity currents. However, when the thickness of marine sediments is tallied globally, a different picture of the slope emerges. While underlying only 6% of the world ocean surface area, slopes contain 41% of ocean sediments, the largest volume for any marine physiographical province (Kennett, 1982). Thus continental slopes are sites of sediment accumulation far more than...
they are regions of erosion and sediment bypassing (Pratson, 2001).

Turbidity currents may be an important cause for this. One reason is that when large turbidity currents pass downslope through submarine canyons, they can overtop the canyon walls and spread sediments into intercanyon areas. This overbank deposition decreases with distance away from the canyon in a fashion analogous to floodplain deposition adjacent to rivers. As a result, the process creates levees along the walls of a canyon, increasing its relief (Rona, 1970). For canyons that have been subject to many large flows, the overbank deposition commonly shows evidence of having been deflected by the Coriolis force, for one side of these canyons (the right side in the northern hemisphere) is significantly higher than the other (e.g. Hudson Canyon; Shor & McClennen, 1988).

Turbidity currents may also build up intercanyon areas on the slope by directly flowing down them. As previously discussed, turbidity currents can be initiated by numerous triggering mechanisms (e.g. earthquakes, waves; Parsons et al., this volume, pp. 275–337), which cause resuspension of seafloor sediments on a sloping bottom. As these sediments settle back onto the seafloor, they are also acted on by the downslope pull of gravity. If concentrated enough, the sediments may begin to move as a turbidity current rather than simply coming to rest (Traykovski et al., 2000). Whether or not this turbidity current causes net deposition hinges on the amount of sediment it gains through erosion of the seabed versus the amount it simultaneously loses to settling. When erosion exceeds deposition in a turbidity current, it gains mass and thus accelerates, causing more erosion in a positive feedback termed ignition (Parker et al., 1986). If deposition exceeds erosion, then the turbidity current gradually loses mass and along with it the excess density driving its movement. Under these circumstances, the current slowly dies, as deposition falls off exponentially with distance downslope (Parker et al., 1986).

One of the most important variables determining whether a turbidity current ignites or dies is the existing seafloor gradient. Theoretical calculations supported by experimental modelling show that if the gradient of the slope is steep enough to cause ignition, repeated turbidity currents will eventually reduce the gradient until the mean drag force caused by the flows combined with the downslope pull of gravity is exactly balanced by the shear strength of the slope sediments (Kostic et al., 2002). If the slope gradient is so low as to cause a turbidity current to die, the seaward-thinning deposits from repeated turbidity currents will eventually steepen the gradient, and the currents will no longer deposit any sediments on the slope but instead bypass it (Gerber et al., 2004). Interestingly, the equilibrium slope in both cases is projected to be within several degrees of the mean angles characteristic of continental slopes (~4°). Consequently, if the slope is more or less steep, repeated turbidity currents should bring it back toward the angle of observed slopes (Fig. 18). Thus, turbidity currents join other mechanisms (earthquakes, pore-water flow and internal waves) that may be keeping the mean gradient of the continental slope far below the internal angle of repose for slope sediments. Turbidity currents are the only one of the mechanisms that will both build and erode the slope to maintain the observed gradients.

**PROCESSES THAT CREATE SUBMARINE CANYONS AND SLOPE GULLIES**

**Turbidity currents versus seafloor failure in forming submarine canyons**

Of the many facets to continental slope morphology, the one that has received the most study is submarine canyons. These are steep-sided, deep valleys that are often linear and largely directed downslope. However, they may also be meandering (von der Borch et al., 1985), or have sharp turns caused by structures such as faults (Song et al., 2000). In general, submarine canyons vary from hundreds of metres to kilometres in width, and from tens to hundreds of metres in relief. The relief is greatest on the upper to middle slope and diminishes toward both the continental shelf and rise.

Essentially all submarine canyons extend to the base of the continental slope. Where canyons begin on the slope, though, appears to relate to their width, relief and overall form. The narrowest and shallowest submarine canyons begin on the middle to lower continental slope, and are commonly U-shaped in cross-section (Fig. 19A). Canyons that begin on the upper slope are often larger, and can have V-shaped as well as U-shaped cross-sections

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The largest submarine canyons are those that incise the shelf break, particularly those that are or at one time were directly connected to a river (Twichell & Roberts, 1982; Farre et al., 1983). These canyons are predominantly V-shaped in cross-section.

Submarine canyons are morphologically similar to large subaerial canyons cut by rivers. Both share similar cross-sectional profiles (e.g. Scripps Submarine Canyon and the Grand Canyon of the Colorado River; Shepard, 1977), and concave-upward profiles along their channel axes. Like rivers, submarine canyons commonly link to form tributary drainage networks that gather sediments from across large reaches of a continental margin and funnel them onto the continental rise (Pratson, 1993). The striking similarity between submarine and subaerial canyons implies potential parallels in their evolution. This attraction has engendered significant investigation into the genesis of submarine canyons, a subject that continues to be debated.

Active submarine canyons occurring along narrow continental shelves suggest that there is a clear linkage between submarine-canyon formation and the supply of sediments from rivers and/or...
the continental shelf. In a few locations, rivers can be traced flowing directly into the heads of these canyons (Fig. 19B), but often the two are separated and even laterally offset from one another by an intervening shelf. In these latter cases, the sediments are supplied to the canyons by shelf currents, some of which flow in well-defined circulation cells (Gorsline, 1970; Felix & Gorsline, 1971). Advecting grains resuspended by waves, these currents gradually transport the sediments into the heads of the canyons where they indent the shelf break. The sediments are then periodically flushed down the remainder of a canyon, often when the buildup fails under cyclic loading by large waves from storms (Dill, 1964). The down-canyon movement of sediments can occur as a debris flow, but many submarine canyons incise older slope strata, and such erosion is more likely the result of turbidity currents. The abundance of turbidites that emanate from the mouths of canyons onto the continental rise suggests that downcutting by turbidity currents is one of the main ways that submarine canyons form (Ericson et al., 1961).

Submarine canyons currently separated from rivers and the nearshore by broad continental shelves commonly appear to be relict features, last
active when shorelines and thus sediment supply were near the shelf break during sea-level lowerings (Emery & Uchupi, 1972). However, a significant fraction of canyons have heads that occur well below the shelf break and show little if any morphological evidence of having been cut by turbidity currents (Twichell & Roberts, 1982; Fig. 19A). Instead, the scars of submarine slides often form the head and sidewalls of these canyons. The scars suggest that in addition to turbidity currents, seafloor failures can form submarine canyons, initiating them on the middle to lower slope (Fig. 19C). Such scars of submarine slides often form the head and sidewalls of these canyons. The scars suggest that in addition to turbidity currents, seafloor failures can form submarine canyons, initiating them on the middle to lower slope (Fig. 19C). Such canyons then appear to grow upslope through repeated failures of their headwall (Farre et al., 1983). Where these canyons breach the shelf break, they then tap into the supply of sediments stored on the shelf and become subject to turbidity-current erosion (Fig. 19C).

Slope failures need not correlate with times of high sediment input to the slope and low sea level, therefore canyon formation by this process can be similarly independent (Farre et al., 1983). However, some mechanism is required to cause continued failure of a canyon’s head wall and its upslope growth. One of the most viable is spring sapping (Robb, 1984, 1988; Orange & Breen, 1992; Orange et al., 1994). In this process, groundwater discharging from the face of a sedimented slope exerts a seepage force on the sediments, which in combination with the pull of gravity offsets the frictional forces pinning the sediments in place and can cause them to fail (Fig. 16). Where seepage failure is recurring and the material is carried away, a pit forms. The pit can then grow upslope as the seepage failure undercut the pit headwall and triggers retrogressive slides (headward-eroding) (Fig. 19C).

One of the attractions of spring sapping as a canyon-forming process is that it provides a possible explanation for the seeming regularity to the spacing of submarine canyons along certain continental slopes (Orange et al., 1994). Creation of a canyon dramatically increases the drop in pore pressure between the seafloor and the subsurface in regions where slope sediments are overpressured. This drives pore fluids to flow toward the canyon head, increasing the seepage force there and facilitating headward growth of the canyon through retrogressive failure (Fig. 16C). The convergence of flow at the canyon head also leads to a divergence of flow between canyons, reducing the likelihood of failure in these regions. Thus a self-organized feedback between hydrological and geomorphological processes develops, which limits the spacing between canyons. The factors that govern this spacing are the bathymetric gradient of the slope, the pore-pressure gradients within the slope sediments, and their material strength. Growth rate and thus size of the canyons is also important, as larger canyons can capture the hydrological discharge of smaller canyons, terminating their development.

However, not all canyons that appear to be created by retrogressive slope failure truly have that origin (Pratson et al., 1994). Some slope-confined canyons have completely buried, upslope extensions that may once have indented the shelf break. When followed downslope, these buried canyons connect with modern canyons, usually where the sediment cover has thinned and a seafloor trough has developed (Mountain et al., this volume, pp. 381–458). Seaward, the modern canyons have exhumed the buried canyons and follow the older paths to the base of the slope. These canyons have captured turbidity currents initiated along the upper slope and shelf break, and confined them to the former paths of the buried canyons. Downcutting by the turbidity currents then oversteepened the canyon walls, triggering failures and giving the appearance of having been formed entirely by slope failure (Pratson et al., 1994).

A greater understanding of submarine-canyon formation would come from knowing their age, but here too there are uncertainties and disagreements. The timing of canyon formation has rarely been established (Miller et al., 1987) because submarine canyons are erosive features that cut through the sediments that mark the onset of their formation. The problem is complicated even more by the likelihood that submarine canyons are formed by multiple processes (Shepard, 1981). Some canyons are long-lived and/or have complicated histories due to repeated episodes of downcutting, infilling and rejuvenation (Ryan et al., 1978; Goodwin & Prior, 1989; Mountain et al., 1996). Even the relationship to sea level is complicated, because sediment can be transported into the canyons at times other than lowstands (Carson et al., 1986; Kuehl et al., 1989; Weber et al., 1997; Droz et al., 2001; Mulder et al., 2001).
Consequently, existing theories for the timing of canyon formation fall between two end-member viewpoints. One is that submarine canyons are most active and probably form during lowstands in relative sea level, when rivers deliver sediment to a shoreline that is near or below the shelf break (Daly, 1936). At these times, there is no space left on the shelf to store sediments arriving from the continent and they are transported to the deep sea by turbidity currents, which play a major role in cutting the canyons (Kolla & Macurda, 1988; Rasmussen, 1994). This is supported by the common occurrence of abandoned (Emery & Uchupi, 1972) and buried shelf channels (McMaster & Ashraf, 1973; Berryhill et al., 1986; Burger et al., 2001) that lead toward canyon heads.

The other line of thinking is that submarine canyons initiated by slope failure may or may not be related to sea-level change. Mechanisms that can trigger failures independent of sea level and sediment supply include earthquakes (Garfield et al., 1994) and spring sapping (Robb, 1984). Another is the dissolution of sediments or sedimentary rock beneath the slope surface (Paull & Neumann, 1987), such as that occurring in carbonate rocks being infiltrated by groundwater. Two others are subduction (Orange & Breen, 1992) and diagenesis (McHugh et al., 1993), both of which can expel fluids through sediments with such force as to cause repeated slope failure. However, even spring sapping can be in-phase with sea level. Lowstands can allow the boundary between fresh and saline groundwaters to be pushed beyond the continental slope, causing freshwater to escape due to aquifer pressures (Johnson, 1939).

Multiple buried canyons incising regional unconformities are interpreted to have formed and infilled over the course of a single fall and rise in sea level (Fulthorpe et al., 2000). However, other nearby buried canyons preserved between sequence boundaries indicate that canyon formation is sufficiently variable in geological history and location on the slope as to occur at times other than just lowstands (Fulthorpe et al., 2000).

Turbidity currents versus seafloor failure in forming slope gullies

Smaller counterparts to submarine canyons on the continental slope are gullies (Fig. 20). These are channel-like features that can have as little as 1 m of relief. Slope gullies have not engendered the same level of interest as submarine canyons, but they are starting to receive greater attention because many recent studies have discovered them throughout the world.

Slope gullies can be separated into two general categories: erosional and aggradational. Erosional gullies can be further subdivided into submarine rills and dendritic gullies. Submarine rills are downslope-trending gullies that converge with one another or a nearby submarine canyon at low angles, in a fashion analogous to subaerial rills formed on terrestrial landscapes by runoff (Pratson et al., 1994; Fig. 20A). They range from narrow furrows 5–10 m wide and 1–2 m deep, to broader features 50–300 m wide and 10–40 m deep. Dendritic gullies intersect submarine canyons at high angles to give the latter a leaf-like, dendritic erosional pattern (Farre et al., 1983; McGregor et al., 1983; Fig. 20A & B). They are relatively short (0.5–5 km long), range from 75 to 250 m wide, and are ~10–20 m deep.

Dendritic gullies and rills are probably formed by different processes. Dendritic gullies occur on the steep slopes of submarine canyon walls (~14°–20°) and their sharply defined, irregular headwalls show clear evidence of having been excavated by slumps and slides (Fig. 20A & B). These failures are likely to be caused when the walls of the canyon

Fig. 20 (opposite) Submarine gullies on continental margins. Submarine gullies are smaller than submarine canyons but appear to be intimately related to them. (A & B) Dendritic gullies, which incise the walls of submarine canyons. Submarine rills are narrow, relatively straight gullies that extend down open-slope areas. (C–E) Aggradational gullies appear to originate from submarine rills, but persist even as the slope surface aggrades. Dendritic gullies are an integral part of submarine canyons. Submarine rills and aggradational gullies (A & D) are located on open slopes between canyons. However, on the New Jersey margin, submarine canyons have captured the drainage of submarine rills (left side, A), suggesting the two have grown from a common sediment source. On the California margin, aggradational gullies (C & D) feed into Trinity Canyon, forming an extensive tributary network (E). (A, modified from Farre et al., 1987; B, modified from Ryan, 1982; C, modified from Spinelli & Field, 2000; D & E, modified from Orange et al., 1994.)
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A

submarine canyons

dendritic gullies

B

1 km

New Jersey margin

C

124°30´W 124°25´ 124°20´

submarine rills

D

15 km

aggradational gullies

E

California margin

submarine canyon
are undercut by turbidity currents flowing through the canyon’s thalweg (McGregor et al., 1983). By comparison, the narrower rills are found on much-lower-gradient open-slope areas between canyons (Pratson et al., 1994) (Fig. 20A). This is also where aggradational gullies are found (Spinelli & Field, 2001) (Fig. 20C). Both appear to be created by turbidity currents. Whereas rills are cut by turbidity-current erosion, aggradational gullies grow by differential sediment accumulation, with less sediment accumulating on gully floors than on intergully areas (Spinelli & Field, 2001).

How turbidity currents generate rills and aggradational gullies on an open slope is an interesting problem. In this setting, turbidity currents have a tendency to diverge and form a sheet flow. The gullies, however, are an expression of flow convergence. In the absence of any pre-existing depressions, such convergence could develop if the head of the spreading turbidity current becomes destabilized and starts to finger (i.e. create isolated extensions of the turbidity-current head; Parsons et al., this volume, pp. 275–337). Such fingering could arise from small spatial and temporal variations in flow concentration and thus speed. Destabilization may also result from the increase in the centrifugal force acting on the turbidity current when the current passes over a curved break in slope. Fingering of a turbidity-current head will lead to variations in deposition and erosion that, if significant enough, could begin to channelize the flow by subsequent turbidity currents.

Submarine rills should form where the channelizing turbidity currents cause net erosion, but why the aggradational gullies develop is less clear. Analyses of seismic-reflection data indicate that aggradational gullies first develop where erosional rills have been buried by uniform and gradual sedimentation across large stretches of the continental slope (Burger et al., 2001; Spinelli & Field, 2001). This switch from an erosional to depositional regime is interpreted to occur following a lowstand, as sea level rises and the supply of sediments to the slope is reduced. While the rills are buried, their topographic expression is preserved even as the seafloor aggrades. Thus the rills are transformed into aggradational gullies. Complete infilling of the gullies is prevented, because turbidity currents and their high shear stresses are focused on gully floors.

When viewed from a more regional perspective, submarine rills and aggradational gullies are often found to feed into a large submarine canyon, and thus are part of a more extensive submarine drainage system (Orange, 1999; Fig. 20D). This relationship suggests that the rills and gullies could play a direct role in the evolution of submarine canyons. One idea in this regard is that submarine rills in an open-slope area grow and coalesce like their subaerial counterparts (Pratson & Coakley, 1996). Due to spatial variability in the frequency and size of the turbidity currents spilling down the slope, some rills will deepen and widen faster than their neighbours, altering the local slope, and capturing drainage in a fashion analogous to stream piracy (Fig. 20A). Eventually, the larger rills become deep enough and their walls steep enough that failure occurs. The chutes excavated by these failures then evolve into headward-eroding canyons that advance upslope following the path of the rills. This advance is caused by retrogressive failure of the canyon head wall each time it is undercut by turbidity currents that flow down the rill.

The above idea invokes the need for both turbidity-current erosion and retrogressive mass wasting in submarine-canyon formation. It also links the evolution of slope-confined canyons to those that breach the shelf break, and provides an alternative mechanism to spring sapping for the continued episodic failure of the canyon headwall (Pratson & Coakley, 1996). Finally, it provides a mechanism for creating submarine canyons with multiple heads, such as Hudson Canyon on the New Jersey slope and Eel Canyon in northern California. Sediment flows that enter a submarine canyon downslope of its existing head can trigger retrogressive failures, which then excavate an additional canyon head along the path of the flows, back to their shelf-break source.

FUTURE RESEARCH

This paper has attempted to synthesize the fundamental ways that major processes operating along shelves and slopes affect the morphological evolution of seabed surfaces. The information has come from work in a variety of geological disciplines. It draws upon findings from plate tectonics, geomorphology and hydrology, physical ocean-
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graphy, geophysics and geological engineering, and, of course, sedimentology and stratigraphy. In many respects, the last discipline, stratigraphy (i.e. sequence stratigraphy), has provided the best context for understanding the evolution of seabed surfaces in shelf and slope settings.

As discussed in Mountain et al. (this volume, pp. 381–458), sequence stratigraphy is founded upon a cogent theory for why strata should be grouped into sequences and systems tracts. More importantly, this theory has engendered a number of landmark studies in a broad range of fields that have refined its fundamental tenets. These include seminal papers on such topics as: the role of thermal and isostatic subsidence in sequence formation (Pitman, 1978; Watts & Steckler, 1979); the synchronicity of global changes in sea level and their implications for past climates (Vail & Mitchum, 1977; Haq et al., 1988); and the power of numerical modelling for gaining insights to sequence evolution (Burton et al., 1987; Jervey, 1988).

The main focus of sequence stratigraphy has been to clarify why certain sediments are found where they are in relation to other sediments. However, development of the underlying theory also has been utilized to develop linkages between shelves and slopes, and the sediments accumulated in both regions. Therefore, much of the recent research described in this paper reinforces ideas envisioned by early pioneers in sequence stratigraphy. However, this is only partially true.

As originally conceived, sequence stratigraphy addressed strata formation under a well-defined, but limited, set of environmental drivers. These were principally changes in eustasy, subsidence and sediment supply. The causes and consequences of changes in eustasy and subsidence are now better understood. By comparison, the role of sediment supply in shelf and slope evolution remains poorly constrained. This is because sediment supply is far more variable in both space and time than either eustasy or subsidence. In addition to sea level, sediment supply to a margin is determined by the three-dimensional geometry of both the ocean basin and the drainage systems feeding into it. These in turn are subject to internal reorganization by the processes transporting sediments to the basin, including river avulsion, delta lobe switching, and large-scale failures on land and beneath the sea. Changes in lithology can also have a significant effect, contributing to lags in the response of sediment supply to environmental change. Then there is the basic question of how and how fast sediment processes actually modify the shelf and slope, and ultimately contribute to the development of their stratigraphy.

The latter is the question initially addressed by the STRATAFORM programme. Continued work needs to be done in documenting the tie between what happens on the seabed, and what is preserved in the shelf-slope stratigraphic record. However, STRATAFORM has demonstrated the importance of viewing strata formation from the dual perspectives of processes and products.

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