# N

## NATURAL BRIDGE

Remnant arch-shaped formation developed through erosion of the surrounding bedrock. Natural bridges, or stone arches, are unusual features that predominantly develop in horizontally bedded sedimentary rocks such as sandstone and limestone, though they hardly ever occur in metamorphosed or igneous rocks. They may form in a variety of ways, though all are ephemeral and will eventually collapse. The most common mode of formation is by water erosion, forming in deep valleys with highly sinuous rivers. Eventually, the river will cut across the neck of the entrenched meander by eroding a route through the obstructing rock outcrop. Often this can be accomplished without the arch collapsing, thus forming the natural bridge. Natural Bridge, Virginia, USA, has an uncertain evolutionary history, though meander cutting by the James River is a strong possibility (Malott and Shrock 1930). The other possible mode of formation is by the collapse of an underground drainage tunnel, leaving a remnant of the tunnel ceiling. Natural Bridge spans 30 m across Ceder Creek and is one of the few remaining natural bridges that is used as a transport bridge, at 60 m high.

Natural bridges may also be formed by the near complete collapse of underground tunnels. Such formations are common on the Hawaiian Islands, where recent lava tunnels roofed by a solidified crust may collapse leaving all but a small arch-shaped portion. Other origins of natural bridges include those cut by the sea resulting in coastal wave-cut arches, while a more unusual natural bridge can be found in Petrified Forest National Park, Arizona, USA, where a silicified tree trunk, known as the Onyx Bridge, spans a canyon 15 m wide.

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SEE ALSO: arch, natural

STEVE WARD

## NEBKHA

Nebkha, or nabkha, is an Arabic term given to mounds of wind-borne sediment (sand, silt of pelletized clay) that have accumulated to a height of some metres around shrubs or other types of vegetation. They are sometimes called shrub-coppice dunes. They may occur on bigger dunes, in interdune areas, on pan surfaces, near wadis and on or behind beaches. Morphometric data are provided by Tengberg and Chen (1998). The largest nebkhas (mega-nebkhas) accumulate around clumps of trees. In the Wahiba Sands of Oman these can be 10 m high and up to 1 km long (Warren 1988).

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A.S. GOUDIE

## NEEDLE-ICE

Needle-ice (synonymous to 'pipkrake' or 'kammeis') is the accumulation of ice crystal growths in the direction of heat loss at, or directly beneath, the ground surface. Although needle-ice usually grows perpendicular to the ground surface, curved ice-filaments are sometimes observed owing to wind and gravity effects. Needle-ice may also connect normal to plant stalks that have drawn sufficient ground moisture. Needle-ice is common to areas of diurnal freeze–thaw, ranging from tropical alpine to subarctic environments.

Needle-ice best develops in moist, fine-textured sediment with at least 10 per cent clay/silt. Precise soil and near-surface thermal dynamics affecting needle-ice growth and decay are complex, thus making it difficult to predict annual frequency. Generally, needle-ice develops within the first hour of ground temperatures dropping below 0 °C. Further conditions necessary for needle-ice development include a relatively low soil water tension to enable ice segregation to take place and adequately rapid movement of unfrozen moisture to the freezing front, so that it corresponds with the rate of latent heat loss, and thus preventing the *in situ* freezing of pore water (Outcalt 1971). Very windy conditions may cause a rapid temperature drop through the soil pores, thus reducing the suction gradient and enhancing the development of pore ice rather than needle-ice. Typically, needle-ice phases will entail periods of growth, stagnation and ablation. The duration of freeze determines growth phases and consequently needle-ice length, which may vary from a few millimetres to several centimetres. Polycyclic or multilayered needle-ice, separated by thin veneers of sediment, occurs where there has been moisture stress. Alternatively, long-lasting growth phases over several days may produce multilayered needle-ice lengths exceeding 400 mm.

Needle-ice has been applied to studies examining SOIL CREEP, SOIL EROSION, the impacts on plant (particularly seedling) disruption and its function as a geomorphic process in miniature landform development. Needle-ice on stream banks or soil terraces commonly extrudes sediment, which is transferred by needle-ice induced direct particle fall, sliding and toppling failure and mini-mudflows. Geomorphological consequences of needle-ice as an erosion agent include notches and undercut fluvial banks, TURF EXFOLIATION and associated depositional microforms. Several soil structures including nubbin soils, gaps around stones and other varieties of PATTERNED GROUND, have been attributed to needle-ice. It is thought that soil stripes aligned parallel to the late morning sun may be a function of shadow and differential thaw effects during the ablation phase.

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SEE ALSO: freeze–thaw cycle; frost heave; ice

STEFAN GRAB

## NEOCATASTROPHISM

Neocatastrophism, as defined by Schindewolf (1963), refers to the explanation of sudden extinctions in the palaeontological record. In geomorphology, George Dury (1975, 1980) expressed the view that high magnitude, low frequency events were more important in an absolute sense than low magnitude, high frequency events in moulding the Earth's landscapes. Dury's statement expresses the essence of the issue. Neocatastrophism is a response to one hundred years of geomorphic thinking in which the predominant role of low magnitude, high frequency events in landform evolution had become the prevailing paradigm. A side issue, expressed in an exchange between Brunsden (1996) and Yatsu (1996), is whether the word catastrophism should be excised from the geomorphic vocabulary, and hence, by implication, also the word neocatastrophism. I am not convinced that we need to fear this word; but there is a need for unambiguous definition. By contrast with catastrophism, which is an outmoded, pre-twentieth century mode of thought, neocatastrophism is thought to be an increasingly relevant way of viewing the geomorphic world.

Circumstances which have favoured the emergence of neocatastrophism include the following:

- improved precision in geochronology has demonstrated unexpectedly rapid past changes;
- the exploration of mass extinctions in the past has intensified;
- some geomorphological features, such as the Channeled Scablands of eastern Washington, are more amenable to explanation by low frequency, high magnitude events than by gradual, semi-continuous processes;
- space exploration has generated a strong interest in galactic scale events;
- interest in global environmental change has provided evidence of rapid past changes, such as found in the polar ice caps and the oceanic deep sediments;
- the rise of non-linear dynamics and chaos theory is beginning to provide ways of synthesizing gradualism and catastrophism.

Within geomorphology, it was the paper by Wolman and Miller (1960) which provoked a critical evaluation of the question of magnitude and frequency (see MAGNITUDE–FREQUENCY CONCEPT) of the operation of geomorphic processes. The authors directed attention to the importance of medium size and medium frequency events as having the greatest cumulative influence on the landscape. This was an important insight, but did not prevail after notable discussions by Wolman and Gerson (1978), Gould (1984), Gretener (1984) and Baker (1994).

Wolman and Gerson (1978), in following up their findings on magnitude and frequency, expanded on effectiveness of climate and relative scales of time such that they were forced to modify their view about the importance of the intermediate magnitude and intermediate frequency event in landform history. Introduction of the idea of the length of time over which a landform survived suggested that, in many cases, it was the extreme events which were most important. The influence of this paper on geomorphic thinking cannot be overestimated as it has emphasized the importance of combining measures of process magnitude and frequency with duration or lifetime of landforms.

Gould's discussion on punctuational change was an equally influential paper for the whole of Earth science (Gould and Eldredge 1977). The essential concept was a recognition that many important changes in Earth history have proceeded by relatively rapid flips between more stable conditions. Systems often absorb stress and resist change until the stresses accumulate past a breaking point. Systems then flip to a new stable state. This hypothesis, known as punctuated equilibrium, has gained widespread acceptance in the palaeontological community and it is viewed by other Earth scientists as a model for processes of inorganic change. Gretener (1984) advances the example of isostatic rebound to illustrate the relativity of gradualism. Isostatic rebound has been active during the last 10,000 years and is still in progress in such places as northern Canada and Scandinavia. The process covers all of humanity's conscious history and is generally perceived as a gradual phenomenon. However, if one considers that the Earth's skin can completely recover from the unloading of 1–2 km of ice within a period of 15–20,000 yrs, this process is effectively instantaneous from an Earth history perspective. This leads to a consideration of what constitutes an event? Gretener suggests that the duration of an event occupies no more than 1/100 of the total time span being considered. On this basis, geological processes may have durations as great as 10 Ma and still qualify as events. Indeed Earth history 'reveals long periods of tranquility interrupted by moments of action' (Gretener 1984: 86). The rare event in geology is a punctuation with such a low rate of occurrence that it has taken place, at most, a few times through all of Earth history. It is unscientific to call such events 'impossible'. Punctuationism would possibly be a better term than neocatastrophism. Nevertheless, either term is preferred to uniformitarianism, which fails to do justice to such extreme events. Brunsden (1996) illustrates Gretener's point well in his Figure 2.3.

Baker (1994) provides the most powerful geomorphic justification for neocatastrophism in his interpretation of the resistance of the geological community to Harlen Bretz's (1923) theory of the origin of the Channeled SCABLANDS of eastern Washington. He explains that the community was blinkered by its slavish adherence to gradualism and its suspicion of the mention of cataclysmic flood events. Nevertheless, Bretz's interpretation was finally vindicated in many of its neocatastrophist details, in part as a result of the identification of a source of this exceptional flooding (glacial Lake Missoula) which Bretz himself had not recognized, but also in part because of the recognition of the erosional and hydraulic concomitants of an extreme flood.

There remains an urgent need for geomorphologists to accommodate our thinking to the new diastrophic ideas associated with global plate tectonics. The focus on short timescales relevant to process studies has been partly responsible for a neglect of the longer timescales. Modes of vertical motion, the onset of Ice Ages and the appearance of volcanism all need to be reappraised in a neocatastrophist framework.

Thorn (1988) points out that there is an important intellectual issue associated with the rise of neocatastrophism. If greater significance is being attached to large events in a series, this only forces an adjustment of magnitude–frequency concepts. If the new perspective is one that identifies unique events as paramount in geomorphological records, then there can be no science of geomorphology because there is no science of uniqueness. Most of us are busily adjusting our magnitude–frequency concepts.

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SEE ALSO: catastrophism; magnitude–frequency concept

OLAV SLAYMAKER



# NEOGLACIATION

Neoglaciation is a geological term, originating in North America, used to describe the period during the latter half of the Holocene when valley GLACIERS in many mountain areas readvanced to their maximum extent following Pleistocene DEGLACIATION. The term was first used by Moss (1951) and Nelson (1954) who attribute it to Matthes (though the term appears in none of his papers). Neoglaciation was formally defined by Porter and Denton (1967) as a 'cool geologic-climate unit ... indicating a probable world wide synchrony of glacier fluctuations in response to climatic change'. Their classic paper established the standard division of the North American Holocene into a warmer and drier early Holocene (the Hypsithermal) followed by a cooler Neoglacial Interval characterized by several periods of glacier advance. The related term 'little ice-age' was first used by Matthes (1939) to define the period when glaciers re-established in the Sierra Nevada of California following the post-glacial climatic optimum. Matthes's 'little ice-age' was, in fact, what is now termed Neoglaciation. Subsequently, the term Little Ice Age (LIA) has been almost universally adopted to describe the latest and most severe part of the Neoglacial during the past few centuries when glaciers in many areas of the world reached their maximum Holocene extent (Grove 2003).

This terminology was established at a time when there were few detailed chronologies of Holocene glacier fluctuations with little absolute dating control (radiocarbon dates were just becoming generally available to Quaternary scientists). Holocene climates were interpreted on the basis of limited evidence from studies of glacier fluctuations and the zonation of pollen diagrams in Europe and North America. As the maximum Neoglacial (Little Ice Age) extent of glaciers at most northern hemisphere sites was between AD 1600 and 1850, almost all morphological evidence of earlier glacier events was destroyed. Stratigraphic evidence from lateral MORAINES and sections within the Little Ice Age limits was fragmentary, difficult to find and the dating often poorly constrained.

Over the past thirty to forty years new information has led to the modification of our knowledge and understanding of these glacial events. Significant glacier recession during the late twentieth century has exposed many new moraine sections and buried forest sites that yield detailed evidence of earlier glacier fluctuations. The advent of AMS and calendar-adjusted radiocarbon dates, plus dendrochronological dating of sub-fossil wood using millennial-length tree-ring reference chronologies, and the development of proximal varve sequences have improved the available record of dated Neoglacial sequences (Plate 83).

Most evidence of Neoglacial glacier fluctuations comes from western North America and western Europe where, generally, the LIA glacier advances were the most extensive. However, in the southern hemisphere several authors have identified deposits of an early Neoglacial advance c.4,400–4,600 yr BP, downvalley of the LIA limits. This evidence is critically reviewed by Porter (2000) who cautions that this conclusion should remain provisional until a larger population of better dated sites are available.

Early work in the northern hemisphere (mainly in Alaska and Scandinavia) identified three phases of Neoglaciation: early (c.6,000–4,000 BP), middle (2,500–3,500 BP) and late (last 1,000 years or LIA) with a minor event c. AD 700–900 (Denton and Karlen 1973). Evidence for the earliest events is fragmentary. Most investigations

Plate 83 Late Neoglacial (Little Ice Age) lateral and terminal moraines, Bennington Glacier, British Columbia, July 1990

date initial Neoglacial advances between 4,000–5,000 $^{14}$C yr BP and link them with other PALEOCLIMATE evidence of climate deterioration at this time. Although the preceding Hypsithermal was originally defined as a time stratigraphic unit (Porter and Denton 1967), dates for the Hypsithermal–Neoglacial transition are clearly time transgressive with evidence for some alpine glacier advances before 6,000 BP. Therefore the early Neoglacial is not well defined. There is widespread evidence for glacier advances between c.3,500–2,800 $^{14}$C yr BP in the Canadian Rockies, Alaska, Switzerland, Patagonia and Scandinavia. There is also evidence from several areas of glacier advances c.1,300–1,500 $^{14}$C yr BP and an 'early medieval advance' c.AD 600–800. However, the most detailed (and best dated) reconstructions of glacier fluctuations from the Alps (Holzhauser 1997) indicate that at least seven advances of the Aletsch Glacier occurred between 3,200 yr BP and AD 1000, plus three major LIA advances. It seems unlikely that the history of glacier fluctuations at less well-dated sites is any less complex than that shown by the Aletsch. Therefore the history of glacier fluctuations during the early and middle Neoglaciation remains incomplete but probably consists of multiple, relatively short-lived (50–200 years?) advances that appear to have been progressively more extensive over time and were separated by periods of glacier recession. In assessing the synchroneity of these events it is critically important to determine both the precision of the dating technique used and the precision of dating control (i.e. its stratigraphic or geographic relationships with the event being dated). In many cases the limiting dates are +/−50 years at best which is often inadequate to differentiate between synchronous and closely spaced events or determine whether the events are correlative and synchronous over large areas rather than simply locally significant records.

The beginning of LIA is traditionally placed at the end of the Medieval Warm Period. The MWP was initially defined from non-glacial evidence in Europe (Hughes and Diaz 1994) and encompasses the period AD 800–1200 when there is little evidence of extended glacier cover. The status of this period as a global interval of generally warmer conditions remains questionable until an adequate database of high resolution palaeoclimate records becomes available. Well-dated, early LIA glacier advances occurred in the twelfth to fourteenth centuries in Patagonia, Canadian Rockies, Alaska, Switzerland and Scandinavia. These early advances were followed by an interval with little evidence of glacier fluctuations until the main LIA advances, dated between the sixteenth and nineteenth centuries. In many areas glaciers reoccupied positions at or very close to their maxima several times during the LIA, e.g. the early 1700s and mid-1800s in the Canadian Cordillera and coastal Alaska or the 1350s, 1650s and 1850s in the Alps. Most glaciers have receded rapidly during the twentieth century. The exposure of old buried forest sites and the alpine iceman suggests that this twentieth-century recession is the most rapid and severe during the Holocene. However, minor advances of glaciers occurred in many alpine areas during the 1960s–1970s and glaciers in western Norway advanced significantly during the 1980s and early 1990s as a result of increased winter precipitation due to changes in atmospheric circulation. In summary, these records indicate several intervals of glacier expansion over the past 5,000 years. Some, such as the nineteenth century, appear to be globally synchronous (at least at the centennial scale) whereas others may reflect local or more regional glacier histories.

The development of independent, high-resolution proxy climate records spanning the late Holocene (using tree rings, ice cores and other techniques) has greatly expanded our knowledge of climate variability and climate history. This work has shown that climate varies continuously at several spatial and temporal scales; that the relationships between glacier fluctuations and climate are complex; and that climate variability is rarely synchronous at the global scale. Recent palaeoclimate work also provides superior records of climate forcing. Some forcing factors have globally synchronous effects (e.g. variations in solar output, sunspot minima, etc.) whereas the effects of others may differ between hemispheres (orbital effects, volcanic eruptions) or between regions (e.g. circulation changes). Global climate variability reflects the interaction of all these factors as do climatically dependent fluctuations of glaciers. However, despite these strong links to climate, glacier behaviour and response times are also influenced by many factors unrelated to climate. The use of glacier-defined terms such as Neoglaciation and Little Ice Age to identify distinct, global, climate-geologic periods is inappropriate and misleading in the context of current knowledge of



Holocene climates. Usage of these terms should be confined to describing the glacial advances of the late Holocene after c.5,000 BP and between c.AD 1000–1900, respectively.

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SEE ALSO: dating methods; dendrochronology; Holocene geomorphology; palaeoclimate

BRIAN LUCKMAN

# NEOTECTONICS

Neotectonics concerns the study of horizontal and vertical crustal movements that have occurred in the geologically recent past and which may be ongoing today. While most crustal movements arise directly or indirectly from global plate motions (i.e. tectonic deformation), neotectonic studies themselves make no presumption about the mechanisms driving deformation. Consequently here 'movements' is a vague catch-all term that encompasses a myriad of competing deformation processes, such as the gradual pervasive creep of tectonic plates, discrete (seismic) displacements on individual faults and folds, and distributed tilting and warping through isostatic readjustment or volcanic upheaval. The phrase 'geologically recent past' is also intentionally vague. Early attempts to define the discipline by arbitrary time windows such as Late Cenozoic, Neogene or Quaternary have given ground to a more flexible notion that envisages neotectonism starting at different times in different regions. The onset of the neotectonic period, or the 'current tectonic regime', depends on when the contemporary stress field of a region was first imposed. For instance, in the Apennines of central Italy the 'current tectonic regime' began in the Middle Quaternary (~700,000 years ago) and it is even younger (< 500,000 years) in California; in contrast, in eastern North America the present-day stress regime has been in existence for at least the last 15 million years.

Typically then, neotectonic movements have been in operation in most regions for the last few million years or so. Over such prolonged intervals, neotectonic actions are revealed by the stratigraphical build-up of sediments in inland and marine basins, the burial or exhumation histories of rocks and the geomorphological development of landscapes. Geological studies of palaeobotany and palaeoclimate, numerical models of landscape evolution and techniques such as fission track analysis and cosmogenic dating are among the disparate tools unravelling this long-term tectonic activity. Over periods of many tens of, to several hundred thousand years, the actions of individual tectonic structures (faults and folds) can be determined, unmasked by their deformation of geomorphic markers, such as marine and fluvial terraces, and tracked with reference to the late Pleistocene glacial-eustatic time frame. The apparently smooth deformation rates discerned over intermediate timescales are revealed to be episodic and irregular when faults and folds are examined over Holocene (10,000 years) timescales. Over millennial timescales, secular variations in the activity of tectonic structures can be gleaned from a diverse set of *palaeoseismological* approaches, from interpreting the stratigraphy of beds that have been affected by faulting to detecting disturbances in the growth record of trees or coral atolls.

Although neotectonic movements continue up to the present day, the term *active tectonics* is typically used to describe those movements that have occurred over the timespan of human history. Active tectonics deals with the societal implications of neotectonic deformation (such as seismic-hazard assessment, future sea-level rise, etc.), since it focuses on crustal movements that can be expected to recur within a future interval of concern to society. Even contemporary crustal movements may reveal themselves in Earth surface processes and landforms, such as in the sensitivity of alluvial rivers to crustal tilting. In addition, geomorphological and geological studies are important in recording the surface expression of Earth movements such as earthquake ground ruptures which, due to their subtle, ephemeral or reversible nature, are unlikely to have been preserved in the geological record. However active tectonics also employs an array of high-tech investigative practices; prominent among these are the monitoring of ongoing Earth surface deformation using space-based or terrestrial geodetic methods (tectonic geodesy), radar imaging (interferometry) of ground deformation patterns produced by individual earthquakes and volcanic unrest, and the seismological detection and measurement of earthquakes (seismotectonics), both globally via the World-Wide Standardised Seismograph Network and regionally via local seismographic coverage. These modern snapshots of tectonism can be pushed back beyond the twentieth century through the analysis of historical accounts and maps to infer past land surface changes or deduce the parameters of past seismic events (historical seismology). In addition, earthquakes can leave their mark in the mythical practices and literary accounts of ancient peoples, the stratigraphy of their site histories, and the damage to their buildings (archaeoseismology). The time covered by such human records varies markedly, ranging from many thousands of years in the Mediterranean, Near East and Asia to a few centuries across much of North America. Generally they confirm that regions that are active today have been consistently active for millennia, thereby demonstrating the long-term nature of crustal deformation, but occasionally they reveal that some regions that appear remarkably quiet from the viewpoint of modern seismicity (such as the Jordan rift valley) are capable of generating large earthquakes.



In reality, the distinction between neotectonics and active tectonics is artificial; they simply describe different time slices of a continuum of crustal movement. This continuum is maintained by the persistence of the contemporary stress field, which means that inferences of past rates and directions of crustal movement from geological observations can be compared directly with those measured by modern geodetic and geophysical methods. Although the terms 'neotectonic' or 'active' are somewhat blurred and are often used interchangeably, societal demands (for instance, regulatory authorities for seismic hazard, nuclear safety, etc.) often require the incidence of tectonic movements to be strictly defined. For instance, the present definition in Californian law of an 'active fault' is one that has had surface-rupturing earthquakes in the last 11,000 years (established when the Holocene was considered to have begun at that time) (see ACTIVE AND CAPABLE FAULT). Other regulatory bodies recognize a sliding scale of fault activity: Holocene (moved in the last 10,000 years), Late Quaternary (moved in the last 130,000 years) and Quaternary (moved in the last 1.6 million years). Neotectonic faults, by comparison, are simply those that formed during the imposition of the current tectonic regime. 'Real' structures, of course, are unconstrained by such legislative concerns. Many modern earthquakes rupture along older (i.e. palaeotectonic) basement faults. Indeed, it is important to recognize that any fault that is favourably oriented with respect to the stress currently being imposed on it has the potential to be activated in the future, regardless of whether it has moved in the geologically recent past.

A more meaningful way to differentiate styles and degrees of neotectonic activity is in terms of tectonic strain rate, which is a measure of the velocity of regional crustal motions and, in turn, of the consequent tectonic strain build-up. Crustal movements are most vigorous, and therefore most readily discernible, where plate boundaries are narrow and discrete. In these domains of high tectonic strain, frequent earthquakes on fast-moving (>10 mm/yr) faults ensure that a century or two of historical earthquakes and a few years of precise geodetic measurements are sufficient to capture a consistent picture of the active tectonic behaviour. Intermediate tectonic strain rates characterize those regions where plate-boundary motion is distributed across a network of slower moving faults (0.1–10 mm/yr). Examples of such broad deforming belts are the Basin and Range Province of western USA or the Himalayan collision zone, where earthquake faults rupture every few hundred or thousand years, ensuring that the Holocene period is a reasonable time window over which to witness the typical crustal deformation cycle. In contrast, low-strain rates ensure that intraplate regions, often referred to as 'stable continental interiors', are low-seismicity areas with slow-moving (< 0.1 mm/yr) faults that rupture every few tens (or even hundreds) of thousands of years, making the snapshot of human history an unreliable guide to the future incidence of tectonic activity.

The global pattern of present-day crustal motions can be accounted for by PLATE TECTONICS theory, that elegant kinematic framework in which rigid plates variously collide, split apart and slide along their actively deforming boundaries. Closer inspection, however, reveals that the basic rules that govern global plate motions (i.e. rigid blocks separated by narrow deforming boundaries) break down at the regional and local scale. This is particularly so on the continents, where a patchwork of pre-existing geology and structure ensures that tectonic stresses are not applied in a uniform, straightforward fashion. Studies of how the contemporary stress field varies across the Earth's surface (Figure 108) distinguish between first- and second-order stress provinces. First-order provinces have stresses generally uniformly oriented across several thousands of kilometres. The largest of these are the midplate regions of North America and western Europe, where the stress fields are largely the far-field product of ridge push and continental collision. In contrast, first-order stress provinces in tectonically active areas are dominated by the downgoing pull of subducting slabs and the resistance to subduction. Second-order stress provinces are smaller, typically less than 1,000 km across, and are related to crustal flexure induced by thick sequences of sediments and postglacial rebound, and to deep-seated rheological contrasts. Although the bulk of the Earth's crust is in compression, significant regions of extension occur. In both the continents and oceans, these extensional domains are long and narrow and correspond to topographically high areas, though notable exceptions are the Basin and Range province and the Aegean region of the eastern Mediterranean. Most first-order stress provinces, and many second-order stress provinces coincide with distinct physiographic provinces.

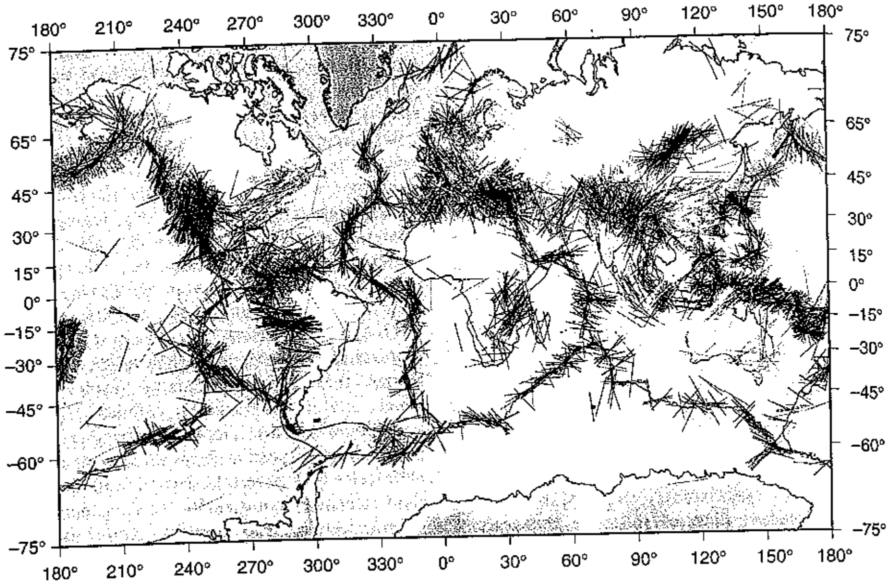


Figure 108 The World Stress Map with lines showing the directions of maximum horizontal compression. Black lines denote normal faulting (extension), dark grey lines denote strike-slip faulting, and light grey lines denote thrust faulting (compression); white lines show an uncertain tectonic regime. The longer the line length, the better the quality of the data. Around two-thirds of the stress data come from earthquakes and so highlights where the bulk of tectonic deformation is occurring; most of the remaining third comes from borehole stress measurements that are concentrated in petroleum-producing provinces. From Mueller *et al.* (2000)

Plate driving forces may exert the dominant control on the contemporary stress field, but another process contributes to crustal deformation at a global scale. That process is glacial isostatic adjustment (GIA), the physical response of the Earth's viscoelastic mantle to surface loads imposed and removed by the cycles of glaciation and deglaciation to which the planet has been subjected for the past 900,000 years (see GLACIAL ISOSTASY). Because large ice-mass fluctuations induce the sub-crustal flow of material, measurable crustal deformation extends for thousands of kilometres beyond the limits of the former ice margins (Figure 109); in short, the effects of GIA are felt globally. In addition, while the crust's elastic response to ice-sheet decay is geologically immediate, the delayed viscoelastic response of the mantle ensures that GIA persists long after the ice has gone. Although the effects of GIA can now be detected from space geodesy, its legacy is most clearly visible in the worldwide pattern of postglacial sea-level changes. Regions that were ice covered at the Last Glacial Maximum are uplifting (i.e. relative sea level is currently falling) as a consequence of postglacial rebound of the crust. Likewise the regions peripheral to the former ice sheets are subsiding (i.e. relative sea levels are rising) due to collapse of the 'glacial forebulge'. The effect of this subsidence outside the area of forebulge collapse is to draw in water from the central ocean basins, which is compensated by uplift in the ocean basin interiors in the far-field of the ice sheets. The final GIA component is the hydro-isostatic tilting of continental coastlines due to the weight applied to the Earth's surface by the returning meltwater load, which produces a 'halo' of weak crustal subsidence around the world's major land masses. For the most part, geological studies of Holocene relative sea-level changes are consistent with the uplift/subsidence



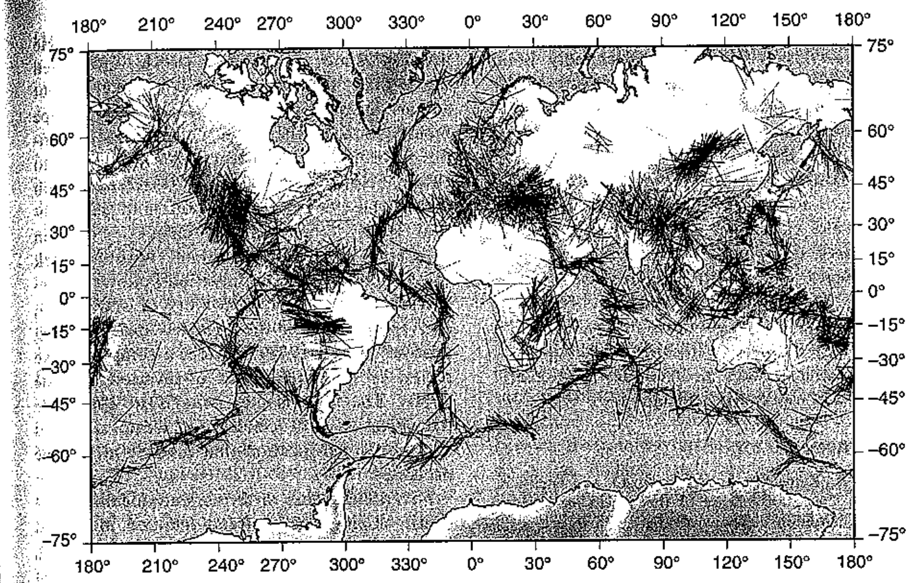


Figure 109 Map showing the outward radial motion of eastern North America (predicted by Peltier's (1999) postglacial rebound model) due to glacial isostatic adjustment to removal of the Laurentide ice sheet, and highlighting the concentration of contemporary seismicity along the former ice margins. From Stewart *et al.* (2000)

pattern predicted by global viscoelastic theory. The key areas of misfit are along plate boundary seaboards (especially subduction zones), where tectonic deformation dominates, and those areas 'contaminated' by local anthropogenic effects (groundwater extraction etc.).

The neotectonic implications of GIA are not confined to the coastline. Glacial rebound is now widely considered as an effective mechanism for exerting both vertical and horizontal stresses not only within the limits of the former ice sheets but for several hundred kilometres outside. Within the former glaciated parts of eastern North America and northern Europe both tectonic and rebound stresses are required to explain the distribution and style of both postglacial and contemporary seismotectonics. Outside in the ice-free forelands, predicted glacial strain rates are still likely to be one to three orders of magnitude higher than tectonic strain rates typical of continental interiors. Consequently, some workers argue that an apparent 'switching on' of Holocene earthquake activity in eastern USA and the occurrence of atypically large seismic events such as the great ($M > 8$) earthquakes that struck the Mississippi valley area of New Madrid in 1811–1812 may be associated with areas where glacial strains are particularly high. Glacial loading and unloading may also disturb the build-up of tectonic strain at glaciated plate boundaries, such as today in Alaska or previously when the Cordilleran ice sheet capped part of the Cascadia subduction zone. More recently, the isostatic component of glacier erosion in the mountain-building process is becoming appreciated.

In summary, the worldwide pattern of vertical and horizontal crustal movements arise from the global effects of plate motions and glacial isostatic adjustment. Regionally and locally, this is augmented by flexure from eustatic or sediment loading, volcanic deformation or anthropogenic change (dam impoundment). While many neotectonic investigations seek to disentangle movements arising from the imposition of tectonic strains from those augmented by non-tectonic processes, this is often a fruitless holy grail; because deformation of the Earth's crust typically induces compensatory flow underlying mantle, neotectonic movements are applied globally. Nevertheless, these disparate contributory mechanisms, coupled with the varying timescales over which their actions can de discerned, ensure that neotectonics encompasses a remarkable breadth of research disciplines. Few other fields easily blend topics as disparate as space science, seismology, Quaternary science, geochronology, structural geology, geomorphology, geodesy, archaeology and history. It is this interdisciplinary marriage that makes neotectonics particularly exciting and especially challenging.

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IAIN S. STEWART

# NIVATION

Nivation is a morphogenetic term introduced by Matthes (1900) to describe and explain the processes associated with late-lying seasonal snow patches and landforms derived from them (nivation benches or terraces, and nivation hollows). The term became entrenched in periglacial geomorphology with little attention to process measurements until recently. One important vein of thinking envisages nivation hollows as precursors of glacial cirques.

While Matthes (1900) fails to produce a sharp definition of nivation, he exhibits a sophisticated grasp of snowpack accumulation dynamics. He invokes static snowpacks with intensified freeze–thaw around snowpatch peripheries, but assigns nivation only modest powers of landscape modification. Furthermore, while Matthes identifies a form continuum from nivation hollow to cirque, he distinguishes sharply between nivation and glacial effects and does not claim that nivation hollows enlarge into cirques. Nivation was soon adorned by others with bedrock freeze–thaw weathering, nivation hollows as precursors of cirques, the mobility of snowpacks, and solifluction as the primary mass wasting process. Thorn (1988) provides a comprehensive review of the development of nivation into the 1980s. The fundamental issue is to appreciate that nivation is a concept of weathering and transport intensification that invokes no unique processes.

Nivation benches or terraces are idealized as a gentle sloping flat or tread mantled in debris, unvegetated where snow is especially late-lying, with a steeper riser at the upslope end. Expansion is by headward incision promoted by the presence of late-lying snow at the inflection of slope. When suitably oriented such landforms, whatever their origin, are highly likely to become snow accumulation sites and it becomes tempting, if not irresistible, to move from correlation to causation, a step fraught with problems in the absence of process measurements. Nevertheless, available evidence does suggest that nivation is a likely mechanism for expansion in poorly consolidated materials (Thorn 1988; Berrisford 1992; Christiansen 1998). Where such forms occur in bedrock, headward incision becomes dependent upon more contentious weathering processes, rather than merely upon excavation by mass wasting.

While it is easy to envisage that the additional water supply associated with late-lying snow has considerable geomorphic potential, detailed field measurements were not undertaken until the 1970s (Thorn 1988). Important subsequent work includes Berrisford (1991, 1992) and Christiansen (1996, 1998). A reasonable summary of present knowledge of the mass wasting component of nivation is to state that late-lying snow does indeed accelerate or intensify periglacial mass wasting processes (e.g. solifluction, surface wash) by several factors, even by orders of magnitude, in comparison to nearby snow-free (or thinly snow-covered) surfaces. However, the literature is not adequate to specify a consistent pattern of process or process rate intensification; indeed, considerable variability appears likely. On unconsolidated surfaces the elimination of vegetation cover by late-lying snow (not always the case) appears to represent an important process threshold. As rainfall inputs decrease and snowfall inputs increase proportionally, snowpatch meltwater influences emerge more starkly, especially in poorly consolidated materials (Christiansen 1998). In the presence of permafrost snowpatches may have important impacts on near-surface water flow (Ballantyne 1978), particularly by raising shallow subsurface flow to the surface where a snowpatch sustains a frozen subsurface.

While the role of nivation within the periglacial transport suite is generally non-problematic, and increasingly emphasizes meltwater impacts, the weathering role of nivation is problematic. For much of its history nivation was generally assigned no chemical weathering role. Intensification of chemical weathering processes beneath and around snowpatches is now documented (e.g. Thorn 1988), with a spatial pattern strongly dependent upon meltwater pathways that may even shift the impact downslope of the snowpatch itself. Knowledge of the role of nivation as a modifier of freeze–thaw weathering is largely constrained by the uncertainties associated with freeze–thaw weathering itself (Hall *et al.* 2002). Relevant ground climates (as opposed to largely irrelevant generalized air climates) are poorly known, laboratory studies do not necessarily mimic field conditions adequately, nor do they effectively isolate freeze–thaw from other possible mechanisms. Within the immediate context of nivation the critical issues rest with the interaction between snowpack insulation modifying, and perhaps eliminating, sufficient thermal regimes versus the obvious addition of abundant and necessary moisture through snowmelt. Berrisford (1991) found morphological evidence in the form of angular clasts beneath some portions of snowpatches to suggest enhanced mechanical weathering of coarse debris. However, he also emphasized the geomorphic importance of the annual temperature cycle, as opposed to shorter cycles, and views perennial snowpatches as protective.

Unlike CRYOPLANATION, of which it is a critical component, nivation research has been reinvigorated in recent years. While field data is increasingly available, definitional problems continue. Thorn (1988) suggested the term is so broad that it will always defy definition and should be abandoned, while Christiansen (1998) would like to expand it to embrace all snow-related processes making it equivalent to 'glaciation' in generality. Perhaps neither path is advisable, but the sharp contrast serves to highlight the problems presently associated with the term.

Snow-derived process will always be central to periglacial geomorphology and lead to some broad issues (Thorn 1978). Most nivation researchers appreciate that the wind-derived nature of a snowpatch means that there is potential for snow-bearing winter winds to orient landscape development through nivation. In fact, such orientation passes through a second filter, namely, available, suitable topographic traps because deep, late-lying snow cannot accumulate on a flat surface. Such concepts lead to ideas focused upon the landforms produced by snow-dominated regimes as opposed to those of full glaciation. Nelson (1989) goes so far as to suggest that not only is nivation central to cryoplanation, but that cryoplanation terraces are periglacial analogues of glacial cirques. His thesis invokes the presence of cryoplanation terraces where snowfall and temperature regimes are inadequate to generate cirque glaciation. Yet another view of nivation juxtaposes it with cold-based, that is non-erosive, glaciation. In such a context Rapp (1983) has suggested that interglacial nivation may represent the erosive, landforming regime and glaciation the quiescent, protective one.

The spatial extent of seasonal snowcover and lengthy interglacial periods, perhaps even more importantly relatively short pleniglacial periods, suggest that geomorphic processes derived from late-lying snow merit considerable attention. Clearly, nivation represents a core concept in such an appreciation, albeit not an exclusive one. Intensification of surficial mass wasting processes by nivation is now well established, but determination of systematic trends and rates must await generation of considerably more data. The weathering regime associated with nivation remains uncertain. Concentration of snowpatch meltwaters promotes chemical weathering; however, the location and degree of mechanical weathering with respect to a fluctuating seasonal snowpatch remains questionable, despite widespread willingness to invoke it. The extent to which nivation is able to shape a landscape is simply unknown.

Nivation cannot seemingly initiate topographic lows, but it can certainly modify them – but to what extent, in what fashion, and at what rates? Christiansen (1998) demonstrates headward expansion of nivation hollows in soft materials infused with permafrost, Berrisford (1992) suggests downslope expansion of nivation hollows. Thorn (1976) calculated nivation excavation rates in colluvium that would not produce a cirque from a nivation hollow in a feasible period of time. Quantitative research into such topics is urgently needed, but is immediately confronted by scale-linkage issues. In particular, periglacial process studies conducted on mesoscale phenomena must contend with the possibility that PARAGLACIAL conditions, rather than prevailing ones, hold the key.

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SEE ALSO: cryoplanation; freeze–thaw cycle

COLIN E. THORN

# NIVEO-AEOLIAN ACTIVITY

Niveo-aeolian processes involve the entrainment, transport and deposition of fine (mainly sand-sized) particles by wind in seasonally snow-covered areas, and modification of such sediments during snowmelt. The defining characteristic of niveo-aeolian activity is the deposition of wind-blown sediment on snowcover. This involves either simultaneous deposition of mixed sediment and drifting snow, or deposition of windblown sediment alone over earlier snowcover. Sediments deposited by wind on snowcover are referred to as niveo-aeolian deposits. The term *denivation* is used to refer to the processes, microforms and sedimentary structures associated with melting of the underlying snowpack.

Niveo-aeolian activity occurs in polar deserts, in subarctic environments, on alpine plateaux (Ahlbrandt and Andrews 1978) and on maritime mountains (Ballantyne and Whittington 1987). Most niveo-aeolian deposits are annual (associated with complete melting of snow in summer), though in exceptionally cold arid areas of Antarctica perennial deposits consisting of interstratified sediment and snow occur. Elsewhere, buried snow may persist under niveo-aeolian deposits for one or more summers (Bélanger and Filion 1991).

Sources of niveo-aeolian sediments include unvegetated or partly vegetated floodplains or outwash plains, raised beaches and deltas, aeolian sandsheets and dunes, glacial deposits, and sandy regolith. Sediments are entrained by wind in autumn or spring when snowcover is incomplete, or during winter when strong winds strip snow from the crests of ridges or hummocks. Sublimation of pore ice and abrasion by blowing sand are important in releasing sand particles from frozen surfaces. Most sand-sized particles travel by saltation or creep over ice or crusted snow, with fine sand and silt particles travelling in suspension. During violent storms coarse granules may travel up to 4 m above the surface and pebbles may be blown over ice (McKenna Neuman 1990).

Niveo-aeolian deposits have poor to moderate sorting and a wide range of modal grain sizes, but most are dominated by medium and coarse sand (0.2–2.0 mm). Fresh deposits often reveal concentrations of sediment at the top and base of the snowpack, layers of mixed snow and sediment, and sediment-rich layers separated by clean snow. During melt, sediment becomes concentrated at the snow surface. Such supranival deposits tend to be thickest on the lee of obstacles, notably on the slip faces of dunes. Melt of the underlying snow produces a range of denivation features including dimpled surfaces, snow hummocks, contorted bedding, sinkholes, cavities, tension cracks and faulted slipface strata (Koster and Dijkmans 1988; Dijkmans 1990). Meltwater fans may accumulate at the lower edge of niveo-aeolian beds (Lewkowicz and Young 1991).

Unless rapidly buried under prograding slipface beds, most denivation structures dry out and are destroyed by summer winds. Niveo-aeolian deposition therefore rarely leaves any distinctive sedimentological or structural signature in cold-climate aeolian sequences. For this reason the role of niveo-aeolian activity in the formation of the extensive Late Pleistocene COVERSANDS and dunefields (see SAND SEA AND DUNEFIELD) of Europe and North America remains contentious.

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SEE ALSO: aeolian processes; nivation; periglacial geomorphology

COLIN K. BALLANTYNE

# NON-LINEAR DYNAMICS

Geomorphologists have long recognized that landforms are the result of a complex set of interactions operating over different scales of space and time (Schumm and Lichty 1965; Schumm 1979; Brunsden and Thornes 1979). More recently, these ideas have been variously strengthened, confronted and extended by incorporating findings and analytical tools from the subject of non-linear system dynamics (NSD) developed in the mathematical and physical sciences. The term 'non-linear' expresses an unequal relationship between the driving force or the stress and the geomorphic response, most simply described as where *outputs are disproportionate to inputs*. A good example is the classic Hjulström curve of velocity of water plotted against the size of entrained sediment particles. It is a non-linear curve showing that the output (sediment size entrained) is not proportional to the input (water velocity); factors such as particle density and inter-particle cohesion are also important. The geomorphology of a landscape is governed by the interaction of a vast array of such processes operating in different parts of a landscape and over different timescales. Hence, the term 'non-linear dynamics' is used to describe the behaviour of the system rather than the behaviour of discrete process interactions. In the example of an unstable slope system, this means that the relationship between the intensity of precipitation falling on a slope and the size and timing of landslides may be complex and non-linear, rather than a simple cause and effect. It is widely believed that all complex systems consisting of interacting components behave non-linearly. The attractions of NSD lie with the possibilities of finding generic insights about geomorphic system behaviour and mapping well-studied model behaviours onto real systems that would normally be non-observable by conventional field methods. In practice, the demonstrable existence of model phenomena in real systems and hence the usefulness of NSD ideas are contested.

NSD may provide useful insights about (1) the predictability and unpredictability of geomorphic phenomena; (2) distinguishing between spontaneous and forced geomorphic change; (3) the sensitivity and resilience of landscapes to impact; and (4) the use of appropriate conceptual and modelling scales and methodologies.

A useful division in the discussion of non-linear system behaviour is to identify intrinsic or extrinsic changes. Intrinsic changes are those that spontaneously occur through self-organization as part of the system's own dynamic without any direct and proportional external forcing, analogous to the idea of biological evolution. One explanation is provided by the so-called 'arrow of time' implicit in the second law of thermodynamics, which predicts that a system will develop towards an equilibrium point where the free energy is minimized and thermodynamic entropy (disorder) is maximized. Thus, free energy in the form of flowing water drives the organization of a river network, especially its drainage density, so that the total potential of the flowing water to erode and carry sediment is minimized (or diffused) across the landscape. As in the classic Davisian cycle of landform evolution, thermodynamic equilibrium is reached when the relief is progressively lowered to a peneplain. However, since the Earth's surface is not dominated by peneplains but rather by an array of other geomorphic forms, it is clear that most landscapes have had their 'arrow of time' development arrested. Therefore, embedded within the idea of 'arrow of time' are other system behaviours that explain geomorphic systems set at some point far from equilibrium, known as dissipative systems (Prigogine 1996).

One of these, emergent complexity, describes the conditions found in open, and often partitioned systems that show ordered states maintained by flows of energy across the system boundary. Within human timescales, they may appear unchanging in their underlying forms but over longer timescales these systems evolve or emerge to become increasingly ordered and complex, as in the example of progressive weathering leading to soil horizons supporting a complex terrestrial ecosystem. Another, chaotic behaviour, may explain the local variability of many geomorphic systems. Chaos is most easily seen in mathematical equations that describe processes such as turbulent water flow, but the implications of chaos theory for geomorphology are large. Chaotic behaviour means that the exact pathway of a set of interacting processes over time is crucially sensitive to initial conditions and to even the smallest external perturbations. Over a long period of time, an observer of a chaotic landscape would see small initial differences in relief, drainage and soils amplified over time: divergence, rather than convergence. One result might be a mosaic of soil types overlying a fairly uniform parent material that initially differed from place to place only in small differences in texture. The variability would be amplified by subsequent vegetation succession and positive feedback controls. In one sense this means that a chaotic system, as we know for weather forecasts, is an unstable system becoming progressively unpredictable as the system moves from its starting point. However, model chaotic systems also evolve to lie within well-defined ranges, known as attractors, which translates to the natural environment in terms of the degree of variability at a higher scale being constrained and therefore predictable in probabilistic terms. Thus the overall range of soil types encountered under the emergent complexity displayed in a temperate woodland is fairly easy to predict from knowledge of the tree species, climate and geology but the local scale variability of soil properties, driven by chaos, may appear almost random.

Embedded system behaviours may also explain the remarkable fact that many emergent geomorphic patterns are the same, irrespective of the spatial scale. Aerial photographs of ripple beds and dunefields may be indistinguishable without an absolute distance scale. Stream networks, as measured by a stream-ordering parameter, such as Horton's bifurcation ratio, are often statistically similar whether viewed at the scale of the whole drainage basin or a small sub-basin. Many other geomorphic features show this so-called scale-invariance or fractal geometry (Figure 110a), which is easily identified by a power law (straight-line) relationship in a log–log plot of the variable and spatial scale. The behaviour of scale-invariant phenomena has been studied under the heading of self-organized criticality (SOC). This term was used to explain the response of a model sand pile to continuous additions of sand from above. The pile maintains a characteristic form (at the macrolevel) through losses of sand by avalanches (negative feedback at the microlevel), whose distribution over time (either size or frequency) conforms to a power law. The power law distribution means that it is possible to predict the overall



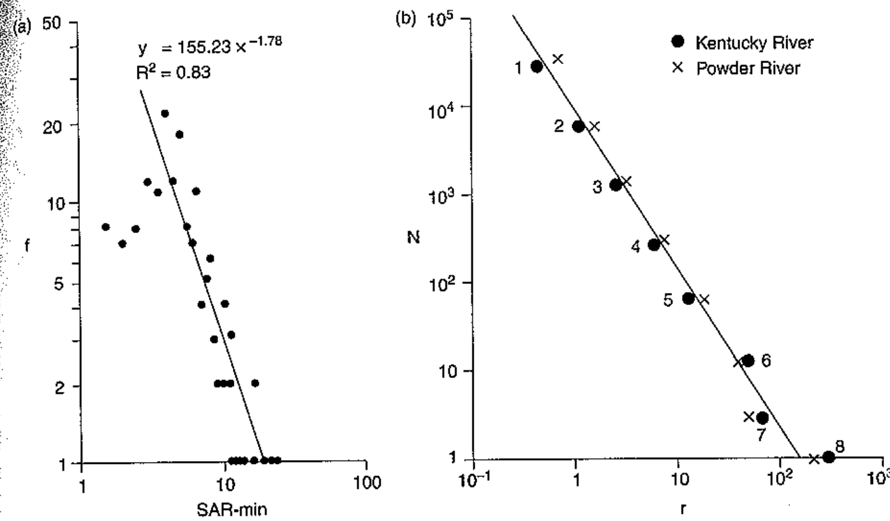


Figure 110 Scale-invariant phenomena in geomorphic patterns and processes. (a) Dependence of the number of streams N of various orders 1–8 on their mean length r for the Kentucky River basin in Kentucky and the Powder River basin in Wyoming (Turcotte 1997); (b) inverse power law relationship between the minerogenic sediment accumulation rate SAR-min and its frequency f for a mid-Holocene lake sediment record of catchment erosion at Holzmaar, Germany, indicative of self-organized criticality (Dearing and Zolitschka 1999)

properties of the system, such that there will be fewer large avalanches than small ones, but the details of timing and size of the next avalanche are unpredictable. A theory has developed that argues for systems of all kinds to evolve to critical states where they may similarly respond to small perturbations in a disproportionate manner, but importantly will evolve back to the original state (Bak 1996). Evidence for SOC now exists for many real spatial patterns, such as river networks and forest fires, and also real time series (Figure 110b), such as earthquakes, landslides and river sediment transport (e.g. Dearing and Zolitschka 1999). Thus, many geomorphic landscapes and landforms may be viewed at one scale as complex and ordered non-linear systems, emerging from the evolution of chaotic processes at another. Chaotic processes may produce highly variable or apparently random forms at a small scale, but through constraints imposed by the nature of the environment (e.g. geology, climate) may produce identifiable and self-similar patterns at large spatial scales, and these may exist in critical states.

Geomorphic systems also respond to external forces, like climate and human activities. The nature of the response depends on the force and the condition of the system and these may vary from direct and reversible to time-lagged and irreversible (Figure 111). Knowing the NSD dynamics that underlie a system may help to define the resilience or sensitivity of a geomorphic system. For example, how close a non-linear landscape system lies to a bifurcation point, as defined by the mathematical representation of that system, helps define its sensitivity to external impacts: the closer it is, the more likely is the system to be driven along a new irreversible trajectory towards an alternative steady state (Thornes 1983).

The implications of NSD in geomorphology are profound. We may have to accept that some geomorphic outcomes are unpredictable and strongly contingent on a landscape's history. It may be that reductionism as a methodology is unlikely to afford extrapolation of mathematical rules to large scales (Lane and Richards 1997). There may be a strong case for adopting some form of 'scientific

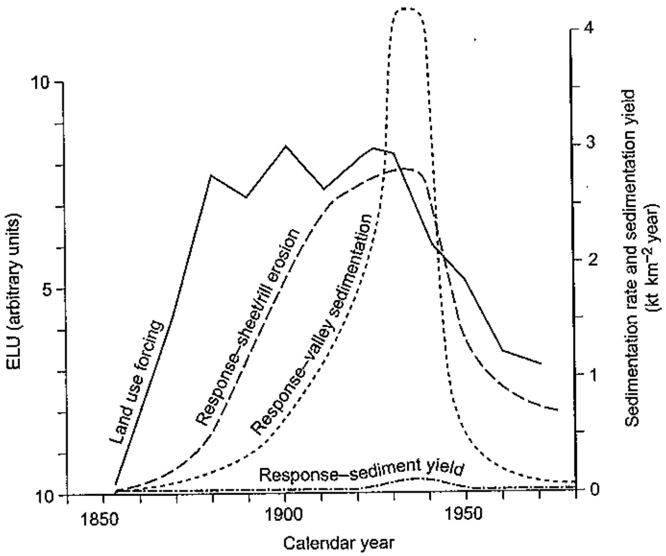


Figure 111 Non-linear relationships between forcings and geomorphic responses in space and time. Erosive land use change ELU in the Coon Creek basin, Wyoming, forces lagged temporal responses in hillslope sheet and rill erosion, main valley sedimentation, and catchment sediment yield (Trimble and Lund 1982; Wasson and Sidorchuk 2000)

realism' as the correct methodology, where we accept the existence of non-observable phenomena, structured and stratified systems with emergent properties, contingent relationships and prediction based on probabilities, while rejecting a belief in direct and enduring relationships between cause and effect (Richards 1990). Paradoxically, overcoming the difficulties of mapping NSD concepts and ideas onto real landscapes may be achieved through the use of simple computational models with simple rules. This 'new kind of science' (Wolfram 2002) uses cellular automata, where grids of interacting cells each containing a set of simple rules are updated at subsequent time steps in order to simulate the evolution of the system. Landscape models based on cellular grids and simple equations for sediment and water flows (e.g. Coulthard *et al.* 2000) show much promise in being able to simulate the spatial and temporal development of complex geomorphic forms with realistic non-linear behaviour. These may prove to be the most efficient means for simulating how non-linear landscapes will really respond to future combinations of climate and human activities.

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JOHN DEARING

# NOTCH, COASTAL

Notches can be cut at the cliff foot by waves, but they are generally poorly defined in fairly homogeneous rocks, and locally restricted to geologically favourable locations in more variable rocks. Notches, typically between 1 and 5 m in depth, are most common and best developed on tropical limestone coasts, where low tidal range concentrates the erosional processes. The formation of notches throughout the tropics is generally attributed to chemical or biochemical CORROSION, or to biological grazing and boring (see BORING ORGANISM), especially in sheltered locations. Nevertheless, ABRASION and other forms of mechanical wave erosion contribute to their formation in some areas. As there is little agreement over the level, or levels, that tropical notches are developing today, the occurrence of double or multiple notches in some places has been ascribed to changes in relative sea level, intermittent tectonic activity, variable rock structure and lithology, and the effect of organisms and other notch-forming mechanisms operating most efficiently at different elevations.

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ALAN TRENHAILE



# NUÉE ARDENTE

A French term most frequently translated into English as a 'glowing cloud', to refer to a pyroclastic flow (see PYROCLASTIC FLOW DEPOSIT) or surge from a volcano. First used by Lacroix (1904) to describe the pyroclastic clouds erupted from Mt Pelée, Martinique, on 8 May 1902 and subsequently. They destroyed the town of St Pierre killing most of its 27–28,000 inhabitants in the worst volcanic disaster of the twentieth century, as measured by loss of life. The term 'ardent' was originally used strictly to refer to 'very hot, burning or scorching' rather than 'glowing', and specified that such clouds were not incandescent at night, except close to the crater (Tanguy 1994).

Nuée ardente has come to be used as a general term for a hot, turbulent, self-expanding gaseous cloud with its entrained particles of rocks and ash, which may be incandescent, that descends the flank of a volcano at high or exceptionally high velocities. In the high velocity flows the denser, lower part hugs the ground surface and becomes strongly controlled by the pre-existing topography. This portion usually forms the bulk of a resultant deposit. Such deposits show massive coarse bouldery facies in channels or the axis of flow, and usually grade to sandy deposits both laterally and upwards. There is evidence of searing heat for only a few brief minutes, but if objects are buried in the deposit they show signs of being cooked. The lighter, upper part of the flow, comprising hot gases and ash particles, rapidly expands upwards as a dark, towering cloud to many kilometres in height. This cloud may spread over a much wider area than the dense lower flow, resulting in a widespread coeval ash.

In exceptionally high velocity surges, a blast of gases and entrained particles travels outward from the volcano, usually independent of the topography, removing most structures and trees in its path. Vertical surfaces become 'sandblasted', the bark of trees may be stripped, trees may be snapped and removed laterally for large distances, rocks may become embedded within materials they impact against, buildings may be razed with their contents twisted or carried from the scene. Resultant deposits show a high degree of fluidity (a low concentration fluid) of the gaseous mixture with cross-stratification, antidunes, highly irregular erosion breaks, considerable lateral variability in lithology and thickness, and frequently charcoal.

Nuée ardente has been used in a broad sense to encompass all pyroclastic flows and surges and in a restricted sense to refer specifically to small volume monolithologic block-and-ash flows generated by the collapse of actively growing lava domes or lava flows on steep terrain (Pelean eruptions).

Nuées ardentes are one of the most feared volcanic hazards. It is not simply the destructive energy of the cloud but the searing heat of gases and ash particles that make them so lethal to life and property.

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VINCENT E. NEALL

## NUNATAK

Nunatak is an Inuit word referring to a mountain top that protrudes above the surface of a GLACIER or ice sheet. Such summits are subject to intense frost weathering but escape glacial erosion. On relatively flat surfaces this often results in the formation of autochthonous blockfields (see BLOCKFIELD AND BLOCKSTREAM), whilst sharper peaks and ridges can become broken along joints to produce sharp ARÊTEs and pinnacles. In areas that were glaciated in the past, the difference in weathering and erosion can be identified as a 'periglacial trimline' that separates glacially eroded terrain from higher ground that retains evidence of prolonged weathering.

A range of simple measures have been used to quantify the difference in degree of rock surface weathering above and below proposed trimlines, including rock surface roughness, joint depth and surface hardness as recorded using a SCHMIDT HAMMER (McCarroll *et al.* 1995). One of the most reliable indicators of past nunataks is the presence of the clay mineral gibbsite at the base of the soils. Gibbsite, an aluminium oxide, is an end product of the weathering of silicate minerals and in mountainous environments in the extra-tropics is thought to represent a long period of *in situ* weathering. Cosmogenic isotope exposure age dating (see COSMOGENIC DATING) can also be used to test the hypothesis that summits escaped glacial erosion and to date the exposure of glaciated bedrock and ERRATICS (Stone *et al.* 1998).

The identification of 'palaeonunataks' using geomorphological and dating evidence provides field evidence with which to test and improve models of past ice sheets in formerly glaciated areas such as the British Isles, northern Europe and North America (Ballantyne *et al.* 1998; McCarroll and Ballantyne 2000). It has also been argued that nunataks in these areas may have provided refugia for plant species to survive glaciations and then re-populate when the glaciers receded. In some cases this would help to explain the rather patchy occurrence of some species, though conditions would have been extremely harsh (Birks 1993).

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DANNY McCARROLL

# O

## ORGANIC WEATHERING

Organisms and organic compounds play a wide range of roles in both enhancing and retarding rock and mineral weathering processes in most environments. Indeed, as Reiche (1950: 5) recognized, weathering involves 'the response of materials which were at equilibrium within the lithosphere to conditions at or near its contact with the atmosphere, the hydrosphere, and perhaps still more importantly, the biosphere'. Thus, by definition, weathering in most places operates in a non-sterile environment and biological processes and influences need to be taken into account if we are to understand weathering fully. The term organic weathering is not widely used by geomorphologists, with biological weathering more commonly found in the literature, but both have a similar broad definition as the suite of biological weathering processes and indirect biological influences on weathering. Bioerosion is an allied term, referring to the erosive activity of organisms especially on bare rock surfaces, and there is a spectrum of such organic influences on weathering and erosion. Many books refer to a three-fold classification of weathering into physical, chemical and biological processes, but it is perhaps simpler to keep the classic division into physical and chemical weathering process groups, acknowledging that many processes within these groups can be seen to be biophysical or biochemical in nature.

Although there was a flourishing of interest in biological weathering in the late twentieth century, as analytical and microscopical techniques became available which permitted closer study of the rock:organism interface, interest in the possible roles of organisms and organic compounds in weathering is nothing new. Several late nineteenth-century workers carried out pioneering experiments on the role of plant seedlings (Sachs 1865), lichens (Sollas 1880) and organic acids (Julien 1879) in weathering common rocks and rock-forming minerals. However, there was little general assessment of the overall nature, rate and importance of biological weathering.

Research into organic weathering has tended to focus on a number of key types of organism or organic influence. First, there has been a huge concentration of effort into understanding the role played by organic acids and other organic compounds within soils on the weathering of minerals (Drever and Vance 1994). Second, there have been a number of studies on the role of plant roots in weathering both rocks and minerals, illustrating the important uptake of many elements by such processes (Kelly *et al.* 1998; Hinsinger *et al.* 2001). Third, a few studies have been made of the role of animals in weathering, especially their contribution to the decomposition of organic material leading to the production of organic acids. Fourth, there have been many studies on the role of lichens in the weathering of rock surfaces. These studies illustrate the complexity of roles played by organisms in weathering, with biophysical and biochemical lichen activity recognized, as well as a bioprotective role in some cases where they retard the action of other weathering processes and bind the rock surface together. Different lichen species play different roles, with some species being highly biodeteriorative and others not. For example, a recent study by Robinson and Williams (2000) in the Moroccan High Atlas show accelerated weathering of sandstone by *Aspicilia caesiocinerea* agg. producing notable scars on the rock surface. Finally, there have been many studies of the roles played by micro-organisms (notably algae, fungi and bacteria) in weathering. In most environments, combinations of different processes and organisms are involved producing a complex biological imprint on weathering. Thus, for example, the large percentage of the terrestrial landsurface covered by soils will experience weathering conditioned by soil organic acids, plant roots, animal decomposition and the many micro-organisms that inhabit soils (e.g. symbiotic mycorrhizal fungi have been shown by Jongmans *et al.* 1997 to bore into feldspars in soils, thereby allowing uptake of Ca and Mg by plant roots). Similarly, most bare rock surfaces are not actually bare at all but are coated with a diverse community of micro-organisms and lower plants which play a range of roles in weathering.

Biophysical weathering often involves the creation of stresses by the expansion and contraction of organisms or parts of organisms. Lichens, for example, can absorb a vast amount of water leading to huge expansion on wetting and a concomitant shrinkage on drying. Crustose lichens, which grow very closely attached to rock surfaces, can cause much weathering to the underlying rock. Fry (1927) and Moses and Smith (1993) provide experimental evidence of the physical weathering caused by such wetting and drying on a range of rock types. Natural processes of growth and decay of rock-dwelling organisms or parts of organisms can also cause biophysical weathering. Some crustose lichen thalli, for example, start to peel away from the rock surface at the centre as they senesce, producing patches of intensive weathering. Plant roots may also force their way into cracks and joints, with large pressures exerted during growth sufficient to induce weathering. Grazing and mechanical burrowing of animals can also produce physical weathering, although this is usually categorized as a form of bioerosion.

Biochemical weathering includes a host of individual processes, with the production of carbon dioxide, organic acids and chelation (often involving organic acids) being particularly important. Within soils, organic by-products can provide a key control on weathering processes, especially within the rhizosphere. Many organic acids, such as humic, fulvic, citric, malic and gluconic, have been shown to be capable of weathering a range of substrates. Several types of micro-organism have been found to be capable of boring into suitable substrates (such as rock, corals and mineral grains) through chemical means. The exact mechanism by which different micro-organisms bore has been much debated, but extracellular acids and other substances probably promote chemical decomposition of minerals. The production of a network of near-surface boreholes can encourage further weathering by other (often inorganic) processes as they weaken the surface and increase the reactive surface area. Biochemical weathering by lichens involves the production of carbon dioxide, oxalic acid and the complexing action of a range of sparingly soluble lichen substances. Lichens such as the highly biodeteriorating *Dirina massiliensis* forma *sorediata* produce oxalic acid which reacts with calcium carbonate within rocks to produce calcium oxalate (Seaward 1997).

Another important aspect of organic influences on weathering is bioprotection. In this case, organisms do not play an active role in encouraging weathering, but rather play a passive role. The very presence of a cover of lichens or biofilm, for example, protects the underlying rock surface from extremes of temperature (thus reducing the potential for damaging freeze–thaw weathering for example). Furthermore, the lichen or biofilm can also act as a sponge soaking up incident rainwater and providing a chemical buffer for the underlying surface, thus reducing the potential for chemical weathering.

Identification of the various organic weathering processes and influences is a challenge; an even greater challenge revolves around trying to quantify their effect and identify their contribution to overall weathering rates. Some measurements of biological weathering rates are presented in Goudie (1995) derived from detailed empirical studies in the field as well as experimental studies. Many biological processes have proved very difficult to measure in the field with currently available techniques (e.g. physical weathering by plant roots) and experimentation is also challenging where growing organisms are involved. Although some organic weathering processes operate at a fast rate and are quite dramatic, their action is often highly localized in time and space. Thus, for example, biodeteriorating lichens can act at a very fast rate, but they are often patchily distributed over a rock surface, so their net effect may be reduced. Furthermore, over time community dynamics will influence the species present on any surface, limiting the net contribution of biodeteriorating species.



Despite the wide range of studies on a variety of organic weathering processes and influences, there is still uncertainty about their general importance and contribution to weathering in different environments. Viles (1995) hypothesized that on bare rock surfaces biological weathering would increase in 'hostile' environments, and that less bioprotection would take place. The reasoning behind this assertion is that in hostile environments organisms extract more nutrients, water and shelter from the rock itself, producing a weathering effect as they do so. In more benign environments organisms tend to have a less close contact with the surface, and their net role will be a protective one. Thus, hot desert, cold tundra and coastal environments should be characterized by high rates of biological weathering, in comparison with humid temperate and tropical ones. However, many harsh environments are also characterized by slower biological growth rates which may complicate this pattern.

Even if biological weathering on bare rock surfaces can be seen to be more intense in some environments rather than others, this does not necessarily imply that it will be the dominant series of processes there. In hot arid areas, for example, although lichen weathering can be spectacular, heating and cooling and salt weathering can also be highly effective and operate at a much faster rate than can slow-growing lichens. Considering the soil-covered landscape it is probable that the reverse applies, with higher rainfall and temperatures encouraging organic growths within the whole ecosystem, thus enhancing the production of organic acids and increasing both the overall rate of weathering and the contribution of organisms to weathering.

In some instances, recognizable small-scale landforms can be produced by biological weathering processes. For example, lichen fruiting bodies create small pinhead-sized pits in rock surfaces and various authors have identified larger pitting and grooving as being organically produced (e.g. Danin and Garty 1983; Robinson and Williams 2000). On limestone surfaces such features may be called BIOKARST. It has proved difficult to establish convincing process-form links in many cases, as there are a multiplicity of ways in which similar landforms at this scale can be produced.

Future challenges for geomorphologists in understanding organic weathering include the need to provide quantitative comparisons of sterile vs. organically mediated weathering rates in different environments, and the need to provide a broader assessment of the overall role of organisms and organic by-products in weathering and sedimentation. Finally, scientists such as Robert Berner have suggested that biotically enhanced weathering has played a major role in the global carbon cycle over long time spans (Berner 1992), and geomorphologists can play a key role in providing reliable empirical data on biological weathering rates and their spatial and temporal variability in order to test such models.

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HEATHER A. VILES

## ORIENTED LAKE

Oriented lakes are subparallel, elongated lakes, which commonly occur in extensive clusters. Many of these clusters, especially in bedrock, are the result of processes antecedent to lake formation. For instance, orientation on the Canadian Shield is commonly associated with glacial fluting. However, there are belts of oriented lakes covering thousands of $km^2$ in unglaciated areas of the sandy Arctic coastal plains of Russia, northern Alaska and north-west Canada. They also occur in non-permafrost areas, such as the Atlantic coastlands of Maryland, the Carolinas and Georgia, USA. Two forms of oriented lake have been recognized: the most common are elliptical, but rectangular shapes occur in the Beni Basin of northeastern Bolivia, and the Old Crow and Bluefish Basins of northern Yukon Territory, Canada.

In North America, the elliptical lakes occur outside the glacial limits, at sites where there is no evidence of glacial deposition. Some occur in postglacial marine terraces. Well-known examples are the Carolina Bays of the southern Atlantic coast, and the lakes near Liverpool Bay, NWT, and Point Barrow, Alaska, on the western Arctic coast. In permafrost environments, the lakes are in depressions formed by thermokarst processes, which have been elongated during their development. The lakes range in size over several orders of magnitude, from ponds with long axes of less than 30 m, to water bodies of over 1,500 ha. The lake-size distributions are skewed, with the mean being less than 250 ha. Along the western Arctic coast, the mean length to width ratio of the lakes is about 1:7. The major axes of the lakes are aligned perpendicular to the prevailing winds, and, in the case of the lakes near Liverpool Bay, the standard error of mean orientation is less than 3°.

Several theories have been advanced for the causes of lake orientation. Many of these, including bombardment by meteorite showers, effects from upwelling of artesian springs or the action of fishes hovering while spawning, have been discredited. However, consideration of the effects of wind action perpendicular to the long axes of the lakes has been supported by field data and laboratory experiment.

The hydrodynamic theory proposes that winds blowing across a lake establish a two-cell current circulation within the water body, with water returning to the windward shore around the ends of the lake (Rex 1961). The maximum littoral drift, and associated erosion, occurs at the ends of the lakes, where the angle between waves propagating in deep water and a line perpendicular to the shore is 50°. A similar circulation has been measured in large lakes of the Alaskan coastal plain (Carson and Hussey 1962), and has been reproduced at laboratory scale by blowing air across a 2 m square box containing a scale model pond. The model pond, initially circular, became elongated in the process. The equilibrium form of a finite shoreline in readily transported, unconsolidated material is cycloidal, which corresponds to the approximate shape of the oriented lakes along the western Arctic coast. The orientation of the lakes in the western Arctic is consistent with the hydrodynamic theory, when applied using the prevailing wind regime of the region (Carson and Hussey 1962; Mackay 1963).

The rectangular form of the oriented lakes in the Old Crow and Beni basins has been associated with aspects of bedrock structure propagating through the overlying sediments (Allenby 1989), but the explanation has not been verified. In northern Yukon, the lakes are THERMOKARST features that have developed in sediments up to at least 150 m thick, in which permafrost is present in the upper 60 m. The permafrost impounds the lakes. In Bolivia, the sediments are similarly thick, but lake levels are maintained by a high water table close to the surface.

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C.R. BURN



## OROGENESIS

Orogenesis is the building of mountains by the forces of PLATE TECTONICS. Driven by the internal heat of the Earth, motions of lithospheric plates produce changes in crustal thickness structure that result in vertical motion of the topographic surface. It is this motion that is responsible for creating impressive mountain landscapes that have been the inspiration of so much geomorphology.

Mountains are built slowly over geologic time as an accumulation of CRUSTAL DEFORMATION. Rates of mountain building are quantified as the rates of vertical motion of rock with respect to the geoid (rock uplift), the surface with respect to the geoid (surface uplift), or rock with respect to the surface (exhumation, see also DENUDATION) (England and Molnar 1990). The relationship, rock uplift = surface uplift + exhumation, is widely adopted. Surface uplift describes topographic growth and creation of the positive landforms of mountains. Long-term surface uplift is difficult to measure, but is approximated using the altitude dependence of fossil organisms or displacement of features with respect to eustatic sea level (Abbott *et al.* 1997). Short-term surface uplift can be measured using geodetic techniques (e.g. GPS and synthetic radar interferometry). Relative surface uplift can be constrained using geomorphic markers, such as river terraces (see TERRACE, RIVER) or erosion surfaces. Exhumation relates to erosion, which is critical for accommodating plate motion by transferring mass from thickened mountain belts to adjacent basins. It is generally inferred from rock cooling histories (see DENUDATION CHRONOLOGY and FISSION TRACK ANALYSIS), basin sedimentation records or COSMOGENIC DATING.

Active orogens may form at rates of rock uplift of $0.01–10\,mm\,yr^{-1}$, and a rate of $\sim 1\,mm\,yr^{-1}$ is representative of many actively growing mountains around the world. Mountain building tends to be stable over millions to tens of millions of years, such that major episodes of orogenesis are commonly given formal names (e.g. the Alpine Orogeny). The timescale of mountain building is short enough, however, that changes in plate motion and global climate throughout the latest Tertiary and Quaternary have had noticeable effects on most orogens.

Orogenic belts are most common along ACTIVE MARGINS, such as the arcuate mountain belts of the 'Ring of Fire' along the continental rim that surrounds the Pacific plate. The characteristics of these mountain ranges depend on the type of plate boundary. The majority of mountain uplift is produced by convergent tectonic motion, where two or more plates collide and increase crustal thickness. One setting in which this occurs is Cordilleran-type orogens, well represented by the Cascades of northwestern North America or the Andes. These mountains stretch above subduction zones and are produced by permanent convergent deformation of the overriding continental plate and thermally driven buoyancy of a magmatic arc. Their anatomy includes coastal ranges near the accretionary complex of the subduction zone, lines of volcanoes that are separated from the coast by elongate valleys, and highlands that rise above foreland fold and thrust belts that verge towards the continent interior. Width of these highlands is largely dependent on the dip of the subducting oceanic lithosphere.

A second setting of convergent mountain building is continental collision, typified by the active collision of India and Asia. India has underthrust beneath Asia over the past 50 Ma, resulting in uplift of the High Himalaya, Tibetan Plateau and associated mountains that penetrate thousands of kilometres into the interior of Asia (Hodges 2000). Several thousand kilometres of plate convergence has been accommodated by a combination of orogenic crustal thickening and lateral escape of microplates via strike-slip faults. That continental collision is the most effective mechanism of mountain building is evident. Half of all mountain peaks worldwide that rise above 7.5 km elevation occur in the Himalaya, while all of the remaining half are associated directly with the India–Asia collision. The crystalline core of the Himalaya has also experienced ~10–20 km of DENUDATION in the Neogene at rates locally as high as $1\,cm\,yr^{-1}$, leading to rapid deposition in the Bengal and Indus fans (Searle 1996). In Tibet, arid conditions and internal drainage basin geometry have hindered erosional exhumation, leading to formation of an orogenic plateau above the under-thrust Indian plate. Geodynamic processes limit the plateau's elevation to ~5 km, via lower crustal flow where strength is exceeded by gravitational load (Royden *et al.* 1997).

Mountains are also built at divergent and transform plate boundaries and within continental interiors. Crustal extension via normal faulting leads to tilting and uplift of large hanging-wall blocks and results in characteristic basin and range topography (e.g. the western USA). Strike-slip plate boundaries may produce narrow zones of orogenesis where plate motions are somewhat oblique in a convergent (transpressional) or divergent (transtensional) sense. The components of non-strike-slip motion along such boundaries may be accommodated by reverse or normal faulting, so that similar but laterally confined mountain systems result (e.g. the Transverse Ranges along the San Andreas fault). Mountains within continental interiors may represent earlier stages of continental deformation or the effect of geodynamic processes. Dynamic topography on cratons can occur where compositionally or thermally induced density contrasts occur in the sub-lithospheric mantle.

Although a mountain may owe its origin to tectonic construction, mountain landscapes are dominated by erosional processes. Mountain topography consists of valleys and hillslopes shaped by erosional agents (e.g. mass wasting, glacial erosion, fluvial erosion). Agents of erosion are poised to reduce the terrestrial surface to low RELIEF, such that tectonic orogenesis is critical for maintaining topographic variation above the mean elevation of the continents. The act of mountain building hence has important effects on the dynamic processes of erosion itself, both directly (e.g. changes in BASE LEVEL) and indirectly (e.g. the effect of mountains on local climate). Because of this erosional character of mountains, topographic character does not always discriminate orogens that are actively forming from those formed by prior plate motion.

Mountainous topography may linger for more than one hundred million years after cessation of active tectonic deformation (e.g. the Appalachian Mountains). Rock uplift and erosion continue as long as an orogen contains a thickened crustal root, that must be removed by DENUDATION. Topography in ancient mountains is maintained, even after erosion has removed many kilometres of rock, because of ISOSTASY. The ductile nature of the upper mantle permits adjustment to gravitational loads over timescales of $\sim 10^3$ yr. The increased thickness of crust beneath orogens is more buoyant than the mantle rocks that lie beneath adjacent crust, such that the topographic surface is higher than the surrounding region. As erosion removes mass from the mountains, the crust rebounds to remain in gravitational equilibrium. The magnitude of rebound is proportional to the ratio of crust and mantle densities, such that mean elevation decreases much more slowly than the rate of regional denudation. Isostatic rebound can even produce the uplift of peaks where mean elevation is decreasing, where valleys are incised faster than interfluves erode.

Although inactive mountain ranges are still referred to as 'orogens', they experienced their tectonic orogenesis in the geologic past and are thus distinct from mountains presently rising along active plate margins. Erosion rates of ancient mountain belts may become quite slow, such as average rates in the Appalachian Mountains of 0.02–0.04 mm $yr^{-1}$ (Mills *et al.* 1987). Denudation this slow is essentially weathering-rate limited, yet it is enough to reduce crustal thickness over long periods. Variations in the geomorphic system, such as climate change, may impose upon the stagnant erosional setting of ancient mountains and force readjustment of erosional processes. This often leads to incision and REJUVENATION of topography.

The character of topography itself has traditionally been used to interpret the surface uplift and exhumation history of mountain belts. This is one goal of landscape evolution studies, which seek to define cyclic changes in landforms or topographic 'maturity' through stages of orogenic and post-orogenic development (see CYCLE OF EROSION). However, this approach requires tenuous assumptions about the relationships between topographic parameters and uplift and exhumation. Studies have shown that many aspects of topography, such as RUGGEDNESS, DRAINAGE DENSITY and hypsometry (see HYPSOMETRIC ANALYSIS), are dependent on the erosional resistance of rocks, the nature of local climate and the individual processes of the dominant erosional agent (e.g. Willgoose and Hancock 1998), such that direct interpretation of orogenic history from topographic parameterization is dubious. One parameter that has been demonstrated to correlate with exhumation rate is slope or short-wavelength relief. Slope is almost synonymous with erosion rate, as rates of all erosional processes increase with slope and slope must increase in areas of rapid surface uplift due to ever-growing gravitational instability. Nonetheless, the clues to orogenic evolution do not lie entirely in topographic parameterization using DIGITAL ELEVATION MODELS.

Because the geomorphic and geodynamic processes that shape orogenic belts are complex and occur over very long scales of time and space, the topographic evolution of mountain landscapes is difficult to study comprehensively with physical experimentation or direct observation. For this reason, the approach of numerical modelling has become important (see MODELS). Models can quantitatively test how elements of landscapes should evolve under a given set of conditions, based on the use of erosion laws derived from previous research. Models can be constructed as grids of cells with set boundary conditions, such as the distribution of tectonic uplift or base-level change, frequency and magnitude of precipitation events, and rheology of eroding materials. Erosion laws, such as stream power bedrock incision or diffusive hillslope transport, can be run under different conditions. The result is a representation of the long-term landscape evolution of a hypothetical setting, and can be instructive with respect to the role of boundary conditions or specific processes (Burbank and Anderson 2001). However, these models are limited by present degree of understanding of individual erosional processes and the difficulty of capturing real-world complexity.

The significance of erosion goes beyond shaping the landscape of orogens. Erosion can itself influence tectonic processes. For example, erosion modulates surface slope in deforming thrust wedges that can in turn affect deformation (e.g. critical taper wedges; Dahlen *et al.* 1984). Concentrated erosion in parts of an orogen also control trajectories of crustal motion, and thus influence deformation partitioning (e.g. the Southern Alps, New Zealand; Koons 1989). This in turn causes faster tectonic uplift and increases gravitational potential energy, leading to more rapid erosion (i.e. positive feedback loop). Eventually, erosion and tectonic rock uplift rates may become balanced in a steady state of mass flux. In the north-west Himalaya, for example, a steady state has been achieved where topography is sufficiently rugged as to result in bedrock landsliding that keeps pace with base-level induced incision driven by tectonic uplift (Burbank *et al.* 1996).

The most important external influence on erosion landscape evolution and mountain building may be climate. Rates of erosion tend to correlate with precipitation. Spatial concentrations of precipitation due to the orographic effect of topography can lead to concentrated erosion (e.g. Irian Jaya fold belt; Weiland and Cloos 1996). This is a core element of models of coupled erosion and tectonic evolution of orogens (Willett 1999). Latitudinal, altitudinal and temporal variations in climate also determine where glacial and fluvial erosion dominate. Glacial erosion is thought to be more effective, and can force relief production via valley incision and intense erosion at equilibrium line (see EQUILIBRIUM LINE OF GLACIERS) altitudes that can 'buzz-saw' mountain landscapes (Brozovic *et al.* 1997). The effects of climate may be so important that increases in erosion and sediment production in many mountain systems worldwide may have been caused by the onset of glacial climate ~4 Myr ago (Molnar and England 1990).

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JAMES A. SPOTILA

## OUTBURST FLOOD

Outburst floods or jökulhlaups, are high magnitude, low frequency catastrophic (see CATASTROPHISM) events involving the sudden release of glacial meltwater (see MELTWATER AND MELTWATER CHANNEL) stored in subglacial (see SUBGLACIAL GEOMORPHOLOGY) reservoirs or ice-dammed lakes. The volume of water discharged is usually orders of magnitude greater than normal flow, with modern outburst floods estimated up to 2,000,000 $m^3 s^{-1}$ and Pleistocene-aged flood peaks estimated at 21,000,000 $m^3 s^{-1}$. However, most discharges usually measure hundreds to thousands $m^3 s^{-1}$. Importantly, outbursts leave characteristic erosional and sedimentary signatures, which enables palaeo-outburst flood reconstructions.

Hydrographs with a slowly rising limb followed by a rapidly falling limb characterize outburst floods through glaciers. The steadily increasing discharge reflects a greater efficiency in subglacial water routing, aided by a positive feedback interaction of channel enlargement caused by melting and abrasion. The increasing discharge may overwhelm the subglacial drainage network, and it is not uncommon for water to burst from SUPRAGLACIAL positions at the ice margin. The rapid decrease in discharge is a reflection of either a drained reservoir, or rapid tunnel closure caused by ice deformation or collapse.

The water source is variable, dependent upon the location of the glacier and the reservoir. In volcanic regions, such as Iceland, high geothermal gradients and subglacial volcanic eruptions lead to the rapid production of meltwater and cyclic outburst floods. For example, Grimsvötn, beneath the Vatnajökull ice cap in Iceland, drains approximately every six years with discharges up to 50,000 $m^3 s^{-1}$. In non-volcanic areas, the collection and storage of water often takes much longer. In these areas, water production is a by-product of lower geothermal gradients, precipitation, insolation (see INSOLATION WEATHERING) and frictional heat from sliding and deforming ice. Water may be stored in supraglacial, englacial, subglacial or ice marginal positions. Supraglacial drainage is dependent upon connections with englacial or subglacial conduits such as crevasses or MOULINS. The largest known possible reservoirs are subglacial. About seventy subglacial lakes have been identified with radio-echo sounding beneath the Antarctic Ice Sheets. The lakes vary in size from a few $kms^2$ to 14,000 $km^2$ with between 4,000 and 12,000 $km^3$ of stored water.

Outburst floods also develop where proglacial (see PROGLACIAL LANDFORM) dams fail. Commonly in mountainous terrain and during glacial recession, proglacial lakes develop behind moraines or in ice-dammed valleys. Dam failure may be initiated by sudden inputs of water or iceberg calving, and is usually the result of rapid fluvial incision initiated by overflow, internal loss of support, or sapping processes, especially with dams composed of sediment or ice. These hydrographs show a rapid increase in discharge with a slowly falling limb.

Outburst floods may deeply cut canyons into bedrock or sediment, and form extensive outwash plains (sandurs) or discrete gravel bars. Deposits may consist of clast-supported, boulder-gravel sequences greater than 10 m thick, which coarsen upward, and are capped with a fining-upward sequence of gravels, sand and silt. However, boulder-gravel deposits described from Pleistocene-aged subglacial outburst floods tend not to show this fining-upward sequence. In backwater areas, rhythmically deposited couplets up to 15 m thick of fine gravel and sand with rip-up clasts and boulders (also called eddy bars), indicate pulsed flow with high sediment concentrations. HYPERCONCENTRATED FLOWS and DEBRIS FLOWS are commonly associated with outburst floods. Giant current ripples, deposited by the Pleistocene-age Lake Missoula floods that scoured out the Channeled SCABLANDS in central Washington, USA, have wavelengths up to 125 m and are up to 7 m high. These BEDFORMS were also instrumental in the development of expansion and pendant bars composed of foreset-bedded gravel. Such bars are also described from the Interior Plains of North America where outburst floods scoured channels across the prairie surface with discharges estimated at $10^5$ $m^3 s^{-1}$. The spillway geometry consists of an inner channel 25–100 m deep and 1–3 km wide, and an upper-scoured zone as wide as 10 km. On the upper scoured zone, channels have an anastomosing (see ANABRANCHING AND ANASTOMOSING RIVER) pattern with residual streamlined hills that resemble DRUMLINS, and boulder lags are common. Within the inner channel, streamlined hills are rare, gravel bars may be found in landslide alcoves or as point bars, and large fans (expansion bars) are found where the outburst flood entered a basin.

Outburst floods have also been used to explain Pleistocene-age subglacial bedforms. In this hypothesis, large subglacial reservoirs are released as sheet flows scouring the subglacial landscape, leaving erosional remnants such as drumlins, fluting, scabland and hummocky terrain up to 100 km wide. Some drumlins and Rogen moraines are explained as sediment moulds from cavities cut into the ice by the outburst floods. Tunnel channels, that often closely resemble spillway channels, are associated with the subglacial bedforms, and develop as the sheet flow collapsed into discrete channels.

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SEE ALSO: geomorphologial hazard; glacier; glaciofluvial; palaeoflood; palaeohydrology

TIMOTHY G. FISHER



## OVERCONSOLIDATED CLAY

Overconsolidated clays are those that have been highly compressed by burial. The burial may be by superincumbent sedimentation or by short-term loading, but glacial ice is the usual agent. Water is expelled from the clay as it assumes a denser packing. If the clay is subsequently unloaded, perhaps by the ice melting, the clay may become fissured and jointed. An example can be given from the glacial sediments of middle England.

Fissure studies in glacial lake clay and tills at Happisburgh and Cromer show well-defined fissure patterns which can be related to the glacial history of the area and the subsequent erosion history. The presence of fissures influences the strength, consolidation, and permeability characteristics of the clay; the strength along a striated fissure plane is almost reduced to its residual value. The coefficients of consolidation and permeability are significantly increased in the presence of fissures. Attention is drawn to the well-developed fissure systems in tills which have commonly been regarded as non-fissured materials.

Such fissured clay presents a lowered strength as a total mass than would be observed by a triaxial test on a small, unfissured sample. Such materials should be analysed in a manner similar to those used in rock mechanics, which consider the state of discontinuities in lowering the bulk strength. Another consideration in overconsolidated clays is that they have generally been sheared past their peak strength by an earlier loading phase. The subsequent performance of the material will thus be dictated by the residual strength, even though laboratory tests might indicate higher peak values.

During the formation of a sedimentary soil the total stress at any given elevation continues to build up as the height of the soil over that point increases.

The removal of soil overburden, perhaps by erosion (perhaps by bulldozer) results in a reduction of stress. A soil element that is at equilibrium under the maximum stress it has ever experienced is normally consolidated, whereas a soil at equilibrium under a stress less than that to which it was once consolidated is overconsolidated. This means that a clay soil whose *in situ* stress is less than the preconsolidation pressure is, regardless of the cause, called overconsolidated. Various geological and landscape factors responsible for causing preconsolidation stress have been recognized. Mechanisms that cause a preconsolidation pressure:

- Changes in total stress due to: (1) removal of overburden, (2) past structures, (3) glaciation.
- Changes in pore-water pressure due to: (1) change in water table elevation, (2) artesian pressures, (3) deep pumping, (4) desiccation due to drying, (5) desiccation due to plants.
- Changes in soil structure due to: (1) secondary compression, (2) environmental changes, such as pH, temperature, salt concentration, (3) chemical alteration due to: weathering, precipitation of cementing agents, ion exchange.
- Changes in strain rate on loading.

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IAN SMALLEY

## OVERFLOW CHANNEL

Ice-dammed lakes are often found in ice-free valleys tributary to glaciers. Meltwaters from snow and ice may be impounded between advancing glaciers and rock walls, between a main valley glacier and a tributary glacier, or by the advance of a glacier in a tributary valley across a valley not occupied by a glacier. Unless water drains beneath the glacier blocking the valley, meltwater accumulates and water levels rise, until the height of the lowest col is reached. At that stage, water overflows either to adjacent proglacial lakes, or to areas free of ice, creating overflow channels. The large volume of water that escapes ensures that the overflow channel is deepened in a comparatively short period of time.

When compared to normal regional drainage patterns, overflow channels are anomalous in terms of position, morphology and size. Characteristically overflow channels are trough-shaped, with flat floor and steep sides, forming an abrupt angle with higher ground above. For many, the longitudinal profile is undulating. Tributary valleys are rare or absent, but if present they display normal river valley morphology. Shorter overflow channels tend to be straight, or nearly so, while larger overflow channels may display a sinuous planform. Overflow channels are sometimes called spillways, although this term is also used for channels created by catastrophic OUTBURST FLOODS following the failure of an ice dam. Most of these channels are now dry, having been abandoned once the ice withdrew and the lakes they drained disappeared. However, on occasions they may be so deepened that the overflow channels retain drainage even after the ice melted and UNDERFIT STREAMS now flow in these channels. Not all anomalous drainage patterns can be explained by overflow, however. RIVER CAPTURE or subglacial drainage, for example, many provide alternative explanations.

Today, proglacial lakes and overflow channels are found in Norway, the Himalayan region, the Rocky and Andes mountains, Baffin Island, Iceland, and particularly Alaska. At the end of the last glaciation, significant volumes of water were impounded around the margin of continental ice sheets. Water levels in these lakes frequently changed configuration, depth and volume because of the interplay between ice margin position, subglacial topography, isostatic rebound and outlet erosion. As water levels changed, often abruptly, water was released into newly opened overflow channels. Thus, complex and extensive overflow channels resulting from late-Pleistocene glacial lakes are found in northern Europe and North America. Sparks (1960) provides detailed descriptions of overflow channels in northern England. Twidale (1968) describes the overflow channel that runs through central Sweden, from Stockholm, through two lakes Mälaren and Vänern and the Göta River, to Göteborg. Teller *et al.* (2002) describe the overflow channels of the largest late-Pleistocene lake, glacial Lake Agassiz, into the Mississippi and Hudson systems.

The term overflow channel has been extended to refer to abandoned channels in a floodplain that may carry water during periods of high flow, or at a dam site to refer to the spillway that can be opened to release lake water when water levels get high enough to threaten the safety of a dam.

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SEE ALSO: outburst flood

CATHERINE SOUCH

## OVERLAND FLOW

Overland flow is the term used for water that flows over the surface of hillslopes. It is important because this route provides the fastest means by which rain falling on hillslopes can reach stream channels. Hence overland flow contributes significantly to the shape of a catchment storm hydrograph. Equally, it may be responsible for high erosion rates on hillslopes. Other terms that are used in this context are sheet wash (see SHEET EROSION, SHEET FLOW, SHEET WASH) and interrill flow, both of which denote unchannelled flow, and RILL flow, which is used to describe overland flow where it becomes concentrated into definable channels on the hillslope. Though interrill flow does not exhibit definable channels, it is common to observe that the flow is not of uniform depth. Instead, the flow converges and diverges around microtopographic obstacles forming anastomosing threads of deeper and faster flow within a layer of water that covers most of the surface. It is generally accepted that the presence of rills indicates that the flow is able to detach and transport sediment, whereas interrill flow, though capable of transporting sediment, does not have the erosive power to detach sediment. Instead, the sediment load of interrill flow is supplied to it by the detaching force of impacting raindrops. Three types of overland flow may be recognized. The first is that which is due to rain falling at an intensity in excess of the rate of infiltration into the soil. This type of overland flow is also termed Hortonian overland flow after R.E. Horton, who first described the process (Horton 1933). The second type of overland flow is termed saturation-excess overland flow, and the third is termed return flow.

### Hortonian overland flow

According to Horton's description of the generation of overland flow, rain reaching the soil can be separated into two parts: one infiltrates into the soil, the other remains on the surface. The rate at which rain infiltrates and its relationship to the rainfall intensity is the basis of Horton's model for the generation of overland flow. Horton developed the equation

$$f = f_c + (f_o - f_c)e^{-kt}$$

to describe the way infiltration would change during a storm, in which $f$ is the maximum instantaneous infiltration rate, $f_c$ is minimum infiltration rate (infiltration capacity), assumed to be a constant for a given soil, $f_0$ is the initial (maximum) infiltration rate (at $t = 0$), $k$ is a constant that varies with soil type, and $t$ is time since the onset of rain.

In general, $f_0$ will be in excess of all but the highest rainfall intensities, so that initially all rain infiltrates. As the rainstorm proceeds, the pore spaces of the soil fill with infiltrated rainwater, cracks in the soil close and fine particles wash into the surface of the soil so that the instantaneous infiltration rate declines through time. Eventually, it may fall to the point where it is below the rainfall intensity, at which time some of the falling rain remains on the surface. The time taken for this to occur is known as the time to ponding and its completion can be recognized by glistening of the surface and the appearance of ponds of water in small depressions on an irregular ground surface. As these depressions fill with water they begin to overtop and interconnect, until the ground surface is covered by a connected series of these pools. The amount of water that is required for this to occur will vary with the surface irregularity of the hillslope, and is known as the depression storage. Once this stage is reached, flow from pool to pool begins to occur throughout the hillslope and water is discharged from the hillslope as overland flow. As the rainstorm proceeds, the rate of infiltration into the soil continues to decline so that the amount of water remaining on the surface increases. As the volume of water at the ground surface increases, so does the discharge from the hillslope. Once the infiltration capacity $f_c$ of the soil has been approximated (assuming rainfall rate is constant), the discharge of water from the hillslope will also begin to level out towards an equilibrium value. The layer of moving water is known as the surface detention. The thickness of this layer and the time delay between the approximation of the soil's infiltration capacity and the achievement of equilibrium runoff will be a function of the runoff hydraulics (depth and velocity) and the length of the hillsope. Runoff hydraulics are, in turn, primarily controlled by the surface roughness of the ground, which determines the frictional resistance it affords to the flowing water. The relationship between water depth and velocity, on the one hand, and frictional resistance, on the other, can be expressed through the Darcy–Weisbach equation:

$$ff = \frac{8gds}{v^2}$$

in which $ff$ is the dimensionless Darcy–Weisbach friction factor, $g$ is the gravitational constant (m s$^{-2}$), $d$ is water depth (m), $s$ is slope (m m$^{-1}$) and $v$ is water velocity (m s$^{-1}$).

Under the Hortonian model for the generation of overland flow it is assumed that flow will be generated more or less simultaneously over entire hillslopes. This is most likely to be the case where soils have very low initial soil moisture at the start of rainfall and/or the rainfall is intense and/or the soil has a very low infiltration rate. These conditions are most commonly met on bare soils (such as cleared agricultural land), in arid and semi-arid environments, and during convective thunderstorms during which peak rainfall intensities of 300 mm hr$^{-1}$ can be attained and many minutes of rainfall at intensities exceeding 50 mm hr$^{-1}$ are not uncommon. Bare soils may have quite low infiltration capacities because of crusting (either biological or mechanical) (see CRUSTING OF SOIL) of the surface. Kidron and Yair (2001) report infiltration capacities as low as 9 mm hr$^{-1}$ on crusted dune soils in Israel, so that Hortonian overland flow can be generated at even quite low rainfall intensities.

Because Hortonian overland flow depends on the relationship between rainfall rate and infiltration rate discharge increases downslope. The ways in which this increase in discharge downslope affects flow width, depth and velocity (the HYDRAULIC GEOMETRY) is highly variable and depend primarily on the characteristics of the hillslope surface (Parsons *et al.* 1996). Both laminar and turbulent flow conditions are present and the flow may vary from laminar to turbulent both spatially and temporally. Horton (1945) used the term mixed flow to characterize this condition.

The downslope increase in discharge may be accompanied by a change from wholly unchannelled flow on the upper part of the hillslope, to a mix of channelled (rill) and unchannelled (interrill) flow on the lower part. The emergence of eroded channels under the operation of overland flow led Horton to term the upper part of hillslopes without rills the belt of no erosion. It is, however, not correct that no erosion takes place in this zone: simply that detachment in this zone is accomplished by raindrops and is spatially diffuse, as it is everywhere in interrill flow. Soil detachment by falling raindrops is controlled by the kinetic energy of the rainfall, but is diminished as the depth of interrill flow increases because the water dissipates some of the energy of the rainfall. The relationship may be expressed by the equation (Morgan *et al.* 1998)

$$D = \frac{k}{\rho_s} KEe^{-bd}$$

where $D$ is the detachment rate, $k$ is an index that varies with soil type, $\rho_s$ is particle bulk density, $KE$ is rainfall kinetic energy and $b$ is a constant that varies with soil texture.

In rills, detachment is achieved by the shear stress exerted by the flow. Although there have been several attempts to quantify the threshold conditions for rill initiation by overland flow (e.g. Slattery and Bryan 1992), Nearing (1994) has pointed out that the shear stress exerted by shallow flow is of the order of a few pascals, whereas the shear strength of soils is of the order of a few kilopascals.

## Saturation-excess overland flow

The acceptance of Horton's model for the generation of overland flow is at odds with the fact that it is very seldom observed in many environments, particularly those in which there is appreciable vegetation cover and/or rainfall is cyclonic, rather than convective. Kirkby and Chorley (1967) argued that rain falling onto an already saturated soil will also remain at the surface and become overland flow. Generation of this type of overland flow depends not so much on the relationship between rainfall intensity and soil infiltrability as on the amount of water that is already in the soil at the onset of rainfall, known as the antecedent



moisture content, and the water-storage capacity of the soil. These amounts are both spatially and temporally variable. Antecedent moisture is likely to be highest on footslopes and in concavities, and areas with thin soils (such as spurs) have low total water storage capacity. Both of these areas will preferentially generate saturation-excess overland flow. Antecedent moisture will also depend on rainfall record prior to an individual storm event. Taken together, these two factors mean that, in contrast to Hortonian overland flow, saturation-excess overland flow is likely to be generated on only some parts of hillslopes (the concept of partial area contribution to overland flow – see Betson and Marius 1969), and be variable for two storms of similar characteristics (the concept of variable source area – see Dunne and Black 1970). Because saturation-excess overland flow is generated locally, particularly in areas close to rivers, it is an important control on catchment hydrographs. Conversely, because much is generated on low-angle footslopes, it is of much less importance for soil erosion on hillslopes.

## Return flow

Water that infiltrates into the soil and moves downslope through the soil as throughflow or in pipes (see PIPE AND PIPING) may encounter saturated soil, thereby having its further downslope movement through the soil blocked. This water may be forced to the surface, where it is known as return flow, and travel further downslope as overland flow. Like saturation-excess overland flow, return flow is generated locally, often on footslopes, so its significance lies in its impact of catchment hydrographs.

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SEE ALSO: soil erosion

A.J. PARSONS

# OVERWASHING

Overwashing is generally regarded as the process of sediment transport across a BEACH RIDGE or barrier beach (see BARRIER AND BARRIER ISLAND), with deposition as a washover deposit on the back slope of the ridge or in the lagoon landward of the ridge occurring during storms. Morton *et al.* (2000) have, however, reported frequent overwash events occurring during non-storm periods. The term 'overwash' is generally applied to the process, and 'washover' to the resulting despositional landform. Overwashing is a major process by which the back slope of a barrier beach is renewed with the barrier building to landward. Overwash-dominated barriers tend to be relatively narrow and flat, with low and unstable dune systems. Where sediment supply is abundant, overwash barriers may grow and stabilize, but where sediment supply is limited, overwash causes barriers to roll over with sediment being moved from the seaward slope to the landward slope, thereby causing the barrier to migrate landward. Overwashing is an important process on both sand-dominated barriers (where AEOLIAN PROCESSES may also be important) and gravel barrier systems.

Overwash can occur along a significant length of barrier crest producing a washover ramp on the landward side, but more commonly flow is concentrated in channels called overwash throats. In some circumstances and in the long term an overwash throat may lead to the development of a tidal inlet. Washover fans develop on the landward side of the barrier where flow is no longer constrained. Orford and Carter (1984) relate the spacing of overwash throats and washover fans to beach rhythmic morphology (see BEACH; BEACH CUSP; WAVE).

Overwashing (which generally leads to the lowering of a beach barrier) must be differentiated from overtopping which is the process by which barrier systems are built up due to swash transport of sediments to the top of a berm or barrier crest, thereby causing the barrier to build in height and width. Overtopping and overwashing may occur at the same locations, dependent upon wave energy conditions.

Although often considered as a somewhat different process, overwash sedimentation also occurs on CORAL REEF islands, and can be important for the maintenance and development of reef beaches and cays.

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SEE ALSO: beach ridge; storm surge

KEVIN PARNELL

# OXBOW

An abandoned meander (see MEANDERING) loop along an alluvial river. It is the most common type of lake of fluvial origin. As a logical consequence of the lateral shifting and the general downstream migration of meanders, it develops from the interplay of channel erosion and accumulation separated in space. Oxbows are produced by meander cutoffs, which occur in two different ways. If a meander neck becomes narrow enough, streamflow is directed along the shortest route of greatest slope, instead of following the whole perimeter of the meander loop (neck cutoff). Alternatively, a new channel may develop along a swale between POINT BARS (chute cutoff). Natural LEVEE formation and FLOODPLAIN deposition soon build a silt or clay plug between the oxbow and the main channel, although a narrow batture (watercourse) may provide a connection. The American name emphasizes the crescent shape, while in many other languages an oxbow is called a 'dead arm' (e.g. in French: *bras mort*). Both types of cutoff cause channel shortening and scour upstream and deposition downstream. Thus the meandering river maintains its average SINUOSITY since each cutoff triggers the formation of further cutoffs on the long term (self-organization – Stolum 1996). The tranquil freshwater makes an oxbow a valuable aquatic habitat. Meander scars are oxbows completely filled up with mineral and organic matter. They remain discernible in the landscape for a long time.

Classic examples are found along several major rivers of the world, including the Mississippi, the Amazon and the large rivers of Siberia. Floodplains of some minor (regulated) rivers also abound in oxbow lakes.

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DÉNES LÓCZY

# OXIDATION

Oxidation is the loss of a negative electron so an element becomes more positively charged, for example ferrous iron, $Fe^{2+}$ (or Iron II) becomes oxidized to ferric iron $Fe^{3+}$ or Iron III. This process commonly occurs in the presence of oxygen:

$$\underset{\text{Iron II}}{4FeO} + O_2 \rightarrow \underset{\text{Iron III}}{2Fe_2O_3}$$

Oxidation is thus a common weathering reaction when a mineral formed in an anoxic environment becomes exposed to air at the surface of the Earth. The process is often combined with HYDROLYSIS, for example the weathering of olivine:

$$\underset{\text{Olivine}}{3MgFeSiO_4} + H_2O \rightarrow \underset{\text{Serpentine}}{H_4Mg_3Si_2O_9} + SiO_2 + \underset{\text{Ferrous oxide}}{3FeO}$$

the FeO then becoming oxidized to iron oxide, as above.

During oxidation, the strength of the minerals is reduced which also makes mechanical breakdown much easier.

The loss of a negative electron can equally occur when iron in an acid solution becomes less acid. The latter process accounts for the deposition of reddish iron oxides in the less acid lower parts of soil profiles where there can be, in fact, less oxygen than nearer the surface.

Iron oxides are an important constituent of tropical soils, being produced as a residual mineral through prolonged, intense weathering. The simple iron oxide, haematite ($Fe_2O_3$) is bright red and leads to the distinctive colour of soils in tropical and subtropical areas. If haematite is subject to HYDRATION then limonite ($2Fe_2O_3{\cdot}3H_2O$) forms which is yellow in colour. The content of iron also appears to be a key factor in the formation and hardening of laterite (Thomas 1974). Many of the theories of laterite formation involve the movement of iron in solution by mobile groundwater, or by upward diffusion, and their subsequent oxidation and mobilization near the surface; the harder laterites having a higher iron content.

Oxidation is not only a weathering process in itself, it can produce further weathering agents. The oxidation of iron sulphides (pyrites) can produce sulphuric acid:

$$\underset{\text{Pyrite}}{2FeS_2} + 15(O) + H_2O \rightarrow \underbrace{2Fe^{3+} + 4SO_4^{2-} + 2H^+}_{\text{Sulphuric acid}}$$

There is a notable example of this at Mam Tor in Derbyshire where pyrite oxidation and the production of sulphuric acid leads to the intense weathering of the illite and kaolinite present in the shale in which the pyrite occurs, leading to a marked rise in porosity and facilitating slope instability, contributing to the collapse of a road (Vear and Curtis 1981). The authors calculate that for 1.5 g of pyrite oxidized by a litre of acid-sulphate water, $0.0125\,gl^{-1}$ $H^+$ is produced.

While geomorphologists often focus on the oxidation of iron, other compounds can also be oxidized, for example manganese, and a key process in the nitrogen cycle involves the oxidation of ammonium produced by the decomposition of organic matter to nitrate:

$$\underset{\text{Ammonium}}{4NH_4^+} + 6O_2 \rightarrow \underset{\text{Nitrite}}{4NO_2^-} + 8H^+ + 4H_2O$$

and

$$\underset{\text{Nitrite}}{4NO_2^-} + 2O_2 \rightarrow \underset{\text{Nitrate}}{4NO_3^-}$$

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STEVE TRUDGILL

# OYSTER REEF

'Among organic reefs, those of the geologically young oysters are now second in size and distribution only to the coralline reefs' (Price 1968: 799). The reefs are built by the modern estuarine *Ostrea* and *Crassostrea* and the marine *Pycnodonte*. The tops of the reefs range from the intertidal zone to depths of as much as 12 m below sea level. Optimum temperatures are from 15–25 °C and optimum salinities occur in the central parts of bays and other estuaries midway between stream mouth and oceanic opening. In addition to forming reefs and visors, they can help to stabilize spits and other constructional features. Oyster reefs are widespread, and major locations include the Gulf of Mexico (*Crassostrea virginica*) and parts of China (*Crassostrea gigas*) (Wang Hong *et al.* 1995), but in some parts of the world they are being destroyed by human activities (including gastronomy and fishing); this has led to attempts to restore them or to provide artificial substances for their colonization (Coen and Luckenbach 2000).

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A.S. GOUDIE

# P

## PALAEOCHANNEL

When a channel ceases to be part of an active river system it becomes a palaeochannel. Palaeochannels vary greatly in age. Those from the recent geological past (perhaps tens to hundreds of years old) include meander cut-offs and longer reaches abandoned by AVULSION. Although these channels are now isolated from the active river flow, except during periods of floodplain inundation, they are scaled to present flow regime. Many of the more ancient palaeochannels indicate discharges greatly in excess of those occurring at present. The oldest palaeochannels known are found on Mars where huge flows of surface water carved a complex pattern of channels more than 3.5 billion years ago. Clearly, the Martian channels could not have formed in the planet's present waterless environment.

Where palaeochannels are well preserved they provide valuable information about past flow regimes. The basis of discharge reconstruction is given by established statistical relationships between channel forming (bankfull) discharge and aspects of channel morphology including cross-sectional area and meander wavelength as documented by the US Geological Survey in the 1950s and 1960s and amply confirmed since. Meandering rivers are particularly suited to this work because of their potential to preserve planform and, sometimes, cross-sectional geometry. The relationship between meander wavelength and stream discharge (Dury 1965) provides a useful, although often imprecise, approximation of palaeoflow. Rotnicki (1983) argued, on the basis of fieldwork on the Prosna River in Poland, that channel cross sections in meander neck cut-offs provided more reliable estimates because of the excellent preservation of channel dimensions at such locations.

The high degree of channel preservation in neck cut-offs results from their mode of formation. Upon initiation of a short circuit, sedimentation rapidly seals the cut-off ends to form an OXBOW lake. In the low energy environment of the lake, fine-grained sediments from suspension form a drape over the old riverbed. The infill sediment cast effectively preserves the former channel cross section, which can be revealed subsequently in a series of auger holes.

In Rotnicki's (1983) comparative study, estimates of BANKFULL DISCHARGE based on fourteen equations linking meander wavelength and discharge were compared with estimates based on preserved cross-sectional dimensions at cut-offs. For a measured bankfull discharge of $22.5\,m^3\,s^{-1}$ the estimates based on wavelength ranged from 0.2 to $34.1\,m^3\,s^{-1}$, or more than two orders of magnitude. The frequently used equations of Dury and Carlston gave errors of 36 per cent and 78 per cent respectively. Estimates of bankfull discharge based on cut-off cross sections and the Manning Equation reduced the error to 10 per cent.

Many palaeochannels also indicate past regimes of markedly different channel pattern and sediment load. The transition from late glacial conditions to those of the Holocene produced a strong channel response globally. In regions directly affected by ice, many large braided (see BRAIDED RIVER) and bedload-dominated proglacial channels gave way to meandering mixed and suspended load channels. In North America the Mississippi provides an excellent example. At the same time, in regions far removed from the great Quaternary ice sheets parallel changes occurred. For example, on the Riverine Plain in southeastern Australia, low sinuosity, aggraded sand-bed channels (PRIOR

STREAMS) were converted to highly sinuous systems dominated by fine-grained sediments. Here, changing upper catchment conditions including rising temperatures and treelines, and shrinkage of the winter snowpack, produced the channel response. Selected examples of palaeochannels reported in the scientific literature are presented below.

## Underfit streams

From the late nineteenth century it was recognized that some sinuous valleys in Europe contain floodplains on which present-day rivers describe smaller wavelengths than the enclosing valley (Davis 1899). Such streams were described as being manifestly (obviously) underfit. The inference being that a reduction in discharge had resulted in a reduction in meander wavelength. George Dury's (1964a,b, 1965) detailed studies in Europe and North America demonstrated the fluvial origin of large meandering valleys and the widespread regional distribution of UNDERFIT STREAMS. Following the elimination of other possible causes, including headwater capture and the loss of glacial meltwater, Dury deduced that regional climatic change had been responsible for the observed reduction in discharge. Radiocarbon dating of valley fills indicated that the last major discharge shrinkage occurred between 10,000 and 12,000 years ago, at the beginning of the Holocene.

Dury estimated the discharges of the large palaeochannels largely on the basis of the statistical relationship between meander wavelength and discharge. Although the computed discharges exceeded those of the present rivers by up to a factor of 60, Dury argued that they could have been produced by glacial climates characterized by reduced evapo-transpiration and a 50–100 per cent increase in precipitation. Many of the reconstructed discharges approach the largest discharges ever recorded on Earth for catchments of equal drainage area and thus were considered by many workers to be excessive, especially in areas of low relief unsuited to the production of extreme flows. In particular, there was concern about the influence of parameters other than discharge on meander wavelength. Various studies have subsequently shown that meander wavelength alone does not provide reliable estimates of bankfull discharge.

## Superflood palaeochannels

In the 1920s, J. Harlen Bretz (1923) first described superflood palaeochannels in the Columbia Plateau region of the northwestern United States. Here, space imagery reveals a complex network of large anastomosing (see ANABRANCHING AND ANASTOMOSING RIVER) channels carved into basalt bedrock and overlying LOESS and other sediments. Fluvial features include great waterfalls, potholes, longitudinal grooves carved into the bedrock, and boulders that were transported and deposited in flood bars and giant current ripples. Bretz attributed these features to a cataclysmic flow he called the Spokane Flood.

The eventual verification of Bretz's catastrophic flood hypothesis, despite vehement opposition at the time of its proposal, is considered by Baker (1978) to be one of the most fascinating episodes of modern science. On the basis of flow reconstructions by Shaw *et al.* (1999) it now appears that the giant scabland floods arose from a combination of sources including late Pleistocene ice-dammed Lake Missoula and large subglacial reservoirs that extended over much of British Columbia. An estimated total volume of $10^5\,km^3$ flowed across the Scablands and achieved a peak discharge of some $17 \times 10^6\,m^3\,s^{-1}$. The power per unit area of streambed generated by these flows was up to 30,000 times greater than that produced in the present Amazon.

Similar outburst floods (jökulhlaups) have been documented for spillways marginal to the Laurentide Ice Sheet and in Swedish Lapland. Cataclysmic flows generated by Pleistocene ice-dammed lake failures in the Chuja Valley in the Altas Mountains of south-central Siberia (Baker *et al.* 1993), which exceeded $18 \times 10^6\,m^3\,s^{-1}$, are comparable to the largest of the Channeled Scabland flows. Spectacular palaeochannel landforms in the Chuja Valley include scoured channels, giant bars and gravel wave trains. The impressive hydraulic parameters associated with the Chuja Valley floods include flow depths of 400–500 m, supercritical flow velocities of $45\,ms^{-1}$ and stream powers approaching $10^6\,W\,m^{-2}$. The outburst floods of the Channeled Scabland and Chuja Valley are Earth's greatest known terrestrial discharges of freshwater.

Landform assemblages characteristic of cataclysmic flooding are also present on Mars (Baker *et al.* 1993). The Martian outflow channels, which were first recognized on the basis of Mariner 9 and Viking space mission imagery, are much larger than those of the Channeled Scabland and may have experienced discharges as great as $10^9\,m^3\,s^{-1}$. The Martian palaeochannel systems are therefore not only the largest known, but also the oldest, dating from before 3.5 billion years ago.



## Palaeochannels in Australia

Australia's modest stream discharges, compared to those of other continents, result from its predominantly subtropical location and low average relief. Not surprisingly, the presence of large palaeochannel systems in Australia's two inland drainage basins with areas exceeding $1{,}000{,}000\,km^2$, the Murray–Darling and Lake Eyre (Figure 112), has been of particular interest to geomorphologists seeking to reconstruct Late Quaternary hydrological regimes.

In the Murray–Darling Basin, palaeochannels of the Murrumbidgee River have been studied extensively since the late 1940s. Previously described as prior streams, large palaeochannels here form an impressive distributary system in a region now characterized by small meandering rivers. Research before 1970 subdivided the palaeochannels into two genetically different categories: older prior streams and younger ancestral rivers. Channels described as prior streams were aggraded bedload systems characterized by low sinuosity, high width to depth levees and source bordering sand dunes. The sinuous ancestral channels were characterized by floodplains of lateral migration and discharges much larger than the present rivers in this region (Plate 84).

Thermoluminescence (TL) dating by Page *et al.* (1996) resulted in a major revision of the prior/ancestral model. It was shown that four major surface palaeochannel systems (Coleambally, Kerarbury, Gum Creek and Yanco) operated between 100,000 and 12,000 years ago with frequent alternations between prior and ancestral modes of channel behaviour. Stratigraphic investigations showed that bedload aggraded channels (prior streams) typically were bordered by fining-upwards deposits associated with laterally migrating channels (ancestral rivers). Clearly, there was a need to revise the existing model in which prior steams preceded ancestral rivers. Page and Nanson (1996) proposed that the first three phases of Murrumbidgee palaeochannel activity were characterized by alternations from laterally migrating to vertically aggrading channel behaviour with each phase terminating in vertical aggradation and the formation of source-bordering sand dunes (Figure 113). Only the final (Yanco) phase failed to terminate in bedload aggradation, probably because the onset of Holocene climates reduced the size of flood peaks, greatly diminished the supply of bedload from the upper catchments, and resulted in streams evolving into their present highly sinuous suspended load morphology.

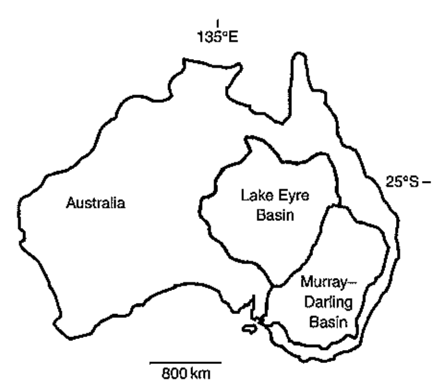


Figure 112 Map of Australia showing locations of Lake Eyre and Murray–Darling Basins


Plate 84 (a) Ancestral Green Gully palaeochannel, and (b) present channel of Murray River in southern Australia at same scale and drainage area

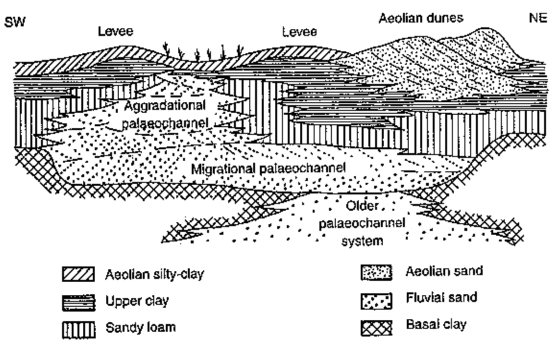


Figure 113 Stratigraphic model of Murrumbidgee River palaeochannels (Page and Nanson 1996: 943)

Discharge reconstructions at preserved channel cross sections of Murrumbidgee palaeochannels suggest that bankfull flows were between four and eight times greater than those of the present rivers.

The Channel Country of the Lake Eyre Basin (Figure 112), which includes Cooper Creek and the Diamantina River, comprises a vast system of low-gradient anastomosing channels dominated by fine-grained suspended load. The anastomosing channels, which date from about 80,000 years ago, are mud-lined, laterally very stable and underlain by extensive muddy floodplains. However, along the middle and lower reaches of Cooper Creek aerial photographs and subsurface exploration have revealed remnant scroll-bars and palaeochannels beneath the mud unit. The scrolls, which are scaled to river meanders larger than any present in the system today, were formed by mixed-load, laterally migrating rivers that deposited extensive sandy units with abundant flow structures (Katipiri Formation). TL dating by Nanson *et al.* (1988) showed that the Katipiri sands date from at least 250,000 years ago.

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KEN PAGE

## PALAEOCLIMATE

The climate of the past: palaeo, from the Latin word meaning ancient or old; climate, refers to the interconnected group of Earth systems that control weather conditions (temperature, moisture, wind, etc. and the spatial/temporal variation in these factors) at the surface over an extended period of time. The study of past climates is referred to as palaeoclimatology. More specific definitions of palaeoclimate exist (e.g. climate prior to instrumental records) but in the present case a broad definition of palaeoclimate is taken as that meaning all climates prior to the present day. Further divisions of specific geological time periods may then be defined (e.g. Holocene, Quaternary, Permian) and some of these are discussed below briefly.

Climate has a significant influence on most geomorphic processes and an understanding of landscape systems cannot be achieved without knowledge of both present day climate and climatic history. Antecedence in geomorphic systems is often controlled to a large degree by palaeoclimate and interpretation of geomorphic records (e.g. through sedimentological records) is dependent upon an appreciation of climatic variations.

Climate varies on all temporal and spatial scales and the notion of scale is particularly important both in geomorphology and in climate studies. The nature of climate change at each spatial/temporal scale is determined to some extent by the factors forcing the change. Over long 'geological' timescales, the movement of tectonic plates, and associated volcanism, effect far-reaching but gradual changes in global climate. Over periods of hundreds of thousands of years, the influence on insolation of orbital variations (as described by the 'Milankovitch theory') may have driven the large-scale, high amplitude, changes in global climate of the Quaternary period (see below). Both tectonic and orbital forcing are examples of factors 'external' to the climate system. At shorter timescales (e.g. at the millennial scale), variations are probably due largely to 'internal' factors, such as the differences in response time of components within the climate system, and the subsequent chaotic dynamics of these strongly coupled (non-linear) systems. Internal controls often give rise to abrupt changes in climate as thresholds are reached and feedback systems evolve. Over these shorter timescales, the climate system is highly complex, meaning that cause and effect are often difficult to separate.

Information on past climate conditions can be obtained, in some cases, from historical records (e.g. farming records, instrumental data). However, such records are often sporadic, of questionable accuracy and limited in extent, both spatially and temporally. A second method is to employ the use of climate-dependent natural phenomena that leave traces in the geological record. Once calibrated, these traces can then be used as proxies for past climates. Calibration involves defining the dependence of each proxy on climate, sometimes using modern analogues, sometimes using theoretical calculations (the principal of uniformitarianism then applied). Broadly, there are two kinds of proxy data: (1) episodic/discontinuous records (e.g. flood deposits, glacial advances), which result from the integration of climatic conditions prior to the event; and (2) continuous/incremental records (e.g. constant accumulation of marine mud) which preserve a quasi-continuous record of environmental/climatic conditions. Some examples of proxy records are:

1 *Glaciological*: composition and macrostructure of ice cored from both large ice sheets (e.g. Antarctica, Greenland) and from mountain glaciers – provides information on both regional air temperatures and on global/regional atmospheric composition;
2 *Geological*: ocean sediment cores (geochemistry – particularly oxygen isotope profiles – and species composition of micro fauna), glacial features, sedimentary deposits (e.g. loess, sand dunes), chemistry of speleothems;
3 *Biological*: pollen recovered from terrestrial/marine sediments, remains of insects and micro fauna, composition and structure of tree rings.

The degree to which these proxies are affected by global/regional/local influences depends upon many factors (some of which are context specific) and, to some extent, what we see depends upon how we choose to look at the climate record. Broadly, globally integrated climate signals can be obtained from marine sediment geochemistry, loess/palaeosol deposits and ice sheet cores. However, all proxies are affected by both random fluctuations (noise), non-climate related processes and lags in response times, and all have different sensitivities to climatic conditions. These factors underline the importance of multi-proxy datasets in palaeoclimate research.

As with other geographical matters, it is not easy to discuss palaeoclimates as sets of primary data or even calibrated indices (e.g. temperatures) and further interpretations are required (sometimes quantitative, sometimes qualitative, e.g. 'wetter' or 'warmer'). For conceptual ease, palaeoclimates are commonly discussed in terms of simplifications such as by warmer/colder, drier/wetter or in other even more generalized terms such as reference to glacial/interglacial conditions. Care is needed when using such terms as they often have different meanings in different parts of the world. However, on the regional scale, these terms are indeed useful and are often retained in the palaeoclimate literature.

Climate has varied widely over the history of the Earth. Geological evidence from Precambrian times (approx. 860–550 Ma) points to large-scale glaciations, although the extent of ice cover during this period is debated. Some evidence suggests the Earth was almost completely ice covered, with glaciers reaching to the tropics, while other evidence puts the effected land masses at higher latitudes during these times. Since the Cambrian, the Earth's mean temperature has varied considerably, between the icy conditions of the Permian ice age to the more tropical temperatures of the late Cretaceous. Over the last ~55 Ma, there is strong evidence for a cooling trend at both poles and across the lower latitudes, culminating in the glacial conditions of the last ~2 Ma, the Quaternary period. While the deep geological past sets the wider stage on which contemporary processes operate, it is perhaps the Quaternary period, with its high amplitude, high frequency climatic changes, that is of most significance in understanding the geomorphic setting of most present-day landscapes.

## Quaternary palaeoclimate

During the Quaternary period (~2 Ma) the Earth's climate has been through many oscillations in climate, that have operated over a range of temporal and spatial scales. Perhaps the most characteristic feature of the Quaternary has been the oscillation between glacial and interglacial conditions. During glacial conditions, large continental ice sheets grew on Northern Europe/Eurasia and Canada/North America, causing SEA LEVELS to fall by up to 120 m compared to the present day. During the warmer interglacials, ice sheets were restricted to the poles and to Greenland and temperatures were similar to those experienced at present. Insolation forcing due to orbital variation (the Milankovitch theory) at periods of roughly 100, 40 and 20 ka are believed to play a key role in these climate oscillations, although other factors cannot be ruled out. At lower latitudes, the effects of insolation forcing were different, affecting, for example, the intensity and distribution of precipitation (particularly the monsoon systems).

Other factors such as heat distribution due to the ocean current circulation and atmospheric gas composition are likely also to have played key roles. Some lines of evidence (e.g. ice and marine core data) suggest rapid high magnitude changes also occurred during the Quaternary, with mean regional temperatures changing by as much as 10 °C over decades or centuries. Such changes would have significant effects on geomorphic processes and these effects are often recorded in the sedimentological and morphological record.

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SEE ALSO: El Niño effects; Holocene geomorphology; ice ages

RICHARD BAILEY

# PALAEOFLOOD

Palaeoflood literally means 'ancient flood', but the word does not necessarily connote a specific age, and is often used for any flood not systematically gauged. The characteristics of ungauged floods may be inferred using historical, botanical or geological evidence (Wohl and Enzel 1995). Historical evidence comes from qualitative flood records kept by humans. High-water marks on buildings or canyon walls, diary entries, newspaper reports or damage reports for insurance purposes may all be used to estimate the magnitude and date of occurrence of floods. Such records may extend back 2,000 years in countries such as China.

Botanical evidence of past floods comes from vegetation growing along the riparian corridor (Hupp 1988). Flood-borne debris may impact riverside trees hard enough to destroy a portion of the tree's cambium and leave a corrasion scar that can be dated using annual growth rings in some tree species. Maximum scar heights may be used to estimate minimum peak stage. Flood debris may also bend or break trees near the channel. Many tree species can survive such damage and develop adventitious sprouts, usually within a year of the damage, which can also be dated using tree rings. Anomalies in the width or symmetry of annual growth rings result from changes in water availability, tilting of the tree or stripping of the tree's leaves, all of which may be associated with floods. The age structure of riparian vegetation indicates minimum time since initial deposition or scour of alluvial surfaces. Each of these botanical indicators may provide chronologically precise information on a range of flows for species with annual rings, but the use of these indicators is limited by the presence and age of the vegetation.

Geological evidence of ungauged floods may come from dimensions of relict channels, the size of fluvially transported sediment, or erosional and depositional features that indicate maximum flood stage. Palaeochannels may be preserved as exposed cross sections, abandoned channels on the surface or exhumed channels (Williams 1988). Form parameters including drainage density, terraces, channel pattern, meander wavelength and channel cross-sectional dimensions have been used to infer flow parameters including mean velocity and discharge by means of empirical equations developed for active channels. Many of these form parameters are best preserved along low gradient, unconfined alluvial rivers where continued lateral movement of the channel has left abandoned channels relatively well preserved. Along these rivers, form parameters commonly record low magnitude floods such as average discharge or mean annual discharge. Estimates of flow parameters using channel characteristics may be inaccurate because of deficient regression equations based on limited data; misapplication of available equations caused by extrapolation to different conditions of channel pattern and climate; inadequate preservation of abandoned channels; improper algebraic manipulation of the empirical equations; and uncertainty in the definition of some variables (Williams 1988).

Sediment characteristics may be related to flow parameters by first relating particle size to some index of local transport capability, such as STREAM POWER, and then transforming the transport variable into a discharge estimate using a hydraulic flow equation such as the Manning equation. Gravel and finer sediments may be used in the aggregate to reflect average flow. This approach is commonly used for lower gradient alluvial channels. Coarser sediments are often treated as individual particles, with a focus on the flow competence necessary to transport the largest particles present (O'Connor 1993). This approach is more commonly used for higher gradient confined channels such as bedrock canyons. Use of both finer and coarser sediments to estimate flow parameters relies on empirical relations developed between transported particles and observed, calculated or inferred flow conditions.

Erosional and depositional features may provide palaeostage indicators that record the maximum stage of individual flows. Erosional features include lines scoured into valley-wall soil and colluvium; truncation of landforms such as debris flow fans impinging on the channel; or vegetation limits below which individual species of vegetation are absent (Jarrett and Malde 1987). Depositional features include silt lines of very fine sediment and organics adhering to the channel banks; accumulations of organic debris from fine particles to logs; and slackwater deposits of sediment settling from suspension in areas of flow separation such as tributary mouths or channel-margin alcoves or caves (Kochel and Baker 1982). Palaeostage indicators are best preserved along confined channels with resistant boundaries where an increase in discharge produces a large increase in stage, and changes in channel geometry during and between floods are minimized; and in drier climates where the indicators are less likely to be weathered or obliterated by non-fluvial processes. Flood chronologies may be established from palaeostage indicators using both absolute geochronologic methods such as radiocarbon or thermoluminescence, and relative methods such as stratigraphic position or soil development. Combined with surveyed channel geometry, the stage indicators can be used to estimate flood magnitude (Webb and Jarrett 2002). Palaeostage indicators are commonly used to estimate the largest floods along a channel.

The majority of palaeoflood studies address floods that occurred during the late Pleistocene and Holocene. The late Pleistocene was characterized by immense outburst floods such as those in the Channeled SCABLAND and Siberia produced by the release of meltwater ponded along the margins of the continental ice sheets.

Geological methods used to estimate palaeoflood magnitude on Earth have also been applied to channels on Mars (Baker 1982).

Palaeoflood studies are distinguished from other types of fluvial palaeohydrology in that they usually focus on maximum flows along a channel rather than the entire range of flows. Palaeoflood studies may be a part of studies focusing on channel change as recorded in terraces (see TERRACE, RIVER), ARROYO formation or COMPLEX RESPONSE, or palaeoflood data may be used to examine issues of flood-frequency analysis, flood hydroclimatology and the geomorphic effectiveness of floods.

Flood-frequency analysis is largely based on the measured or extrapolated recurrence interval between discharges of a given magnitude. Measured recurrence intervals are limited by the time span of systematic discharge measurements, which is rarely longer than a hundred years. Extrapolated recurrence intervals may come from extending an existing flood-frequency curve beyond the time span of measurement, or from combining records from neighbouring regions and using the cumulative record length. Both approaches assume that the statistical properties of the hydrologic time series do not change with time, a condition known as stationarity. However, changes through time in the type or frequency of flood-producing storms, or changes in rainfall-runoff generation resulting from land use, are widespread (Hirschboeck 1988). Extending the systematic flood record with palaeoflood information avoids the problem of nonstationarity in the past because palaeoflood indicators record actual rather than hypothetical past floods (Baker *et al.* 2002). Palaeoflood records can also help to constrain the estimate of the probable maximum flood, the largest probable flood that could theoretically occur in a drainage basin. Statistical incorporation of palaeoflood data into systematic data relies on recognition of differences in the two types of data. For example, systematic data may include all floods above a fixed magnitude threshold, whereas the magnitude threshold for palaeoflood data may have varied through time (Blainey *et al.* 2002).

Palaeoflood indicators that record changes in flood frequency through time can also indicate changes in climatic circulation patterns (Redmond *et al.* 2002). And records of the magnitude and frequency of large floods may be used to infer rates of geomorphic change for channels dominated by floods (Wohl 2002) (see FLOOD).

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SEE ALSO: flood

ELLEN E. WOHL

## PALAEOHYDROLOGY

Palaeohydrology is the study of past occurrences, distributions and movements of continental waters. It is the highly interdisciplinary linkage of scientific hydrology with the sciences of Earth history and past environments (Schumm 1967). The linkage extends in both directions in that modern hydrological data can be used to create the means of reconstructing past environments (Schumm 1965), while data from past hydrological processes can be used to calibrate and test modern hydrological models (Baker 1998).

The term *palaeohydrology* was first used by Leopold and Miller (1954) in their study of past hydrological conditions associated with a sequence of late Quaternary alluvial terraces in Wyoming. Nevertheless, it is applicable to all elements of the hydrological cycle. Thus, many aspects of cave development in karst aquifers preserve indicators of paleohydrology for those aquifers. Similarly, past changes in lake levels can be documented in terms of a hydrological balance. All these branches of palaeohydrology derive from long traditions in geology and related Earth sciences. For example, Patton (1987) documents the interest by nineteenth and early twentieth-century geologists in past changes in river processes, as evidenced in deposits, terraces and other landforms. Particularly important was the example of Bretz (1923), who discovered the catastrophic flood origin of the Channeled Scabland region in the northwestern United States. Subsequent palaeohydrological quantification (Baker 1973) showed that immense catastrophic flood discharges generated the scabland features during the late Pleistocene bursting of ice-dammed glacial Lake Missoula.

### Modes of palaeohydrological inference

There are three general modes of reasoning in palaeohydrology. In one mode general theories of hydrology are used to infer specific effects that can then be discerned in evidence of past hydrological processes. This is the classical deductive mode of rational inquiry. An example would be the problem of the catastrophic flooding associated with the failure of ice-dammed glacial lakes. The palaeohydrologist can use an existing theoretical model for how such a dam fails. Of course, the effective use of this model requires that the correct mode of dam failure be matched with the model (Figure 114). With this condition satisfied, the model may be capable of predicting the hydrograph of the resulting flood. Matching the predicted hydrograph properties to preserved field evidence then constitutes a kind of reconstruction of the past hydrological process.

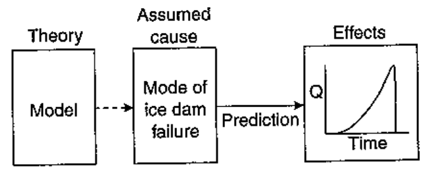


Figure 114 Schematic representation of the deductive mode of palaeohydrological inference applied to the problem of predicting an outburst flood hydrograph from a general theory for the failure of an ice-dammed lake

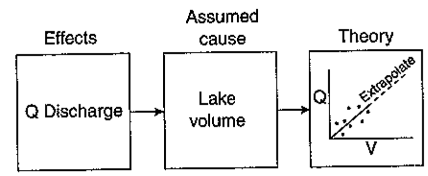


Figure 115 Schematic representation of the inductive mode of palaeohydrological inference applied to the problem of estimating the relationship between lake volume and peak outflow discharge for the failure of ice-dammed glacial lakes

Another common mode of palaeohydrological inference uses empirical relationships that are developed from numerous observations of related hydrological phenomena. This mode of scientific reasoning is inductive. Returning to the problem of ice dam failure, one can collect data on modern glacial lakes. By relating the peak outburst discharges to the associated lake volumes, one can derive an empirical relationship between these two variables (Figure 115). This relationship, extrapolated to the evidence for past lake volumes (or peak discharges) can then be used to estimate the associated discharge (or lake volume). Of course, this exercise must presume that the past phenomena fall in the same class as the data set on modern outburst floods. This is a limitation on all inductive reasoning, because nature is not constrained to behave as we presume it should from our limited set of observation.

Finally, a third mode of reasoning that is used extensively in palaeohydrology is retroductive, or abductive inference (Baker 1996, 1998). For the flood problem, retroductive inference can be accomplished by studying evidence or signs of the

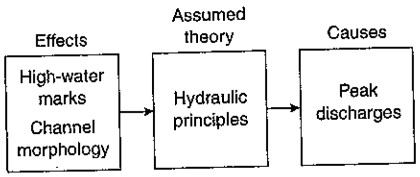


**Figure 116** Schematic representation of the retroductive or abductive mode of palaeohydrological inference applied to the problem of estimating peak palaeoflood discharges from various evidence of palaeoflood stages, utilizing hydraulic theory

past floods. These might include the slackwater deposits emplaced marginal to flood channels, or other high-water marks for the past flow stages, as done in *palaeoflood hydrology* (Baker 1987). Then, using a hydraulic model, it is possible to associate the past flood effects with the causative discharges (Figure 116). Thus, retroductive reasoning proceeds from effect to cause, in contrast to the deductive reasoning that proceeds from cause to effect.

## Fluvial palaeohydrology

The most basic relationships between river morphology and hydrology involve the supply of water and sediment from upstream of a channel reach of interest. The important dependencies are summarized in the following relationships:

$$Q_W \propto \frac{w, d, \lambda}{S}$$

$$Q_S \propto \frac{w, d, S}{d, P}$$

where $Q_W$ is a measure of the mean annual water discharge and $Q_S$ is a measure of the type of sediment given by the proportion of bedload (usually sand and gravel) to the total sediment load (which may include considerable clay and silt). $Q_W$ and $Q_S$ are the controlling, independent variables. The dependent variables include the channel width $w$ and depth $d$, the slope or gradient of the channel S, the sinuosity of the reach P, and the meander wavelength $\lambda$. For palaeohydrological applications the above relationships are usually quantified by empirical equations, using the inductive approach.

An example of the foregoing reasoning is in regard to the phenomenon of *underfit streams.* These are streams for which some practical measure of the modern river, usually the meander wavelength $\lambda$, is too small in relation to the valley that contains the stream. Long recognized as being caused by stream capture, underfitness was also recognized in the context of climatic change by Dury (1954, 1965). Dury (1965) reasoned that, because meander wavelength is directly proportional to bankfull channel width, and because bankfull width is a function of discharge to the 0.5 power (Leopold and Maddock 1953), the wavelength of modern river meanders $\lambda$ must be proportional to modern bankfull discharges $q_b$. Applying the same arguments to the enlarged valley meanders of wavelength $L$ formed by ancient discharges Q, Dury (1965) finds

$$Q/q_b = L^2/l^2$$

Dury's study of many rivers in the United States and Europe showed that the ratio $L/l$ varies from 5 to 10, which implies that the ancient discharges Q were 25 to 100 times larger. The immense climatic implications of such large changes led many to question Dury's estimates. In subsequent work it was discovered that the discharges responsible for valley meanders, which are often developed in bedrock, may have very different relationships to channel size than the empirical relationships that apply to modern alluvial rivers.

A significant discovery in fluvial palaeohydrology came when attempts were made to apply Dury's theory to underfit streams on the Riverine Plain of southeastern Australia. The modern Murrumbidgee River is underfit relative to the very large meanders of an ancestral channel. It has a much narrower channel and a much smaller meander wavelength. In addition, there are *prior channels*, which constitute an older system of paleochannels filled with sediment that is much coarser than that conveyed by either the modern Murrumbidgee or by its ancestral stream. Because the prior channels are much wider and have much greater meander wavelengths than the modern river, Dury's theory predicts that they should have experienced much larger bankfull discharges. However, Australian soil scientists insisted that the conditions at the time of prior stream activity were extremely arid. The apparent paradox was resolved when Schumm (1968) showed that the prior channels were formed by relatively high-gradient, low-sinuosity, coarse-sediment-transporting streams. From the proportional



expressions above it is clear that the discharge factor relates to parameters other than meander wavelength and channel width, as presumed in Dury's theory. Slope, sediment size, sinuosity and the width-to-depth ratio are all factors, and these combine to produce the result of prior channel development during a drier climatic period.

While much of the foregoing concerned a regime approach to fluvial palaeohydrology, there are other procedures. The sizes of bedload particles moved during past flow events can be related to various measures of the event magnitude, including flow velocity, bed shear stress and power per unit area of bed (Costa 1983). One can also determine the past stages of flow events from a variety of palaeostage indicators, including the study of flood slackwater deposits (Baker 1987). These various techniques have now achieved global application both for the practical study of flood risk assessment, and for the academic study of extreme river processes that defy direct measurement in the field.

## Lacustrine palaeohydrology

Ideally, the water balance for a closed basin can be described by the expression

$$dV/dt = d(P_L+R+U)/dt - d(E+O)/dt$$

where $V$ is the water volume in the lake, t is time, $P_L$ is precipitation input to the lake, $R$ is runoff from the tributary basins that feed the lake, $U$ is subsurface (groundwater) flow into the lake, $E$ is evaporation from the lake, and O is the subsurface flow out of the lake. For any given lake stage, the hydrological balance can be considered in equilibrium, so that

$$dV/dt = 0$$

Subsurface inflow and outflow are generally rather small for many lakes, or they may be very difficult to estimate. By ignoring these factors, the equilibrium water balance equation can be simplified in relation to the area of the lake $A_L$ and the area of the tributary catchment $A_C$ from which water drains into the lake, as follows:

$$A_C\, P_L + A_C\, (P_C k) = A_L\, E_L$$

where $P_C$ is the mean precipitation per unit area over the catchment, $k$ is a runoff coefficient such that $P_C k$ will equal the runoff per unit area from the catchment, $P_L$ is the mean precipitation per unit area over the lake, and $E_L$ is the evaporation per unit area from the lake. Usually only $A_L$ and $A_C$ are known for ancient lakes, leaving a problem in estimating the relative influences of evaporation versus precipitation on the overall lake balance, as follows:

$$A_L/A_C = P_C k/E_L - P_L$$

Note that the area of the lake can expand if the evaporation $E_L$ is reduced, if the runoff from the catchment $P_C k$ increases, if the precipitation over the lake $P_L$ increases, or if some combination of these changes occurs. Because evaporation depends on temperature and other climatic factors, its determination may require some independent means of estimating the past climate. Additional complexities occur for precipitation. Thus, the relative simple appearance of expressions for lake palaeohydrology can be misleading in regard to the problem of actually estimating ancient lake balances.

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SEE ALSO: cave; palaeochannel; palaeoflood; pluvial lake; prior stream; scabland; underfit stream

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## PALAEOKARST AND RELICT KARST

*Palaeokarst* refers to KARST landforms that are completely decoupled from the hydrogeochemical system that formed them, as distinct from *relict karst* that is removed from the morphogenetic situation in which it was formed, but remains exposed to and may be modified by present geomorphic processes (Ford and Williams 1989). The terminology associated with palaeokarst can be complex and ambiguous, but a definitive discussion and explanation is provided by Bosak *et al.* (1989). Figure 117 illustrates the main geomorphic relationships encountered.

Palaeokarst is usually found buried unconformably beneath other rocks, the cover beds being younger than the karst. This is sometimes referred to as *buried karst*. When the burial is relatively recent, it tends to be by unconsolidated allochthonous clastic sediments such as alluvial, volcanic, marine or glacial deposits. Relict karst is still subject to modification by modern solution processes beneath the covering sediments and tends to be only partly buried.

Old and deeply buried palaeokarst arises from tectonic subsidence. It can also involve geological deformation. The caprock constitutes a confining formation, and the palaeokarst is interstratified between it and an underlying non-karst formation. This is a form of interstratal karst, but unlike currently active interstratal karst, the palaeokarst is older than the confining cover

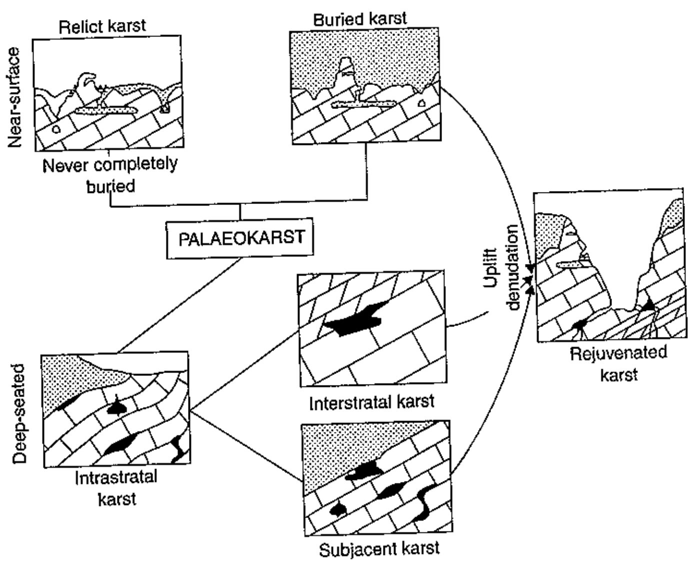


Figure 117 Main geomorphic relationships of palaeokarst



rocks and unconnected to the modern hydrogeological system. There is also an unconformity between the karstified rocks and the caprock. However, sometimes the palaeokarst is quite subtle and without major landforms, and is only recognizable by a disconformity in the carbonate sequence marked by a thin layer of insoluble residue. Such a situation represents a relatively brief interval of subaerial weathering followed by marine transgression.

A distinction is sometimes made between *interstratal* and *intrastratal* karst. The former develops along bedding planes and unconformities, whereas the latter is not restricted to such boundaries between strata. Karst beneath cover beds is sometimes referred to as *subjacent* karst, although if it is currently active it does not constitute palaeokarst. Deeply buried palaeokarsts serve as excellent traps for migrating hydrocarbons and contain some of the world's major oil and gas reserves. Later uplift may result in exposure of the cover beds to erosion and the exhumation of the palaeokarst. When this occurs it can sometimes be reintegrated into the modern hydrogeochemical system and therefore becomes rejuvenated.

Relict karst can arise in two ways: its hydrogeological context may change or its climatic (morphogenetic) situation may alter (Ford and Williams 1989). The first case is commonly found underground as a consequence of the incision of cave streams, because this leads to the de-watering and abandonment of high level cave passages, thus leaving them relict. They are not totally removed from the active hydrogeological system, because they remain in the vadose zone and receive percolation water and accumulate speleothems but, like river terraces in the case of surface rivers, they are removed from the streams that formed them.

The second case results from climate change on a timescale of $10^5$ years or more. Climate change associated with major latitudinal shifts of climatic zones has resulted in landforms developed under one morphogenetic system (say humid subtropical) being exposed later to radically different process conditions (perhaps arid, cool temperate or even periglacial). This can arise from global changes to the Earth's climatic system, as experienced in the transition from the Tertiary to the Quaternary, or from continental-scale movement over millions of years arising from plate tectonics, which can result in latitudinal displacement and in wholesale uplift of very large tracts of land (including its karst), such as in Tibet. This leads to the karst being forced out of equilibrium with its process environment. Such landscapes in a different morphogenetic context than the one in which they were developed are sometimes referred to as fossil karst. Shorter term climatic changes, as experienced in glacial–interglacial cycles, can also have profound effects on landscapes, exposing them to polygenetic conditions without necessarily making them relict, but forcing frequent readjustments to new process regimes.

Although most karst is developed by processes associated with the circulation of cool meteoric waters, some is produced by dissolution by hydrothermal waters and some by hot hypogean solutions associated with the intrusion of magma bodies. Deep subjacent karst is formed where heated water is circulated in a confined aquifer. These karsts are often encountered during mineral exploration, because the cavities produced are often heavily mineralized (Bosak 1989) frequently with sulphide ores. When removed from the situation in which they were produced (which was often at depth and many millions of years ago), such karsts constitute *hypogene palaeokarst*.

Palaeokarst is widespread throughout the world and occurs in carbonate rocks to at least Cambrian age. Contributors to the book edited by Bosak (1989) provide the best international review currently available.

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PAUL W. WILLIAMS

## PALAEOSOL

A palaeosol is a soil that formed on a landscape of the past (Retallack 2001). Soils are products of the physical, chemical and/or biologic weathering of sediments and rocks (see SOIL GEOMORPHOLOGY). Palaeosols typically occur at (1) major unconformities and (2) within basin-fill deposits representing aggradational systems. Although alluvial palaeosols are probably the most common type of palaeosol, they also occur in palustrine, aeolian, deltaic and coastal sediments, as well as in carbonate deposits (Kraus 1999). Palaeosols are especially abundant in Quaternary deposits and have been identified in rocks as old as 3.5 billion years (Retallack 2001).

Palaeosols are identified by a wide range of features including root traces, burrow fills, mottles, nodules, peds, clay films, cemented horizons (i.e. CALCRETE, SILCRETE, FERRICRETE), slickensides and matrix microfabrics. Concentrations of these features are used to identify palaeosol horizons, and individual profiles consist of vertically stacked horizons. Palaeosols should show vertical and lateral variations that mimic those observed in modern soils (see CATENA). One major difficulty in recognizing palaeosols is the effect of burial diagenesis. Diagenetic processes such as compaction, cementation and mineral transformations can significantly alter the texture, mineralogy and chemistry of palaeosols.

Palaeosols provide important records of past environments. Palaeosols at major unconformities are used to interpret past climates and changes in base level, and serve as important lithologic markers for correlating sedimentary deposits. In thick successions of sedimentary rocks, alluvial paleosols record the mode and tempo of basin filling (Kraus 1999). Weakly developed palaeosols are associated with rapid sediment accumulation rates and form close to ancient channel systems. In contrast, well-developed palaeosols reflect slow sediment accumulation rates and settings where sediment input is negligible. Rapid subsidence and sedimentation produce vertically stacked profiles whereas cumulative profiles reflect slow but steady rates of concurrent sedimentation and pedogenesis. Palaeosols also provide opportunities to study landscape development at a variety of spatial scales. At local scales, palaeosol properties vary according to changes in grain size and topography. At more regional scales, palaeosols reflect differences in climate, topography, tectonic setting and lithology.

One of the most promising applications of palaeosol research involves paleoclimate studies. Field-based and stable-isotope studies of iron- and carbonate-rich palaeosols have been used to document increases and decreases in atmospheric oxygen and carbon dioxide concentrations, patterns of global cooling and warming, and ancient mean annual temperature and precipitation (Retallack 2001). Mass balance studies of palaeosols have been used to quantify chemical weathering trends and ancient floodplain hydrology. Finally, palaeosols contribute to the record of ecosystem evolution. The colonization of land by plants and the development of forest and grassland ecosystems are recorded by the development of new palaeosol morphologies such as the mollic horizon (Retallack 2001).

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SEE ALSO: calcrete; catena; chemical weathering; chronosequence; climatic geomorphology; duricrust; ferricrete; silcrete; soil geomorphology

ANDRES ASLAN

# PALI RIDGE

A Hawaiian term for a steep slope or large cliff. Palis are steep-faced scarp ridges between stream valleys, commonly composed of basalt, and typically over 1,000 m in height. Various mechanisms have been suggested for their origin, such as being the eroded wall of a dissected shield volcano, being shaped by higher past sea levels, extreme fluvial downcutting, and catastrophic landslides. It is likely that several of these processes are involved in the formation of a pali.

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STEVE WARD

# PALSA

Palsas are small mounds of peat rising out of mires in the subarctic region characteristic to the discontinuous circumpolar permafrost zone provided that the peat layer is thick enough. They contain a permanently frozen core of peat and/or silt, small ice crystals and thin layers of segregated ice, which can survive the heat of summers. An insulating peat layer is important for preserving the frozen core during the summer. The peat should be dry during the summer, thus having a very low thermoconductivity, and wet in autumn, when freezing starts, giving a much higher thermoconductivity. This allows the cold to penetrate so deep into the peat layers that they do not thaw during the summer.



Palsas can be classified according to their morphology: dome-shaped, elongated string-form, longitudinal ridge-form, and extensive plateaux palsas as well as palsa complexes with many basins, hollows and ponds of thermokarst origin (Plate 85). The diameter of dome-shaped palsas ranges from 10 to 150 m and the heights from 0.5 up to 12 m. Longitudinal ridge-form palsas could be up to 0.5 km long and 6 m in height. Palsa plateaux rise 1–1.5 m above the surface of the surrounding peat surface and can cover areas of several square km.

Once a palsa hummock rises above the mire surface peat formation on its top ceases almost entirely. The surface peat on an old palsa is produced mainly by *Bryales* mosses, lichens and *Ericales* shrubs. It could also be by wind eroded old moss peat. Below the dry surface peat is the original mire peat formed by *Sphagnum, Carex* and *Eriophorum* remains. It is normally permanently frozen forming the permafrost core. In Finnish Lapland the summer thawing forms only a 50 to 60-cm thick active layer on the palsa surface. On the southern slopes of palsas the active layer gets deeper and on the edges the permafrost table is almost vertical. To date a palsa formation, samples should be collected from the contact of normal mire peat and of the dry peat formed on the palsa after its formation.

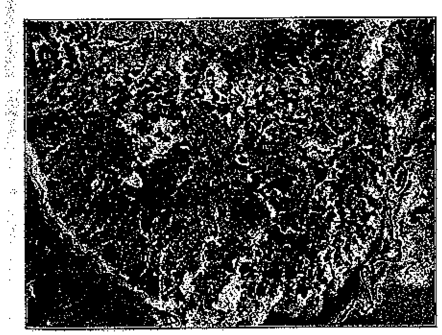


Plate 85 Aerial photograph (nr. 8634 17) of Linkinjeäggi palsa mire, Utsjoki, Finland. Published with the permission of Topografikunta

Low air temperatures together with low precipitation and a thin snow cover are found to be the most prominent factors for palsa formation. The hypothesis that palsas are formed in places with thin snow cover has been proved experimentally by cleaning the snow off the mire surface several times during three winters; a permafrost layer formed in the peat and a man-made small palsa.

Wind drift controls the thickness of snow cover on the mire surface. Thin snow cover allows the frost to penetrate deep into the peat, and in these places the frost fails to disappear completely during the seasonal thawing and part of it remains under the insulating peat. In the following winters the unthawed layer of frost becomes thicker and the mound starts to rise. The wind then carries away snow from the exposed hump more easily and the freezing process accelerates. The freezing front sucks moisture and segregated ice lenses are formed in the frozen core. This process increases the water content of the frozen core which can be 80–90 per cent of the volume.

The concept of cyclic palsa development is based on field observations and experimental studies in Finnish Lapland (Figure 118):

(A, B) The formation of a palsa begins when snow cover is locally so thin that winter frost penetrates sufficiently deeply to prevent summer heat from thawing it completely. The surface of the bog is then raised somewhat by frost processes.

(C) During succeeding winters the frost penetrates still deeper, the process of formation accelerates and the hump shows further upheaving due to freezing of pore water and ice segregation. As the surface rises, the wind becomes ever more effective in drying the surface peat and keeping it clear of snow.

(D, E) When the freezing of the palsa core reaches the till or silt layers at the base of the mire then the mature stage of palsa development begins. By this time the palsa stands well above the surface of the mire, typically displaying a relief of about 7 m in western Finnish Lapland.

(F) Degradation now starts, and peat blocks from the edges of the palsa collapse along open cracks into the pools which often surround the hummocks. During later stages,

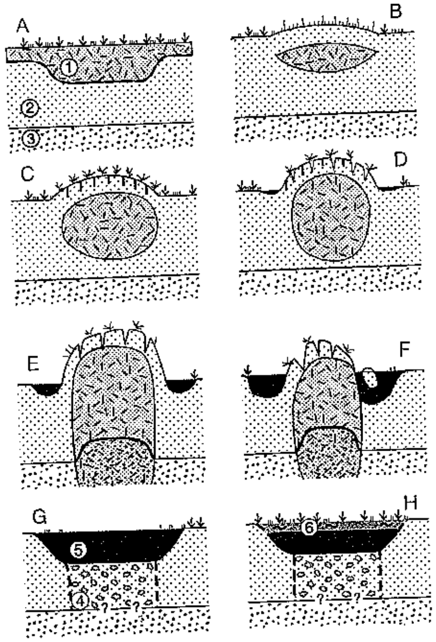


Figure 118 A general model of the formation of the frozen core (1) of a palsa in a mire (2) with a silty till substratum (3). A: the beginning of the thaw season. B: the end of the first thaw season. C: embryo palsa. D: young palsa. E: mature palsa. F: old collapsing palsa surrounded by a large water body. G: fully thawed palsa giving a circular pond on the mire (5). The thawed peat is decomposed (4). H: new peat (6) formation starts in the pond (Seppälä 1982, [illegible]

the vegetation may be removed so that the palsa surface is exposed to deflation and rain erosion.

(G) Old palsas are partially destroyed by thermakarst, and become scarred by pits and collapse forms. Dead palsas are unfrozen remnants: either low (0.5 to 2 m high) circular rim ridges; or rounded open ponds and pond groups; or open peat surfaces without vegetation.

(H) From such pools a new palsa may ultimately emerge after a renewed phase of peat formation, and the cycle of palsa development recommences from the beginning.

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MATTI SEPPÄLÄ

# PAN

[illegible]alled *playas*, *pfannen*, [illegible], [illegible], etc. are closed topographic depressions that are features of low-angle surfaces in the world's drylands (Jaeger 1939). Their characteristic morphology has often been likened to a clam, a heart or a pork chop. They are especially well developed on the High Plains of the USA, in the Argentinian Pampas, Manchuria, the West Siberian steppes and Kazakhstan, western and southern Australia and the interior of southern Africa (Goudie and Wells 1995). Pans evolve on susceptible surfaces. In southern Africa, for example, they are best developed on the sandy Kalahari Beds and on fine-grained Ecca shales. They also occur in particular topographic situations – deflated lake floors, old drainage lines, in interdune swales, in the noses of parabolic dunes and on coastal plains (e.g. the Carolina Bays of the eastern seaboard of the USA). They are sometimes, though by no means invariably, associated with LUNETTES (Sabin and Holliday 1995). They are often oriented with respect to regional wind trends, and tend in many cases to have bulbous lee sides. In areas like the Pampas, the High Plains of the USA and the interior of South Africa there are literally tens of thousands of pans, and they may cover as much as a quarter of the ground surface.

The origin of pans has intrigued geomorphologists for over a century. Hypotheses have included deflation, excavation by animals, and karstic (see DAYAS) and pseudo-karstic solution. Arguments on this issue are recurrent and recent years have seen some important contributions to the debate (e.g. Gustavson *et al.* 1995). What is becoming clear is that a range of processes has been involved in the initiation and maintenance of pans and that no one hypothesis can explain all facets of their own long histories and their variable sizes and morphologies.

An integrated model of pan development is as follows (Goudie 1999). First, pans occur preferentially in areas of relatively low effective precipitation. This predisposing condition of low precipitation means that vegetation cover is sparse and that deflational activity can occur. Moreover, once a small initial depression has formed, and the water in it has evaporated to give a saline environment, the growth of vegetation is further retarded. This further encourages deflation. The role of deflation in the removal of material from a depression may be augmented by animals, who tend to concentrate at pans because of the availability of water, salt licks and a lack of cover for predators. Trampling and overgrazing expose the soil to deflation and the animals would also physically remove material on their skins and in their bladders. Aridity also promotes salt accumulation so that salt weathering could attack the bedrock in which the pan might be located. It is also important that any initial depression, once formed and by whatever means, should not be obliterated by the action of integrated or effective fluvial systems. Among the factors that can cause a lack of fluvial integration are low angle slopes, episodic desiccation and dune encroachment, the presence of dolerite intrusions and tectonic disturbance. This model of pan formation is similar to that developed for the USA High Plains by Gustavson *et al.* (1995).



In addition to their occurrence in deserts, various types of oriented lake are also a feature of some tundra areas (Carson and Hussey 1962).

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A.S. GOUDIE

# PARAGLACIAL

Paraglacial is a term that was introduced by June Ryder (1971) to describe alluvial fans in the interior of British Columbia that had accumulated through the reworking of glacial sediment by rivers and debris flows following late Wisconsinan deglaciation. She mapped alluvial fan distribution throughout south-central British Columbia, noted that they were essentially inactive at present and concluded that they must have been dependent on the reworking of till, glacifluvial and glacilacustrine deposits by streams and debris flows in the earliest Holocene. She showed that fan accumulation was initiated soon after valley floors became ice free and continued until after the deposition of Mazama tephra (*c.*6,000 yrs BP).

The paraglacial concept was formalized by Church and Ryder (1972). They defined paraglacial as non-glacial processes that are directly conditioned by glaciation and added that 'it refers both to proglacial processes and to those occurring around and within the margins of a former glacier that are the direct result of the former presence of ice'. In this remarkable paper, they synthesized evidence from their field areas in

British Columbia (Ryder) and Baffin Island (Church) and they used the contemporary Baffin Island environment as an analogue for the early Holocene environment in south-central British Columbia. They concluded that, although fluvial sediment transport rates were likely to be greatest immediately after deglaciation, fluvial reworking of glacigenic sediments was likely to continue as long as such sediment was accessible to rivers. They identified three aspects of the influence of paraglacial sediment supply on fluvial transport: (a) the dominant component of reworked sediment may shift from till to secondary sources, such as alluvial fans and valley fills; (b) regional uplift will condition the timing of changes in the balance between fluvial deposition and erosion such that the cascade of sediment evacuation can be interrupted by sediment deposition; and (c) consequently, the total period of paraglacial effect is prolonged beyond the period of initial reworking of glacigenic sediments.

Slaymaker (1977) and Slaymaker and McPherson (1977) noted that in British Columbia primary upland denudation rates are low and that a large component of contemporary sediment load is derived from secondary remobilization of late Pleistocene and Holocene deposits. Slaymaker (1987) also showed that in the British Columbia and Yukon region, medium-scale (100–10,000 $km^2$) river systems exhibit the highest specific sediment yield, in contradiction to the conventional model of sediment yield vs basin area relations. Church *et al.* (1999) confirmed this result.

Church and Slaymaker (1989) emphasized the generality of the definition, specifying that it is applicable to all periods of glacier retreat and that a paraglacial period is not restricted to the closing phases of glaciation but may extend well into the ensuing non-glacial interval. The essence of the concept is that recently deglaciated terrain is often initially in an unstable or metastable state and thus vulnerable to rapid modification by subaerial agents. Effectively then the 'paraglacial period' is the period of readjustment from a glacial to a non-glacial condition. Different elements of paraglacial systems adjust at different rates from steep, sediment-mantled hillslopes (a few centuries) to large fluvial systems (>10,000 years). Increase in specific sediment yield with basin area for basins smaller than *c.*30,000 $km^2$ shows that specific sediment yield equals area raised to the power of 0.6. Isometry would dictate an exponent of 0.5 (because specific sediment yield can be reduced to a length dimension); hence sediment is recruited to streams at a rate greater than expected simply from an increase in area. The additional sediment is derived from erosion of both *in situ* and reworked glacigenic deposits along riverbanks. Effectively, these rivers are degrading through valley fill deposits forming entrenched trunk streams flanked by Holocene terraces. For basins greater than 30,000 $km^2$ specific sediment yield tends to decline as conventional models predict because non-alluvial riverbanks are protected from erosion.

These data demonstrate that the timescale of paraglacial sediment reworking in British Columbia includes the whole of Holocene time. Church and Slaymaker (1989) estimate that ultimate dissipation could take several tens of thousands of years. This implies that interglacial fluvial systems were still relaxing from the previous glacial period when the succeeding glacial period arrived. They also imply that there is no equilibrium between hydro-climate, denudation rate and sediment yield because all 'fluvial sediment yields at all scales above *c.*1 $km^2$ remain a consequence of the glacial events of the Quaternary'.

Owens and Slaymaker (1992) have examined sediment accumulation rates in three small lake-drained basins of less than 1 $km^2$ over the last 6,000 years and confirmed that these rates are 1–2 orders of magnitude lower than those of larger basins. Souch (1994) has traced the paraglacial signal downstream through a system of lakes progressively further from the glacial sources. Church *et al.* (1999) have expanded the analysis of suspended sediment yields across Canada and described seven Canadian regions with adequately monitored sediment data. Five of these regions were shown to have trends comparable with those of British Columbia; one, southern Ontario, is influenced by intensive land use disturbance and the data show no trend; one, the eastern Prairies, is a region of net fluvial aggradation and specific sediment yields decrease with basin area in accordance with conventional models. Evidently, paraglacial effects persist throughout the majority of Canada's regions.

Ballantyne (2002), in a magisterial summary and extension of the paraglacial concepts developed in British Columbia, points out that between 1971 and 1985, the paraglacial concept was largely ignored outside North America. Since 1985, he sees four trends: (a) an extension in the geomorphic contexts in which the paraglacial concept has been explicitly used; (b) a focusing of research on present-day paraglacial processes and land systems; (c) use of the paraglacial concept as a framework for research across a wide range of contrasting deglacial environments; and (d) a growing awareness of the palaeo-environmental significance of paraglacial facies in Quaternary stratigraphic facies. The working definition that he adopts for 'paraglacial' is 'non-glacial Earth-surface processes, sediment accumulations, landforms, land systems and landscapes that are directly conditioned by glaciation and deglaciation'.

The new perspective given by Ballantyne (2002) is most remarkable in its overview of the wide range of geomorphic contexts in which the paraglacial concept is already explicitly being used. These contexts are, in addition to the original debris cone, alluvial fan and valley fill deposits: (a) rock slopes; (b) sediment-mantled slopes; (c) glacier forefields; (d) glacilacustrine systems; and (e) coastal systems.

Wyrwoll (1977) was the first to identify rock slope response in a paraglacial context. Ice downwasting and retreat has resulted in the debuttressing of rockslopes and yields three responses: large-scale catastrophic rock slope failure; large-scale progressive rock mass deformation and discrete rock fall events.

The work of Ballantyne and Benn (1994) is significant in identifying sediment-mantled slopes in a paraglacial context. They note the processes of reworking sediment-mantled slopes yielding intersecting gullies, coalescing slope foot debris cones and valley floor deposits of reworked drift. Gully erosion and debris flow activity are the most obvious paraglacial processes invoked in this environment.

Matthews (1992) is credited with the first explicit identification of glacier forefields (forelands) in a paraglacial context. Effects conditioned by the former presence of a glacier include unconsolidated diamicton, steep slopes, unvegetated surfaces and the acceleration of mass movement, frost action, fluvial and aeolian processes.

Leonard (1985) was one of the early investigators of the paraglacial response of lake sediments. Such work accelerated in the 1990s and is now one of the most commonly used ways of assessing the changing rates of sediment production during the Holocene, specifically estimating the duration of the paraglacial effect in specific lake-drained basins.

The extension of the paraglacial concept to coastal systems is perhaps the most dramatic extension of the concept. Forbes and Syvitski (1994) defined paraglacial coasts as 'those on or adjacent to formerly glaciated terrain, where glacially excavated landforms or glacigenic sediments have a recognizable influence on the character and evolution of the coast and nearshore deposits'. They specifically exclude the effects of glacio-isostatic rebound and glacio-eustatic sea-level change on the grounds that these effects are more widely or even globally distributed.

It is clear from Ballantyne's discussion that the paraglacial concept has even wider significance than had previously been imagined. The data bring into question the possibility of any equilibrium or balanced condition in landscapes that have undergone Quaternary glaciation.

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SEE ALSO: alluvial fan; glacifluvial; glacilacustrine

OLAV SLAYMAKER

## PARALIC

Term referring to environments by the sea where shallow waters predominate, though nonmarine. Paralic environments are particularly associated with intertongued marine and continental deposits situated on the landward side of a coast. This includes lagoonal, littoral, alluvial and shallow neritic environments. Paralic sedimentation incorporates basins, swamps (paralic swamps), deltaic zones, heavily alluviated shelves and platform marshes. The word paralic is derived from the Greek word *paralia* meaning sea coast.

Paralic environments typically exhibit localized, abruptly changing facies tracts with a large variety of lithologies. The deposits are distinct by their thick terrigenous accumulations of clays, sands and silts (orthoquartzite to subgreywacke), intimately mixed with estuarine, marine and continental deposits. Paralic sediments can offer important stratigraphical information concerning the long-term changing coastal environment. Often the deposits are zones of subsequent coal formation (termed paralic coal), while petroleum accumulation is frequent within paralic basins. Paralic ecosystems are characterized by a large spectrum of biological species that are strictly bound to the particular environment, and are able to remain stable despite changing environmental conditions (Guelorget and Perthuisot 1992).

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STEVE WARD

## PARNA

A deposit of dust (suspended wind-blown mineral material) differentiated from LOESS by its higher clay content. The term was coined for deposits found in the interior of southeastern Australia and is attributed to an aboriginal word meaning 'sandy and dusty ground' (Butler 1956: 147). Parna can be regarded as synonymous with desert loess. High clay content (30–70 per cent) lead to differentiation of parna from the glacial loess of Europe, but is a feature of desert loess in Africa and elsewhere. High clay content, and particularly the inferred or observed presence of clay in the form of detrital aggregates, remains the main criterion for recognition of parna although the quartz fine sand or silt component is the most readily recognizable feature. These, and other, properties are inferred to arise from the origin of parna from the deflation of soils which have already experienced considerable weathering. Thus other deposits of clay-rich aeolian sediment, derived from deflation of lakes (see LUNETTE), for example, are not considered to be parna, although there is inconsistency in the application of the term. Other properties of parna, such as colour, calcium carbonate, salts, texture and structure vary with soil drainage in a catenary relationship. The depth and number of parna layers are variable and relate strongly to local topography and post-depositional erosion as well as proximity to the source areas. Parna layers were deposited during arid climate phases of the late Quaternary.

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PAUL HESSE

## PASSIVE MARGIN

In plate tectonic theory, oceans are spreading from mid-ocean ridges, and being consumed by



subduction at active margins. Passive continental margins are those that are not also edges of plates. They are also known as 'trailing edges' and 'Atlantic-type margins'.

They are presumed to be initiated as rift valleys, and when the rifts turn into oceans by seafloor spreading they become continental margins. The new margins may undergo some changes, but may also inherit landforms from pre-breakup times. In contrast to active margins which have many volcanoes, volcanicity is rare in passive margins: only east Australia has abundant volcanoes. In India the vast flows of the Deccan traps accompanied creation of the passive margin.

Based on morphotectonics there are two main types of passive margin: (1) passive margins without significant vertical deformation; and (2) passive margins with a marginal swell and Great Escarpment. We have no good explanation for the two types, or their distribution. Why does eastern Australia have a marginal swell-type margin, but most of the south coast is without vertical deformation? Why does most of southern Africa have Great Escarpments, while East Africa does not?

Passive margins without significant vertical deformation are formed by simple pull-apart of a continent. The Red Sea is an example of the early stage of the process. The Great Australian Bight exemplifies a later stage. Horizontal Tertiary limestones underlie the flat Nullarbor Plain, which is almost an old seafloor. In Patagonia (Argentina) the Atlantic is bordered by an extensive plain cut across ancient rocks. The offshore zone is characterized by many listric faults.

Margins with a marginal swell are the dominant type of passive margin (Ollier 2003), and include the Drakensberg, the Western Ghats, the Appalachians, parts of Greenland, Brazil, Antarctica, and elsewhere. The basic geomorphology of such margins is shown in Figure 119.

*Plateau* are upland areas with relatively flat topography and most are erosion surfaces. They may be extensive or dissected until only fragments are left. They occur on a wide range of rock types including horizontal strata, metamorphic rocks, granite and massive lava flow sequences.

The *marginal swell* is a widespread swell or bulge along the edges of a continent (*Randschwellen* in German; *bourrelets marginaux* in French). The whole land surface has been warped into an asymmetrical bulge, with the steeper slope

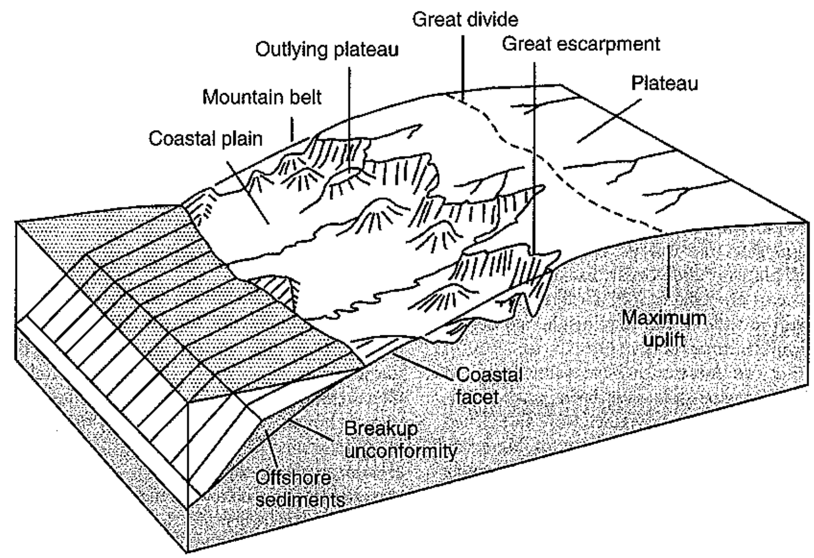


Figure 119 Morphotectonic features of a passive continental margin with marginal swell

to the coast (though the 'steep' slope is still only a few degrees). The marginal swell formed after the planation surface of the plateau, and after formation of major valleys.

*Great Escarpments* are scarps hundreds or thousands of kilometres long, and up to a thousand metres high. They occur on all sorts of rocks and are not structurally controlled. This is demonstrated especially in the Western Ghats of India, where the Great Escarpment in the north is cut across horizontal basalt, and continues with no change of form into the Precambrian gneisses and granites of southern India. Great Escarpments run roughly parallel to the coast, and they separate a high plateau from a coastal plain. The top of the Great Escarpment can be very abrupt. They are undoubtedly erosional. In places they are rather straight, but elsewhere can be highly convoluted. In some instances the top of the escarpment is the drainage divide between coastal and inland drainage (Brazil, Namibia); in other places the major drainage divide of the continental margin may be hundreds of kilometres inland of the Great Escarpment (east Australia). Many large waterfalls are found where rivers cross the Great Escarpment.

*Mountainous areas*, often quite rugged, form below the Great Escarpment, where the old plateau surface has been largely destroyed. Occasionally a patch of plateau is isolated to form a peninsula or isolated tableland.

*Coastal plains* lie between the mountains and the sea.

The landforms on such margins depend to some extent on present and past climates. In southern Norway the landscape is dominated by fjords and glaciated valleys, but the major features of plateau and Great Escarpment were present earlier. Glaciation straightened and deepened the valleys, but they originated before glaciation (Lidmar-Bergström *et al.* 2000). Greenland and Antarctica have some marginal swell-type margins that have been much modified but not obliterated by glaciation. In marked contrast is the Great Escarpment of Namibia, created by fluvial erosion but now in a largely desert environment.

A wedge of sediments is deposited offshore from the continental margin above an unconformity called the breakup unconformity (meaning related to the breakup of a supercontinent). The sediments record the history of uplift in their hinterland. In Scandinavia and southern Africa interpretation of the offshore sediments suggests that there were two main uplift phases – Palaeogene and Neogene. Individual river sources of sediment may be indicated: a delta developed after 103 Ma near the mouth of the present Orange River, South Africa.

There are two main models of passive margin evolution. One school, placing emphasis on fission track data and similar methods, believes there was a continuing uplift towards the margin, where the continental rim ended at a massive fault. Slope retreat moved from the initial faulted margin to the present Great Escarpment. The alternative is warping of the palaeoplain to below sea level. Valleys eroding the steeper, coastal side coalesced to form a Great Escarpment, which then retreats. This model equates the breakup unconformity with the plateau surface, and equates the marginal swell with the raised shoulder of present-day rift valleys (e.g. the Lake Albert rift, Uganda). This implies that the marginal swell dates back to the earliest days of continental breakup.

Some passive margins have simple drainage patterns with streams flowing in opposite directions away from a ridge at the top of the Great Escarpment (Brazil, Western Ghats). On some marginal swells major rivers were in existence before continental breakup and can still be traced in the modern landscape (Australia, South Africa). Drainage may be modified and even reversed. Original drainage divides may relate to the original tectonic movement that made a marginal swell, with the crest of the swell being the drainage divide. The location of divides can be modified by drainage evolution, especially headward erosion of coastward flowing rivers.

What caused uplift of the marginal swell is not known, though a wide range of proposals have been made, and it may not even be the correct question to ask. Some passive continental margins might have been high originally, like the high plateau bounding many present-day rift valleys. A secondary mechanism is isostatic adjustment to erosion of the land, loading by offshore deposition and (in places like Greenland and Antarctica) loading by ice sheets.

The geomorphic evolution of some margins has been traced back to the Mesozoic, as in eastern Australia. Elsewhere landscape evolution is



thought to be Miocene and younger, as in the Piedmont of the USA. There may be more than one period of movement. Several passive margins are now thought to have Mesozoic beginnings modified by further movement in the Neogene.

Several passive margins do not fall into the two groups outlined above. Some margins are dominated by deposition of sediment as the crust sinks. The Gulf Coast of the United States, and the coastal part of the Chaco-Pampean Plain in Argentina are examples. The country east of Perth in Western Australia is dominated by the north–south Darling Fault. This makes a topographic fault scarp, bounding an erosion surface cut across Precambrian rocks. West of the fault, the Perth Basin has about 11 km of Silurian to Cretaceous sediments showing long-continued downfaulting. The basin is not a rift valley as no fault further west has been located, so this is a faulted passive margin. The southern coast of Western Australia, west of the Great Australian Bight, exhibits a simple warp, bending the West Australian planation surface to sea level. Ancient valleys that flowed across the land from Antarctica before breakup can be traced across the warp as lines of salt lakes, and the drainage is reversed to form the present south flowing rivers (Clarke 1994). Despite the long time available (Antarctica started to separate about 55 Ma) there is no sign of initiation of a Great Escarpment, suggesting that it requires a greater relief than this margin offers.

Considering the relationship of big rivers, deltas and global tectonics, Potter (1978) pointed out that the twenty-eight biggest rivers in the world all drain to passive margins. Twenty-five of the world's largest deltas are also found on passive margins.

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SEE ALSO: active margin; isostasy

CLIFF OLLIER

# PATERNOSTER LAKE

A body of water set in a formerly glaciated environment, often divided by either moraine deposits and/or rock bars, and aligned with similar neighbouring lakes. They are often linked by a stream, rapids or waterfalls running through the valley so that it resembles, when viewed in plan, a string of beads. The term paternoster lake is derived from this pattern, with each lake resembling a paternoster (bead) in a rosary. Paternoster lakes are formed by the plucking and scouring out of a valley bed by a glacier, though they may also form through the damming effects of glacial deposits (by moraines, rock bars or by riegels). The varying rock resistance means that the glacier will erode away the weaker rock more quickly, forming depressions in the valley floor. Water then accumulates in these depressions upon glacial retreat, leaving a series of usually elongated lakes, reflecting the direction of scour from which they developed. The number, size and shape of paternoster lakes varies as a function of weakness, jointing and lithology of the underlying rock, alongside the varying characteristics of the glacier, valley steepness, and extending or compressive flow. Paternoster lakes are common in Sweden, draining into the Baltic. Also, Llyn Dinas and Llyn Gwynant, Snowdonia (UK), are examples of paternoster lakes.

STEVE WARD

# PATTERNED GROUND

Patterned ground consists of a range of phenomena – circles, nets, polygons, steps and stripes – developed in surface materials. Such phenomena occur in a wide range of environments and have a large number of causes.

Particularly in the seasonal tropics, swelling clay and texture-contrast soils develop micro-relief consisting of mounds and depressions arranged in random to ordered patterns. These are normally termed GILGAI. Most mechanisms of gilgai development involve swelling and shrinking of clay subsoils under a severe seasonal climate.

In some arid and periglacial regions patterned ground is in the form of DESICCATION CRACKS AND POLYGONS. These result because of the volume reduction that takes place in fine-grained, cohesive sediments as they dry out by evaporation of water. This creates sufficient tensional stress for rupture to occur and cracks to be formed.

Elsewhere in dry regions, patterned ground can be associated with the presence of salt, particularly on the floors of playas and on SABKHAS (Hunt and Washburn 1960). Thrusting structures can develop which are called tepees, because of their resemblance to the shape of the hide dwellings of early American Indians (Warren 1983).

In other dryland regions patterns can be produced by vegetation banding. From the air many dryland surfaces can be seen to be characterized by alternating light and dark bands called BROUSSE TIGRÉE (tiger bush). The banding reflects differences in the proportions of grasses and shrubs. This in turn is related to the action of sheet flow on low angle surfaces (0.2–2 per cent) in areas with 50–750 mm mean annual rainfall (Mabbutt and Fanning 1987).

Organic processes also create patterns through the building of mounds by such organisms as communal rodents and termites (see TERMITES AND TERMITARIA). In the case of the MIMA MOUNDS of the USA and the Heuweltjies of southern Africa their mode of origin is uncertain (Reider *et al.* 1996).

However, patterned ground (Plate 86) is especially prevalent in periglacial regions (see PERIGLACIAL GEOMORPHOLOGY; ICE WEDGE AND RELATED STRUCTURES) and in areas underlain by PERMAFROST. A great diversity of forms and processes are involved (Washburn 1956), including thermal contraction cracking, seasonal frost cracking and desiccation cracking. Circular forms are produced by FROST HEAVE (cryoturbation). Periglacial areas also show the development of Earth hummocks (Thufur) (Schunke and Zoltai 1988), and cryoturbation plays a role in their formation. Relict periglacial patterned ground phenomena developed during former cold phases are widespread in mid-latitude areas (Boardman 1987).

Plate 86 Late Pleistocene patterned ground developed under periglacial conditions in the Thetford region of eastern England. The stripes, which have analogues in present-day Alaska, are formed by alternations of heather (*Calluna vulgaris*) and grass

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A.S. GOUDIE

## PEAT EROSION

Peatlands cover large tracts of the microthermal northern hemisphere, in countries like Canada, Russia and Finland, but mostly these are low-lying and the peat remains largely intact. Upland blanket mire is much more rare and because of higher rainfall and greater slope angles, erosion is more likely. Some 8 per cent of the land surface of the British Isles is covered by blanket peat, mainly in the north and west. These blanket bogs form the largest single contribution (10–15 per cent) to a globally scarce resource. These areas of blanket peat are important for many reasons: water catchments, hill farming, shooting, recreation and landscape.

The blanket peat of the southern Pennines is unquestionably the most degraded in Britain with gullying affecting three-quarters of the blanket peat. These peatlands lie close to large urban areas (Manchester, Leeds, Sheffield), important sources of air pollution, and, compared to other areas of blanket bog, are climatically marginal, being more southerly than most and in areas receiving barely 1500 mm annual rainfall (Tallis 1997). Peat erosion has been studied over the last century, but it was largely Margaret Bower (1960, 1961) who stimulated recent work. She identified two types of gully system: a dense network of freely and intricately branching gullies on very flat ground (less than 3°); and, more linear gullies with much less tendency to branch on sloping land. Erosion rates from the heavily gullied blanket peat are high for the UK. Labadz *et al.* (1991) used reservoir sedimentation surveys to establish the long-term sediment yield: in total over 200 t $km^{-2} yr^{-1}$ including an organic fraction of almost 40 t $km^{-2} yr^{-1}$. These high sediment yields mean that many of the small reservoirs built in the nineteenth century are now largely full up with sediment and effectively useless for water storage. Whilst the peat erosion rates seem relatively small, given the low bulk density of peat, these do in fact represent large volumetric losses, implying that most of the gullies may have developed during the past three centuries.

John Tallis, in particular, has studied the history of peat erosion in the southern Pennines. His analysis of pollen profiles shows that there was drying out of the mire surface during the 'Early Mediaeval Warm Period' in the twelfth and thirteenth centuries. This was followed by a cooler, wetter period, and it seems possible therefore that climate change could have triggered gully erosion at that time (and perhaps in earlier dry phases too). More recently, human-induced pressures on the blanket peat have probably been more important, sometimes working in tandem with climate change. Fire (accidental or deliberate) and overgrazing by sheep are the most important direct pressures leading to erosion; the loss of pollution-sensitive mosses, particularly *Sphagnum*, is also likely to have been significant. Complete loss of *Sphagnum* soon after the start of the Industrial Revolution in the eighteenth century may well have initiated more widespread gully erosion than might have developed because of climatic change alone.

In intact peat, the water table remains close to the surface except during severe droughts. Most storm runoff is produced by saturation-excess overland flow therefore, although locally pipeflow may be important (Holden and Burt 2002). On flat ground a hummock and pool micro-topography often develops. If the peat dries out, gullies begin to form, further lowering the water table, especially during summer. As the peat re-wets in autumn, there is an increased tendency for leaching of dissolved organic carbon (DOC) discolouring local water supplies and leading to significantly increased costs of water treatment (Worrall *et al.* 2002). Together, enhanced export of particulate and dissolved carbon means that the blanket peat no longer continues to build up a store of carbon and increasingly becomes a source of carbon export instead. From the 1950s, many areas of blanket peat were drained (using narrow slot drains or 'grips') in an attempt to increase productivity. More recently, landowners are beginning to fill in the grips, in an attempt to restore habitat and reduce DOC export.

Further north in the Pennine Hills, there are clear signs today (2002) of revegetation of previously heavily eroded blanket peat. This may indicate that previous pressures leading to erosion have been reduced. In the southern Pennines, some gullies have begun to infill over the past twenty years but generally there is little sign of revegetation, perhaps showing that the combined influences of sheep grazing and air pollution continue to hinder recovery there.

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TIM BURT

## PEDESTAL ROCK

A pedestal rock is an isolated erosional rock mass comprising a slender stem, neck or column supporting a wider cap. Also known as mushroom rocks balanced rocks and perched blocks, by local names such as *loganstones* (south-west England) and *hoodoo rocks* (North America), and by non-English equivalents such as *rocas fungiformas, roches champignons, Pilzfelsen*, pedestal rocks are developed in various climatic and lithological contexts; but especially well in sandstone, granite and limestone.

They are due to differential weathering and erosion of the cap and stem. Some are structural, the caprock being inherently more resistant than that of the stem. Pedestal rocks have been attributed to various epigene effects, and certainly, some standing in rivers and on the coast, especially where limestone is exposed, are due to physical, biotic and biochemical attack around water level. The occurrence of pedestal bedrock shapes just below the surface, however, shows that moisture attack there produces incipient indents, alcoves and concave shapes in the bedrock surface. Subsurface weathering all around the base of a block or boulder followed by lowering of the surrounding area produces a mushroom form.

Epigene effects such as differential wetting and drying on different aspects contribute to development and maintenance of the form after exposure. Sand blasting may be responsible for pedestal rocks in an immediate sense, but may exploit bedrock already weakened by weathering. Pedestal rocks are convergent forms, but most are of two-stage or etch origin.

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C.R. TWIDALE

## PEDIMENT

A pediment is a gently inclined slope of transportation and/or erosion that truncates rock and connects eroding slopes or scarps to areas of sediment deposition at lower levels (Oberlander 1989). They have been reported from polar, humid and arid zones (Whitaker 1979), but they are most widely reported and studied in dryland environments, and are generally perceived as a phenomenon of DESERT GEOMORPHOLOGY.

Pediments are part of a family of landforms developed in the piedmont zone, an area of diverse geomorphology juxtaposed between uplands dominated by sediment erosion and lowlands dominated by sediment transport and deposition. Piedmonts may, therefore, be subjected to EROSION, transport or depositional process domains. These domains can vary spatially and temporally, giving rise to a complex variety of piedmont landforms, including ALLUVIAL FANs, BAJADAs and pediments.

Pediments have been variously defined in the literature. These definitions range from very general, such as 'a terrestrial erosional surface inclined at a low angle and lacking significant relief' (Whitaker 1979), to more specific, such as 'surfaces eroded across bedrock or alluvium, usually discordant to structure, with longitudinal profiles usually either concave upward or rectilinear, slope at less than 11°, and thinly and discontinuously veneered with rock debris' (Cooke 1970). Although gradients can range between 0.5° and 11°, pediments steeper than 6° are rare in the natural environment (Dohrenwend 1994). Where two pediments meet across a divide, breaking up the continuity of a mountain mass, a pediment pass is formed, often creating useful trafficable routes through upland regions. A pediplain is formed by the coalescence of numerous pediments. It should be distinguished from a PENEPLAIN, where slope decline is thought to be the major process, rather than slope retreat.

Pediments were first described by Gilbert (1877), but the term 'pediment' was coined by McGee (1897), and is derived from the architectural term for a low-pitched gable, especially the triangular form used extensively in classical architecture. Subsequently, the term has been applied to a variety of geomorphological forms, giving rise to considerable confusion and problems of definition (Whitaker 1979).



Difficulties are encountered when trying to define the outlines of pediments for the purpose of GEOMORPHOLOGICAL MAPPING. This also makes it difficult to derive their MORPHOMETRIC PROPERTIES, such as length, area, mean gradient, etc. (Cooke 1970). The upper margin is generally agreed to be at the piedmont angle (the junction between pediment and mountain front, usually defined as the line of maximum change in gradient in the slope profile), or at the watershed if a tributary upland area is absent. However, the downslope margin is more difficult to define. Cooke (1970) suggests this boundary should be placed where alluvial cover becomes continuous; Howard (1942) and Tator (1952, 1953) suggest it should be placed where the depth of alluvial cover equals the depth of stream scour (15 m). Other researchers have placed this boundary where the thickness of alluvial cover exceeds a small proportion (e.g. 1 per cent) of total pediment length (Dohrenwend 1994).

A variety of pediment types have been recognized and classified. Three different pediment forms can be identified using simple geomorphological criteria; an apron pediment is the common form that extends between an upland source area and a lowland depositional area; a pediment dome is formed by coalescing pediments, when the upland area has been removed; a terrace pediment is developed adjacent to a relatively stable base level such as a through-flowing stream. Other classifications distinguish between covered and exposed forms: a mantled pediment is one where crystalline bedrock is veneered by a residual weathering mantle and which is inferred to have been formed by subsurface weathering of the crystalline bedrock and wash removal of the resulting debris; a rock pediment is thought to be formed by removal of the overlying debris from a mantled pediment; and a covered pediment is developed discordantly across sedimentary rocks, having a veneer of coarse debris.

Oberlander (1989) makes an important distinction between two fundamental types of pediment; those that truncate softer rocks adjacent to a more resistant upland, and those where there is no change in lithology between upland and pediment. The first form has been widely reported, most notably along the northern margin of the Sahara Desert. These landforms have been widely studied by French geomorphologists, who term them GLACIS D'ÉROSION. These landforms truncate weak materials, and tend to be veneered by alluvial gravels, indicating the importance of fluvial processes in their creation. The second form, which has proved much more difficult to explain, is referred to by Oberlander (1989) as a 'true' pediment. Here, the pedimented surface has been cut across a lithology of similar resistance to the adjacent upland, usually a relatively resistant igneous or metamorphic rock. They typically lack the alluvial cover of the glacis-type, and show little clear evidence of fluvial processes in their formation. In the absence of an obvious mechanism a number of theories have been proposed for their formation and development, but they remain ill-defined and controversial.

At a basic level, pediments are normally viewed as a result of erosion of upland areas. This material is transported across the pediment into the lowland depositional area, and the retreating upland leaves behind an enlarging transportational pediment surface. However, problems arise when attempts are made to identify specific pediment-forming processes. Numerous processes have been proposed, but their significance is very difficult to demonstrate in practice. These proposed pediment-forming processes can be considered under three headings; surface WEATHERING, subsurface weathering and fluvial processes.

Surface weathering processes cover a wide range of SUBAERIAL processes leading to the breakdown of bedrock and REGOLITH. However, these processes do not explain the formation of a distinct piedmont angle on rocks with the same resistance to weathering. This has led Mabbutt (1966), and others, to emphasize the importance of subsurface weathering in the formation of pediments. This is largely based on the observation that pediments are widely developed on granitic bedrock, a lithology particularly susceptible to subsurface weathering. Perhaps most important here is the nature of the material produced by deep weathering of granite. The well-sorted, sand-sized GRUS forms non-cohesive channel bank material that are highly susceptible to lateral channel shifting and planation, resulting in a limited amount of channel incision (Dohrenwend 1994). The fine-grained grus can also be transported down the low pediment slopes. Mabbutt (1966) attributes the formation of a piedmont angle to slope foot notching (weathering in the subsurface layer at the base of the mountain front). However, much of this model of formation is based on assumptions based on form and occurrence, rather than on observed and well-characterized processes that can be easily validated.

Fluvial processes have been widely implicated in pediment formation, with the major emphasis being given to lateral planation. Streams debouching from the upland drainage basins are thought to erode back the mountain front by lateral channel migration. Channel incision is thought to be limited due to the high sediment load of these streams. Other research has focused on the importance of sheet floods, but this process occurs rarely in the natural environment, and its significance in pediment formation must be questioned. Sheet flooding cannot produce a planar surface, because a planar surface is necessary for sheet flooding to occur (Cooke *et al.* 1993). This vital distinction between pediment-forming and pediment-modifying processes was emphasized by Lustig (1969), who suggested that contemporary pediments were the wrong place to look for an explanation of how they were formed, since the pediments would already have to exist for these processes to operate. He suggested that geomorphologists should instead concentrate on studying the erosional processes operating in the adjacent upland drainage basins, as this is where erosion is most active. Other workers have suggested that much of the erosion leading to formation of pediments takes place in embayments, formed where streams debouch from the upland area (Parsons and Abrahams 1984). As with the subsurface weathering model detailed above, the fluvial models must be treated with caution due to the difficulties of linking observed forms with clearly defined physical processes.

The main difficulty in explaining development of pediments is the problem of maintaining parallel rectilinear retreat of permeable slopes in a SAPROLITE-mantled landscape. Oberlander (1989) proposes that rectilinear retreat occurs because sediment transport processes are limited by deep permeability of grus, eluviation of fines by throughflow, and accelerated subsurface weathering by soil moisture, concentrated at the base of slopes. Twidale (1978) suggests that lithological and structural features within granitic massifs (petrological variations and differences in joint density) are important controls on pediment morphology, but other work has failed to demonstrate any clear relationships. The importance of tectonics in pediment formation is also uncertain, although, in general, pediments appear to be best developed in areas of long-term stability (Dohrenwend 1994).

With improvements in dating techniques, there is a growing amount of evidence indicating that some pediments are extremely ancient. In the Sahara and Mojave desert (Cooke and Reeves 1972; Plate 87), lava flows can be seen to bury existing pediment surfaces. This raises the possibility that they may be relict landforms formed under different climatic conditions pertaining in Tertiary or even Late Mesozoic times (Oberlander 1989). Specifically, the arid-zone processes acting on contemporary desert pediments may not be appropriate to explain landforms that developed over timescales that embraced humid as well as arid phases. A variety of conditions have been proposed as being optimal for pediment development of crystalline rocks; these include seasonally wet, low-latitude forest, savanna and cold-winter deserts affected by cryogenic processes. Oberlander (1989) suggests that pedimentation is currently active in parts of central Arizona, which appears to replicate conditions in the Mojave Desert in the Miocene. It seems likely that many pediments must be regarded to some extent as relict landforms, currently being modified under very different environmental conditions from those that pertained during their initial formation.

Plate 87 Pediment in the Mojave Desert, southwest USA

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SEE ALSO: alluvial fan; desert geomorphology; glacis d'érosion

KEVIN WHITE

# PENEPLAIN

Peneplain is the term coined by W.M. Davis to mean a surface of low relief worn down to near sea level and formed through erosion over protracted spans of time. His own words are:

> Given sufficient time for the action of denuding forces on a mass of land standing fixed with reference to a constant base-level, and it must be worn down so low and so smooth, that it would fully deserve the name of a plain. But it is very unusual for a mass of land to maintain a fixed position as long as is here assumed.... I have therefore elsewhere suggested that an old region, nearly base-levelled, should be called an almost-plain; that is a peneplain.
>
> (Davis in Chorley *et al.* 1973: 190)

The peneplain is thus not the end-product of a cycle of erosion and, if keeping with Davis's way of thinking, it should not be confused with an endless and featureless plain as is often implied. Rather, a peneplain is a regional landscape at the penultimate stage of development which is yet to be eroded down to a true plain. In another place Davis himself says:

> At a less advanced stage of degradation, the land will still possess low, unconsumed hills along the divides and subdivides between the broad-floored rivers. It will then be almost-a-plain, or a peneplain. A peneplain will be hardly above sealevel at its base, but if the area is large it may attain altitudes of 2,000, 3,000, or 4,000 feet far inland near the river heads, and its residual mounts and hills may rise still higher, although with gentle slopes.
>
> (Davis in King and Schumm 1980: 8)

The processes leading to a peneplain would be mainly subaerial, chiefly fluvial and gravity-driven hillslope processes. They ought to be in action long enough to obliterate the effect of unequal rock resistance, so only the hardest rocks would form bedrock-built hills rising above the peneplain, the monadnocks. Otherwise, gentle slopes would be underlain by deeply weathered rock, with the thickness of weathering mantle being in excess of 10 m. As far as the relative relief of a peneplain is concerned, Davis seemed to be rather vague in defining any critical hill heights or slope angles. Therefore, there are two crucial characteristics of a peneplain. One is its temporal context within the cycle of subaerial erosion. To be a peneplain, the surface of low relief must have formed in the course of protracted denudation. The second prerequisite is grading down to sea level.

Much of the substantial confusion around the term results from the fact that subsequent workers did not always keep with Davis's original definition and used the term in various contexts. For example, peneplains were often equated with PLANATION SURFACES, or a particular mode of slope evolution was implied for a peneplain. Many geomorphology textbooks contrast peneplains formed mainly through slope downwearing and consequent relief reduction, with pediplains formed by slope backwearing and relative relief maintained high until a rather late stage of development. In other cases the condition of being located close to, or graded to, sea level was ignored. In consequence, flattened summit surfaces within mountain ranges were frequently called peneplains despite the fact that neither their origin nor the age were known sufficiently to warrant the use of the term in its strict, Davisian sense. Davis himself suggested the term 'pastplain' to describe a peneplain which has been uplifted and now shows the initial stage of dissection.

Free usage of the term and its obvious connection with the Davisian model of cyclic landform development and the DENUDATION CHRONOLOGY approach, themselves strongly criticized since the 1960s, had eventually led to its declining popularity and gradual abandonment. Preference was given to more neutral 'planation surfaces' in describing landscapes, whereas in the field of theory a search for non-cyclic models of geomorphic development was pursued strongly.

Nonetheless, Fairbridge and Finkl (1980) proposed to return to 'peneplains', but realizing the potential for confusion and misuse they suggested disassociation from restrictions implied by the Davisian definition. Instead, they preferred to give the term a non-genetic meaning, simply to describe a near flat surface regardless of its origin, setting and evolutionary stage. This point has apparently been taken forward by Twidale (1983) who describes peneplains as 'rolling or undulating surfaces of low relief', without referring to their position in respect to BASE LEVEL. Moreover, he argues that there is no means to decipher the mode of past slope evolution leading to the present-day peneplain; hence the argument focused on the backwearing-or-downwearing issue is by and large pointless. On the other hand, he firmly adheres to the Davisian understanding of the peneplain as a landscape at the penultimate stage of evolution and introduces 'ultiplains' as the true end-products of relief development. By contrast, Phillips (2002) in his most recent review offers a broader definition of the peneplain in which the condition of being at any certain stage of a cycle has been made redundant. Given all these divergent views, the term is very difficult to recommend for routine application in describing and explaining landforms.

There has been much debate as to whether peneplains and peneplanation, in the truly Davisian sense, really occur. Phillips (2002), following his many predecessors, claims that contemporary peneplains eroded to near sea level are almost non-existent and seeks the reasons in constant variations in tectonic forcings, climate and base level, especially in the Quaternary. All these changes would not allow peneplanation to last for long, and induce surface dissection rather than planation. On the other hand, Twidale (1983) gives a number of examples of almost perfect rock-cut plains, but demonstrates their antiquity at the same time. Many of these plains date back to the Cretaceous or even beyond. In Fennoscandia, a peneplain of subcontinental extent undoubtedly existed at the end of the Precambrian (Lidmar-Bergström 1995). Further examples of past plains, or palaeoplains, have been reviewed by Ollier (1991). It appears that reconciling the evidence for little or no peneplanation at present and widespread planation in the geological past is one of the challenges for evolutionary geomorphology.

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SEE ALSO: Cycle of Erosion; denudation chronology

PIOTR MIGOŃ

# PERIGLACIAL GEOMORPHOLOGY

Perhaps surprisingly, there is no agreement as to what exactly constitutes terrain which can be regarded as periglacial for there are no quantitative defining parameters which have gained universal acceptance. However, most would accept the proposition that there are two possible approaches to the demarcation of what is periglacial, and that both can be justified.



One would emphasize the requirement for intense frost action in the form of frequent FREEZE–THAW CYCLES and deep seasonal freezing. If these criteria are used to delimit the distribution of the periglacial domain, it follows from this that some 35 per cent of the Earth's continental surface (mainly in the northern hemisphere) falls into the periglacial category. The other approach would stipulate that the presence of perennially frozen ground, i.e. PERMAFROST, is paramount. Permafrost may be defined as any earth material that has maintained a temperature at or below 0 °C for a minimum period of two years. Note that there is no reference to water content or lithology in this definition. If permafrost is the fundamental attribute, then a more rigorous climate than that needed for frost action alone is required to qualify for periglacial status, as permafrost demands a mean annual temperature of below 0 °C. As a result the global periglacial area would be substantially less at around 20 per cent of the total terrestrial surface. Nevertheless, from both perspectives the total area regarded as periglacial remains a considerable part of the Earth's terrestrial environment. By way of comparison, the area of permanent snow and glacial ice is only 3 per cent.

Relief also has an influential role in determining the distribution of periglacial regions. Both freeze–thaw cycles and permafrost are related to climate and this is influenced by both latitude and altitude. The outcome is that the most widespread periglacial areas are mainly lowlands in northern Eurasia and North America and these incorporate both tundra and boreal forest landscapes. Mountain temperatures are influenced by elevation which is sensitive to the lapse rate. This can produce sufficient cooling at lower latitudes to over print the more temperate conditions in the adjacent lower areas. As a consequence alpine periglacial areas can occur even at the equator. Usually they reveal a tundra zone with the lower limit roughly corresponding with the upper treeline.

## Basic periglacial processes

Periglacial geomorphology focuses primarily upon those terrestrial surface processes, sediments and resultant landforms which characterize the cold non-glacial areas of the Earth's surface. Basic to comprehension of these is a knowledge of the somewhat anomalous physical behaviour of water substance. First, there is a 9 per cent increase in volume as the phase changes from liquid to solid and conversely by a decrease from solid to liquid which occurs during freezing and thawing. This is accompanied by a large latent heat of fusion (84 calories per gram) not much less than the heat required (100 calories per gram) in changing the temperature of the liquid phase from the solid to gas transitions. The net result is that the rate of both freezing and melting are delayed more than might be expected. Second, volumetric changes occur during temperature variations in the solid (frozen) state when cooling produces contraction and vice versa. This in itself is not unusual but to avoid confusion it has to be seen as being totally independent on the 9 per cent shift at the liquid–solid–liquid phase change. Third, maximum density is achieved in the liquid phase, some 3.98 °C above the freezing point, ensuring that the solid phase floats on the liquid. This has profound implications for life as it means that water is present beneath lakes and rivers with ice covers. Even in the harshest climates water bodies over 3 m deep do not freeze to their beds as annually developed ice rarely exceeds 2 m in thickness. Fourth, within the sediment pore the freezing point of water can be lowered down to −22 °C in extreme cases. This is especially effective in fine-grained sediments (clays and silts) where the movement of thin films of water occur even though the ground is technically frozen. This facilitates the aggradation of ice masses with volumes well in excess of the pore capacity. All these factors contribute to the landscape-forming processes associated with periglacial environments and collectively these determine the nature of periglacial processes. Driving these processes is temperature change and in its turn weather and climate.

## Palaeoperiglacial activity

Apart from the unusual physical behaviour of water substance, a further complicating factor in understanding periglacial features is the temporal dimension, particularly that allied to climatic change. This can be illustrated by the example of Britain. If delimiting areas subject to periglaciation on the basis of freeze–thaw cycles is accepted, then the higher summits of Britain remain within the ambit of periglacial activity. This is the viewpoint taken by Ballantyne and Harris (1994) in their major regionally based synthesis. However, adoption of the alternative less permissive approach based on the presence or absence of permafrost, inevitably means that the British Isles cannot be regarded as periglacial since the last permafrost finally dissipated some 11,500 years ago towards the end of the Last Glacial stage. Since then through the ensuing Flandrian interglacial (postglacial) even the climatically most extreme locations have been unable to sustain any permafrost. This was the position taken by Worsley (1977) in reviewing British periglaciation when it was concluded that all the periglacial evidence in Britain was effectively relict.

Prior to 11,500 years ago, during the Last Glacial stage, most of Britain had experienced episodic extensive permafrost. Indeed, in many areas of the world the Last Glacial stage witnessed the dramatic expansion of the periglacial realm by up to 50 per cent. But this was probably never so dramatic as in western Europe where the mean annual temperature dropped by some 20°C during the time when the Gulf Stream was inoperative. An underemphasized aspect of the global glacial stages is that if sea level is taken as a proxy for palaeoclimate then the durations of very low sea levels (maximum glacial ice volumes) were relatively short. Similarly Quaternary marine oxygen isotope ratio records are interpreted as reflecting in large part the degree of global glaciation and the negative ratio peaks in the curves are taken to correspond with the periods of most extensive glacial ice cover (corresponding to the lowest sea levels). In many areas of the world covered by glaciers in the Last Glacial stage, the stratigraphic evidence indicates that the glacial advance which culminated in the most extensive ice cover occurred late in the stage. These data imply that for much of the glacial stage those areas which were to become glaciated were cold but non-glacial in character, i.e. periglacial processes rather than glacial processes prevailed. Hence many of the glaciated landscapes bear a partial periglacial imprint. Naturally those areas immediately outside the maximum ice extent limits witnessed a periglacial regime for much of the glacial stage and hence the effects of periglaciation are clearer. Finally, the earlier phases of ice retreat from the maximum limits were primarily the result of reduced snowfall rather than a temperature increase. This enabled the perglacial environment to extend into the areas recently vacated by the ice. PARAGLACIAL conditions follow ice withdrawal and to an extent these might be regarded as part of periglaciation.

## Historical development of the periglacial concept

The term periglacial was first coined in 1909 by the Polish geologist and pedologist W. Łoziński in his account of the mechanical weathering of sandstones and blockfield (see BLOCKFIELD AND BLOCKSTREAM) production under inferred cold climates in the Carpathian Mountains. Three years later Łoziński introduced the concept of a periglacial facies produced by mechanical weathering, although he did not give it any quantitative climatic parameters. However, it is clear that Łoziński was proposing the periglacial concept in an attempt to reconstruct the palaeoenvironmental context of his facies. He envisioned it as diagnostic of the former processes operative in terrain immediately adjacent to glaciers and ice sheets of Pleistocene age rather than as a function of contemporary activity. His second paper was published as part of the proceedings of an international geological congress held in Stockholm in 1910 and this ensured that the term periglacial was widely disseminated. Other activities at the congress included a field excursion to Spitzbergen where periglacial facies could be related to contemporary environmental processes, and thereby gave further impetus to the scientific study. Strictly therefore, periglacial, as originally envisaged by Łoziński, should refer to the area or zone formerly subject to arctic-type climatic conditions peripheral to a glacier.

From the standpoint of modern usage, an erroneous impression might be given that periglacial features are exclusively associated with the area around glacial margins. On the contrary, some of the major areas of permafrost today have never been glaciated or indeed been peripheral to former ice sheets. The prime example of this is east Siberia where the permafrost can exceed over 1 km in depth. This is probably explained by the fact that it has experienced the longest history of sustained permafrost development anywhere on the Earth. A further disadvantage is that there might be the assumption that areas peripheral to glaciers experience a rather less severe climate and that a climatic deterioration would necessarily lead from the periglacial to glacial, with glaciation representing the ultimate severe climatic state. Despite a number of workers having argued for the term periglacial to be abandoned because of its imprecision, its usage is now widespread and a degree of permissiveness in its definition is automatically accepted by its users. It is interesting to note that there was a change in the title between the first and second editions of A.L. Washburn's synthesis of periglacial environments (Washburn 1973, 1979). In the latter the term GEOCRYOLOGY was introduced (i.e. the investigation of frozen Earth materials). The word is derived from the Russian equivalent and although it can include glaciers it is usually directed towards frozen ground.

Although Łoziński formalized the notion of a periglacial zone early in the twentieth century, as with many concepts in Earth science earlier workers anticipated later formulations. For example, the commencement of government-sponsored geological mapping in England in the 1830s soon led to the identification of a relict mantle of rock debris by De la Beche overlying the slopes of Cornwall and Devon in the south-west peninsula. Field relationships demonstrated to him that this 'head' blanket of angular fragments had been derived from mechanical weathering of the bedrock cropping out on the slopes above and that there had been an ubiquitous downslope movement which had tended to even out any terrain irregularities, thereby anticipating something akin to SOLIFLUCTION.

## Unique periglacial processes and landforms

The research literature arising from investigations of periglacial geomorphology has given rise to a wide range of specialized terms and names of unique landforms. These have come from a number of language sources and some duplication and confusion have arisen because of inconsistent usage.

This was exemplified in the planning of this encyclopedia since the draft list of topics specified both altiplanation and CRYOPLANATION as separate entries. In reality there is no difference between the two. Altiplanation terraces were first described by H.M. Eakin in 1918 following field mapping in part of eastern Alaska where he encountered benches and summit surfaces cut in bedrock largely independent of lithology and structure. Similar relationships had earlier been identified in Russia and the term 'goletz' terrace applied to them. Other terms which have been used include equiplanation and nivation terraces. Bryan (1946) undertook a wide-ranging review of the then existing periglacial terminology and amongst others proposed a new term – cryoplanation – to express the unified concept of frost action and frost-related downslope movement of debris to produce a degradation system eroding and lowering hillslopes. This is now the internationally agreed term for such features.

Fortunately the confused nomenclature has been subject to clarification in a comprehensive glossary produced by a very experienced interdisciplinary team of Canadian permafrost workers (Harris *et al.* 1988). This is an excellent source of current usage, definitions and synonyms used in periglacial geomorphology. It also has the additional merit of thoughtfully discussing many of them and the authors are not reticent in recommending the abandonment of some cherished terms!

Periglacial geomorphogy, like any morphogenetic geomorphology, should consider all the geomorphological agents which contribute to the landscape character. Naturally there is a tendency to concentrate upon those agents which are either unique to or are readily associated with it at the expense of those which are common to a range of environments. To illustrate this point there are over fifty entries in this encyclopedia which are relevant to periglacial geomorphology. Yet only half of these are likely to be discussed or referenced to the periglacial realm. Examination of most periglacial environments, in the field or from maps and air photographs, normally reveals an essentially fluvial landscape displaying a 'normal' drainage network. There are some exceptions and significant parts of the periglacial regions display desert landscapes. This is not surprising considering some of the most arid areas of the world are underlain by permafrost. But these deserts are cold. There is tendency to regard deserts as hot places for this is where the vast majority of desert geomorphologists work.

## Applied periglacial geomorphology

Wherever there is ice within the subsurface there is always the possibility that it might melt. Under natural conditions this is an ongoing process and can occur through a range of incidents such as forest fires, coastal and riverbank erosion, or climatic amelioration. Indeed the prospect of global warming carries severe implications for the entire permafrost world.

Over the last century there has been progressive settlement of the permafrost terrain by people from more southerly regions who had an expectation of a similar range of facilities to those south of the permafrost region. This movement was given a particular impetus by the Second World War and operation of defence facilities during the Cold War. Later economic exploitation of mineral resources and hydrocarbon exploration placed further demands for transport, urbanization and allied installations.

Construction of all kinds on permafrost terrain is potentially hazardous if the natural ground thermal equilibrium is disturbed as this will induce melting of the ground ice and cause thaw consolidation. Under the stress of war a number of mistakes were made in road and pipeline construction but experience was gained from tackling the challenges presented by permafrost. A landmark publication was the compilation by Muller (1947) of the then state of the art understanding of permafrost and its allied engineering problems. This drew extensively upon Russian experience. It led to the founding of the US Army's Cold Regions Research and Engineering Laboratory which has subsequently been one of the leading institutions engaged in periglacial research. Similar laboratories were established in Yakutia and Canada with primarily civilian missions.

In Canada in the 1950s, Aklavik, the pre-existing administrative centre in the Mackenzie delta region, was suffering from annual breakup floodings. A decision was taken by the Federal Government to construct a new town to replace it which would incorporate the 'best practice' in permafrost construction (Johnston 1981). A number of sites were assessed in terms of their periglacial geomorphological attributes with that at Inuvik selected for development. Inuvik has since become the show piece of how a small town offering the facilities of the non-periglacial world can be created without significant environmental damage. There all the buildings are well insulated and usually placed on piles which penetrate pads of non frost-susceptible materials carefully placed on the original vegetation. A 1 m high air gap through the tops of the piles enables the maintenance of the natural ground thermal regime. Using the same approach, a system of water, sewage and heating pipes were installed in a duct network (utilidor). In some instances, such as power generating units, piles were not feasible and thick pads of granular materials, through which ventilation pipes were inserted, have succeeded in achieving the same objectives.

A vastly improved appreciation of periglacial LANDSCAPE SENSITIVITY has largely ensured that land use activities can be undertaken without major disastrous consequences. Even so construction has to be closely monitored by environmental managers versed in the basics of periglacial geomorphology and in the field of hydrocarbon exploration a number of drill sites have been closed in the summer for fear of excessive disturbance to the ground ice within the permafrost.

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SEE ALSO: active layer; alas; cryostatic pressure; frost and frost weathering; frost heave; hummock; ice wedge and related structures; icing; loess; needle-ice; nivation; niveo-aeolian activity; oriented lake; palsa; patterned ground; pingo; protalus rampart; rock glacier; thermokarst

PETER WORSLEY



# PERMAFROST

Permafrost is defined as ground (soil or rock) that remains below 0 °C for at least two years, and the term is defined purely in terms of temperature rather than the presence of frozen water (Permafrost Subcommittee 1988). Permafrost may, therefore, not contain ice, or may contain both ice and unfrozen water. In many cases, however, ground ice forms a significant component of permafrost, particularly where the substrate comprises fine-grained unconsolidated sediments. The geothermal gradient below the ground surface averages around 30 °C $km^{-1}$ (Williams and Smith 1989) and this increase in temperature with depth determines the thickness of the permafrost (Figure 120). Seasonal temperature fluctuations lead to above zero ground surface temperatures in summer, and the downward penetration of a thawing front. The surface layer that freezes and thaws seasonally is called the active layer, and its thickness depends on the ground thermal properties and on the ratio of the summer thawing index (the accumulated degree-days above freezing) to the winter freezing index (accumulated degree days below freezing). The annual cycle of winter cold and summer warmth is propagated downwards into the permafrost, but rapidly attenuated, so that it becomes undetectable below around 15 m (Figure 120). This is termed the depth of zero amplitude (Brown and Péwé 1973). Longer term changes in ground surface temperatures cause downward propagation of a thermal perturbation, and in many permafrost sites today the geothermal gradient is non-linear, with warm-side deviation that increases towards the surface (Lachenbruch and Marshall 1986), indicating warming over the past century or more (Figure 120).

In northern Canada, permafrost is up to 600 m thick (Figure 121) and its thickness decreases southwards as the climate becomes warmer. Eventually, local variation in ground conditions leads to breaks in the continuity of permafrost, and a complex pattern of discontinuous permafrost results. Under still warmer climatic conditions

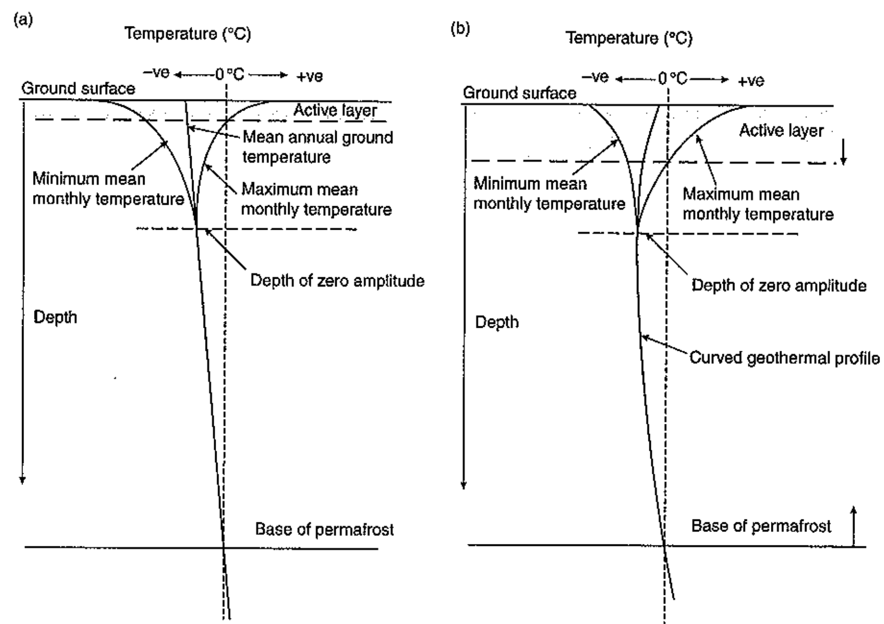


Figure 120 Thermal profiles in permafrost: (a) equilibrium and (b) during thermal adjustment to surface warming

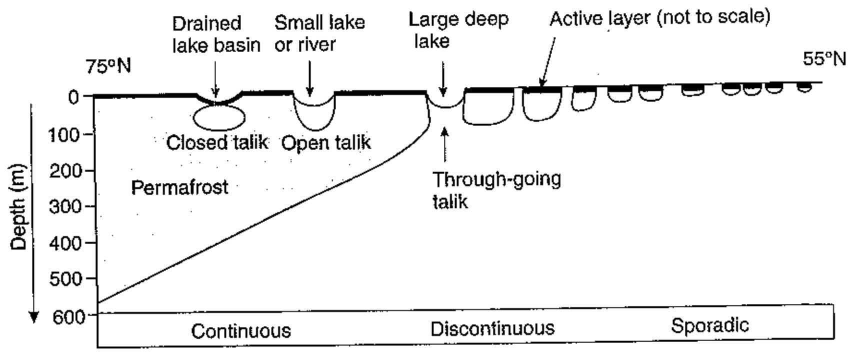


**Figure 121** Typical permafrost characteristics along north–south transect, north-west Canada (after Lewkowicz 1989)

permafrost may form only isolated patches (often related to areas with peat cover), and is described as sporadic. Siberian permafrost is generally colder than North American and in places is in excess of 1,000 m thick (Williams and Smith 1989). However, it takes many millennia for thick permafrost to adjust to surface warming, and it is likely that Siberian permafrost remains chilled by the severity of the last Quaternary cold stage, and is not in thermal equilibrium with present-day conditions.

*Taliks* are unfrozen zones within permafrost terrain that generally occur beneath large bodies of water such as lakes or rivers that do not freeze to their beds in winter. The unfrozen lake or river water is warmer than 0 °C and therefore constitutes a heat source causing a thaw bulb to develop in the underlying permafrost. Drainage of lakes causes downward advance of permafrost, creating a closed talik, entirely surrounded by permafrost (Figure 121). Hydrochemical taliks may be cryotic (below 0 °C), but remain unfrozen due to the flow of mineralized ground water, while hydrothermal taliks may remain non-cryotic due to heat supplied by groundwater flow.

Mean annual permafrost ground surface temperatures are usually higher than the corresponding mean air temperature by a few degrees, so that defining permafrost distribution on the basis of air temperature can be misleading. However, Brown *et al.* (1981) used a mean annual air temperature of approximately −8 °C to delimit the boundary between continuous and discontinuous permafrost in North America, and −1 °C to define the southern limit of discontinuous permafrost. Williams and Smith (1989) stress the multitude of factors that influence the development and survival of permafrost, and point to a gradual southward transition from continuous to discontinuous to sporadic permafrost, with local factors leading to wide variations in associated mean air temperatures.

Where permafrost is developed in unconsolidated sediments, it commonly contains ground ice. Mackay (1972) has provided a classification of ground ice, identifying four categories: pore ice, segregation ice, vein ice and intrusive ice. Both pore ice and segregation ice occur in seasonally frozen soils, but vein ice and intrusive ice occur only in permafrost. Pore ice refers to the ice occupying the pore space in ice-cemented permafrost, and is particularly important in sands and gravels. In fine-grained soils (silts and clays) and porous rocks, much of the pore water occurs as thin films within which capillary and adsorption effects lower the freezing point by several degree celsius (Burt and Williams 1976; Williams and Smith 1989). Progressive freezing of such water results in development of cryosuction, causing water to migrate towards the freezing front. Here it freezes to form lenses of clear ice (segregation ice), increasing ice contents to well in excess of the natural saturated moisture content. Ice segregation during freezing of fine soils causes a significant increase in soil volume and upward frost heaving of the ground surface. Vein ice is the ice that accumulates within permafrost as ice



wedges as a result of thermal contraction cracking (see ICE WEDGE AND RELATED STRUCTURES).

Finally, intrusive ice may form layers up to several metres in thickness as a result of pressurized water flow towards the freezing zone. The pressurized water may be derived from groundwater flow beneath permafrost (open system), or arise from porewater expulsion ahead of a penetrating freezing front in saturated coarse sands and gravels (closed system). Expulsion of water results from the expansion that occurs as pore water changes phase from water to ice. Freezing of pressurized water close to the ground surface in both open and closed systems is responsible for the formation of distinctive conical hills, or PINGOS, the pingo ice being a common form of intrusive massive ground ice (Mackay 1998). Not all massive ice bodies within continuous permafrost originated as intrusive ice, however. In Siberia and parts of northern Canada, ice bodies are considered by some to represent buried glacier ice (Astakhov *et al.* 1996; French and Harry 1990).

The presence of ice-rich permafrost results in high terrain sensitivity to surface thermal disturbance. Permafrost degradation caused by climate warming leads to slope instability and differential settlement as ground ice thaws (French and Egginton 1973). The resulting irregular surface relief is termed THERMOKARST. Widespread thaw settlement in the Arctic has been predicted due to twenty-first-century global warming (Nelson *et al.* 2001). In high altitude mountains, such as the Rockies, Himalayas and the European Alps, discontinuous permafrost is commonly present. Mountain permafrost distribution is generally complex, reflecting altitude, aspect and ground cover, particularly snow cover in winter. Terrain sensitivity to atmospheric warming is again high, and the presence of steep mountainsides increases potential hazards from landslides, debris flows and rockfall (Harris *et al.* 2001).

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CHARLES HARRIS

# PHYSICAL INTEGRITY OF RIVERS

Human activities have dramatically altered the forms and processes of the Earth's river systems. In the northern third of the globe, almost 80 per cent of the rivers are segmented by dams (Dynesius and Nilsson 1994), while in technologically advanced countries such as the United States more than 98 per cent of rivers are significantly impacted by human activities (Echeverria *et al.* 1989). The recognition that the far-reaching effects of channelization, levee building and dam construction have affected biodiversity of aquatic and riparian environments has produced governmental policies

**Plate 88** US Geological Survey LANDSAT image of the middle Missouri River in north-central United States. North is at the top of the image, which extends about 175 km east–west and about 88 km north–south. The Missouri River, largely lacking physical integrity, flows from the left (west) side of the image to the right (east) side. The dark, wide areas are reservoirs behind large dams, while the remaining connections between the dams and the next reservoir downstream are channels where fluvial processes are controlled by the dams. The Niobrara River enters the view from the lower left (south-west) corner of the image. It retains some of its physical integrity and does not include large dams

in many countries to restore rivers and their environments to more natural conditions. Although it is rarely possible to restore rivers to primeval anatural conditions, many nations have policies to promote the physical integrity of rivers, a term that often appears in legislation. Thus, the term integrity has its origins in legal language and usage.

From a scientific and engineering perspective, a physical integrity for rivers refers to a set of active fluvial processes and landforms wherein the channel, floodplains, sediments and overall spatial configuration maintain a dynamic equilibrium, with adjustments not exceeding limits of change defined by societal values (Plate 88). Rivers possess physical integrity when their processes and forms maintain active connections with each other in the present hydrologic regime (Graf 2001). Each term in this definition has particular meaning for the geomorphologist:

- streams and rivers: those parts of the landscape with confined water flow;
- fluvial processes and forms: those features related only to the fluvial domain;
- channel, near-channel landforms, sediments: channel area that is active in the present regime of the river (having a return interval of interaction with flow of 100 years or less), near-channel landforms include the functional surfaces that interact with fluvial processes and channels in the present regime, sediments that are active in the present regime;
- configuration: planimetric and cross-sectional arrangement of functional surfaces, landforms and sediments;
- dynamic equilibrium: the tendency for parameters describing the river to change annually about mean values which also change over periods of decades or centuries;
- limits of change defined by societal values: dimensional and spatial changes in forms and processes within ranges that are acceptable for economic, social or cultural reasons; changes greater than limits imposed by society result in re-engineering the channel to protect lives and property;
- present hydrologic regime: decade or century-long behaviour of daily stream-flow values for magnitude, frequency, duration, seasonality and rates of change.

Measurement of physical integrity for rivers depends on use of a few easily defined, readily assessed indicator parameters. These parameters must have strong roots in the geomorphological literature so that researchers may take advantage of existing knowledge and theory, but the parameters must also be understandable by decision-makers and the public. The parameters must be few in number, and be readily available or easily measured by non-specialists, because river management entails contributions from a variety of observers. Although the range of choices for such parameters in the literature is large (Leopold 1994), the following are the most commonly used and are measured at cross sections: daily water discharge, active channel width, sinuosity, pattern and particle size of bed material.

Of these indicator parameters, daily water discharge is the most important. These data, often collected and made widely available by governmental agencies, provide insight in to the primary driving forces and masses that control the river system. Human interactions with rivers often directly impact the water discharge, with subsequent effects rippling through the geomorphic system. Active channel width is the most easily assessed morphological variable for rivers, and it is the variable that is most sensitive to changes imposed by human activities through discharge adjustments. Classic hydraulic geometry usually shows that width adjusts more than depth or velocity with changes in discharge (Knighton 1998).



SINUOSITY and channel pattern are easily measured on aerial photography for small or medium-sized rivers, or on satellite imagery for large rivers. These parameters reflect upstream impacts of human activities that alter the delicate balance among water, sediment and channel form. Bed material size is readily assessed in field measurements and is sensitive to sediment supply as well as transport capacity (total amount of sediment the river is capable of transporting) and competence (the maximum size of particle that can be transported). Sediment discharge data are also informative, but such data are often not available because they are expensive to measure.

Physical integrity for rivers is important because it underpins biological integrity of river environments. Biological integrity, often characterized by biodiversity and sustainability of ecosystems, depends on the physical substrate of water, sediment and landforms. Efforts to restore rivers to more natural conditions are often thought of in biological terms by planners and managers, but restoration of the underlying physical system must occur first before the biological components of the system can assume more natural conditions.

The most significant human activity in reducing the physical integrity of rivers is the installation of dams. Dams segment the original river system and at least partially control the flow of water and sediment in downstream reaches. Dams reduce peak flows for flood protection, but high flows are important in activating functional surfaces near the channel. As a result, the floodplains downstream from many dams become inactive, causing substantial ecosystem changes that include wholesale vegetation changes. Birds and animals dependent upon the pre-dam vegetation face a loss of habitat under post-dam conditions when the physical and biological integrity of the river decline. In dryland settings, dams often divert the entire flow of the river, resulting in the dessication of reaches that once supported riparian habitat for diverse flora and fauna. In some urban areas, the original stream is replaced by a totally artificial channel without floodplains or other active features, and engineering works often seek to replace braided channels with single thread channels. The restoration of rivers with physical integrity in such cases is a scientific and engineering challenge (Brookes and Shields 1996; Petts and Carlow 1996) that must balance competing objectives subject to social valuation.

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SEE ALSO: floodplain

WILLIAM L. GRAF

# PHYSIOGRAPHY

A word that has obscure origins, although it was in common currency in eighteenth-century Scandinavia, and in regular usage in the English-speaking world in the nineteenth century (see Stoddart 1975, for a historical analysis). Dana defined it in 1863:

> Physiography, which begins where geology ends – that is, with the adult or finished earth – and treats (1) of the earth's final surface arrangements (as to its features, climates, magnetism, life, etc.) and (2) its systems of physical movements and changes (as atmospheric and oceanic currents, and other secular variations in heat, moisture, magnetism, etc.).

One of the most notable exponents of physiography was the British naturalist T.H. Huxley, who published a highly successful text, *Physiography*, in 1877. Huxley's *Physiography* has some geomorphological content including chapters on 'the work of rain and rivers', 'ice and its work', 'the sea and its work', 'slow movements of the land' and 'the formation of land by Animal Agencies'. In the USA W.M. Davis preferred the term to GEOMORPHOLOGY, but he used it without the catholicity of meaning that it had for Huxley. Various other American geomorphologists, including J.W. Powell and N. Fenneman, divided up the USA into what they termed Physiographic Regions, Provinces or Divisions (see Atwood 1940).

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A.S. GOUDIE

# PIEZOMETRIC

A term used in the study of groundwater hydrology referring to the underground water pressure, though the term has particular relevance to underground aquifers and stability analysis of slopes (and PORE-WATER PRESSURE). Underground flow is largely unseen and so instruments called piezometers are used to provide an indication of pressure potential in the water table at particular depths. There are several types of piezometers, including well piezometers, standpoint piezometers and hydraulic piezometers, the commonest being a standpoint piezometer. A typical standpoint piezometer consists of a generally imperforated tube (typically 1–3 m length) inserted into the layer or horizon of interest, with a porous screen at one end to permit flow (about 25–50 mm diameter). Clean sand is placed around the screen, and the borehole surrounding the piezometer tube is filled with a seal (e.g. cement) to ensure that the pressure value given reflects only that on the screen tip. More comprehensive descriptions of piezometer instrumentation is provided in Goudie (1994: 237).

The piezometer gives the potential pressure in a soil or rock by measuring piezometric head, referring to the energy possessed by the water (also termed potentiometric head, hydraulic head and pressure head). Thus, the piezometric head at a point on the water table is the level above an appropriate datum (for instance sea level) that the water table reaches. Differences in head between points in the same water table will result in the transportation of energy (and underground flow) from the point of high piezometric head to that of lesser piezometric head. The velocity of flow between the two points should be directly proportional to the difference in head between, as long as all else remains equal (known as the piezometric gradient). The piezometric gradient can be influenced by external factors, such as precipitation (high amounts of precipitation resulting in a greater piezometric gradient).

By interpolating measurements of piezometric head, an imaginary surface, termed a piezometric surface (though the term potentiometric surface is preferable) can be formed. This represents the distribution of potential energy within the water body. A piezometric surface that lies above ground level will result in flowing water on land. This is termed an artesian well, and is typical of synclinal structures. Insufficient piezometric pressure to reach above ground level is termed subartesian. In unconfined aquifers, the slope of the piezometric surface defines the hydraulic gradient.

Maps of the piezometric surface can be developed, joining together points of equal piezometric head (using contours known as equipotential lines). Flow takes place perpendicular to these lines (down the piezometric gradient), and largely parallel to the overlying surface (Jones 1997: 93). Contour maps provide indications of the piezometric gradient and the pattern of the subsurface flow, and can also be used in stability analysis of soils and rocks.

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STEVE WARD

# PINGO

A pingo is a perennial PERMAFROST mound or hill formed through the growth of a body of ice in the subsurface. The term 'pingo' has been taken from a local Inuktitut word (meaning conical hill) used in the Mackenzie Delta region of the western Canadian Arctic. The term is now used globally to refer to this particular type of ice-cored mound, although in Siberia the Yakutian term *bulgannyakh* is sometimes used.

There are estimated to be some 5,000 or more pingos worldwide with the highest concentration occurring in the Tuktoyaktuk Peninsula area of the western Canadian Arctic, where there are around 1,350 pingos (Mackay 1998). Other areas with considerable numbers of



pingos include the Yukon Territory (Canada), the Canadian Arctic Islands, northern Québec (Canada), northern Alaska (USA), northern areas of the former Soviet Union, the Svalbard archipelago (Norway) and Greenland. A few examples have been noted from Mongolia and the Tibetan Plateau.

Pingos can attain heights in excess of 50 m, with basal diameters of over 600 m. The ice volume of large pingos can exceed 1 million $m^3$. Whilst pingos can form almost conical hills, they may have more irregular forms and can be oval or elongate rather than circular in plan.

In order for a pingo to form and grow, water under pressure must be delivered to a position beneath the surface within a continuous or discontinuous permafrost environment. This water is frozen to form the ice core which is often described as intrusive or injection ice. As the pingo ice core grows the material overlying it (the pingo 'skin' or overburden) is forced upward forming the mound or hill.

Pingos can be classified as either hydraulic (previously termed 'open system' or 'east Greenland' type) or hydrostatic ('closed system' or 'Mackenzie Delta' type). This classification is based on the origin of the groundwater feeding the growth of the pingo ice core.

Hydraulic pingos are initiated by water under pressure of a hydraulic head/potential coming towards the surface in a valley bottom or lower valley side position and they are, therefore, features of high relief environments (for example, east Greenland, Alaska, Svalbard). The position of the upwelling of the ground water may shift over time and in this way a group or complex of pingos may develop within a relatively small area. The largest documented group is the 'Zurich Pingo group' in the Karup Valley of Traill Island, east Greenland (Worsley and Gurney 1996) which has some eleven pingos in various states of growth and decay.

Hydrostatic pingos are initiated by the drainage of a deep lake in a continuous permafrost environment. Following lake drainage the unfrozen saturated sediments that were beneath the lake are aggraded by permafrost and the pore water is progressively squeezed out. It is this water which feeds the growth of the ice core. In general the size of the pingo will be governed by the size of the lake basin from which it grows and usually one pingo will grow in the centre of the former lake basin. If the lake which drains has an irregular form and has two or more basins then one pingo may form in each of the basins.

Studies of the internal structure of hydrostatic pingos in the Tuktoyaktuk Peninsula have shown that beneath the pingo ice core there may be found a pressurized sub-pingo water lens and it is this water which feeds further growth of the pingo. Once such a pingo has become established there is little increase in its diameter and all subsequent growth is upwards, increasing its height. Similar details have not been proven for hydraulic pingos.

One of the longest surveys of a growing pingo was conducted on Ibyuk Pingo in the Tuktoyaktuk Peninsula (Mackay 1998). Although this large pingo was already some 47 m high at the initiation of the survey, the pingo was still seen to grow higher at an average rate of 2.7 cm per year during the survey period 1973–1994. Using this growth rate, along with other data concerning the geomorphological evolution of the area, suggests that Ibyuk may be of the order of 1,000 or more years old (Mackay 1998).

Growth of the pingo ice core leads to the progressive stretching of the overburden causing it to fail through cracking (the generation of dilation cracks) and slumping. The sediment cover at the summit will be thinnest due to this stretching and slumping and pronounced radial cracks may form here. The thinning and rupture of the thermally protective overburden will lead to the decay of the ice core and this will often result in the development of a crater at or near the summit which may contain a pond in summer. When pingos ultimately collapse, whether in a permafrost environment or due to climate change which sees the decay of the permafrost, they do so from the top down and invariably leave a circular or oval rampart surrounding a depression which may contain a pond or marshy area.

Since pingos only form in a permafrost environment, evidence of their previous existence can be used to infer the former presence of permafrost and hence they are extremely useful for palaeoclimatic reconstruction. The remains of pingos of Pleistocene age are often referred to as 'relict pingos' and such features have been documented from North America and western Europe. That these features have always been correctly interpreted, however, is still a matter of some dispute.

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SEE ALSO: ice wedge and related structures; palsa

STEPHEN D. GURNEY

# PINNING POINT

Pinning points are topographic constrictions at which glaciers halt during advances or retreats. They are places where troughs shallow and/or narrow, bifurcate, join another valley or bend sharply. They operate at a range of scales; the fluctuations of CALVING GLACIERs in FJORD systems and ice-contact lakes are sensitive to pinning points, and the topography of land masses and continental shelves affects the behaviour of the floating extensions of ICE SHEETs (ice streams and ice shelves). At non-calving GLACIERs ice is lost primarily through melting so that ice losses are closely tied to climate change. At calving glaciers, however, calving may represent a significant proportion of total ablation, yet calving is only indirectly affected by climate. Calving rates increase with water depth and with the cross-sectional area of the calving terminus, so the mass of ice lost through calving is determined primarily by these non-climatic factors. This means that the fluctuations of calving glaciers can be largely controlled by the topographic geometry of the valley. At pinning points calving rates are reduced, and they therefore represent places of enhanced stability where ice losses are balanced by ice supply. If a glacier retreats from a pinning point into deep water it must continue to retreat until calving rates decrease to match ice supply at a pinning point upstream. Equally, a calving glacier may be unable to advance during periods of regional glacier growth if such an advance would take its terminus into deep, open water. Glacier stillstands are common at pinning points and large MORAINEs may be constructed at these locations. Because these halts are determined by topography and not by climate, such moraine systems may have limited palaeoclimatic significance.


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SEE ALSO: mass balance of glaciers

CHARLES WARREN

# PIPE AND PIPING

Natural soil pipes are linear voids formed by flowing water in soils or unconsolidated deposits (Plate 89). They occur throughout the world and vary from a few millimetres to several metres in diameter (Figure 122). Attempts have been made to provide a quantitative distinction between pipes and other soil macropores based on size, but none has been entirely satisfactory. The most fundamental property of soil pipes is that they actively drain water through the soil, which means that 'connectivity' and a drainage outlet are generally more critical than size in defining a pipe.

As pipes develop, they tend to create subsurface drainage networks akin to surface streams. Horizontal networks up to 750 metres long with

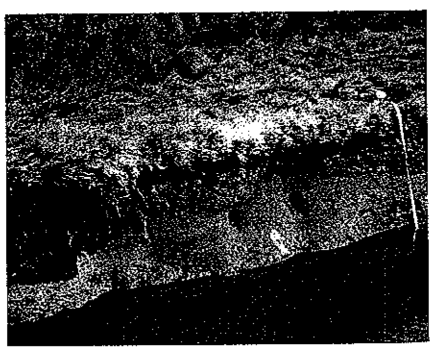

Plate 89 Pipe outlets in a riverbank in the English Peak District

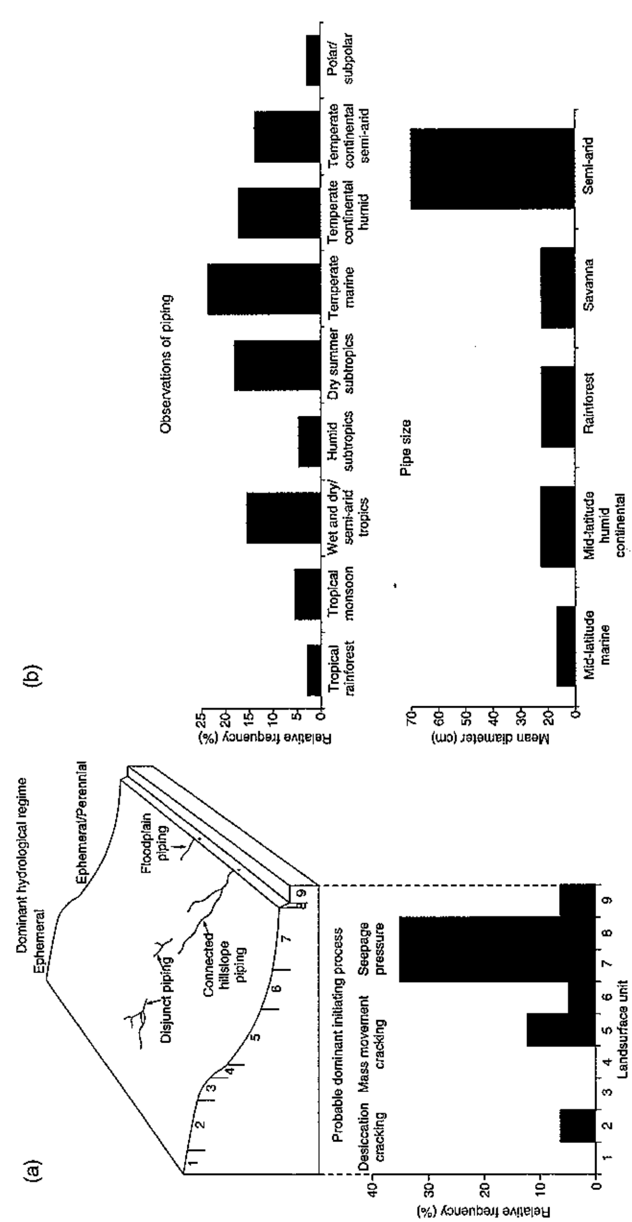


Figure 122 Geomorphic and climatic distribution of pipes: (a) frequency of piping in different landsurface units (Conacher and Dalrymple's (1977) NULM classification); (b) frequency and mean size of pipes in different climatic zones

many branching tributaries, reminiscent of dendritic stream networks, have been reported in shallow upland soils in Britain. In contrast, in badlands or deep loess deposits the networks tend to be more three-dimensional, as in the Loess Plateau in northern China or badlands in Alberta, Arizona, South Africa and Spain. Sometimes these consist of a number of horizontal networks formed above impeding layers with lower permeability and greater resistance to erosion linked by pipes eroding through the layers.

Pipes may be both a cause and a product of GULLYing. Gullies or channels may be formed when pipe roofs collapse. Pipe collapse provides an alternative to the 'classical' theory of channel extension based on headcutting by OVERLAND FLOW entering open channels (Jones 1987a). Conversely, gullies may trigger pipe formation by increasing the surrounding hydraulic gradient. American engineers have been concerned about pipes causing riverbank collapse (Hagerty 1991a,b). In such cases, traditional tests of bank stability based on the properties of the bank material may not give a good indication of risk, as the triggering mechanism may owe more to the source of pipeflow many tens of metres away.

The relationship between piping and LANDSLIDES is equally ambivalent. Pipes may initiate slides if they become blocked or water pressures exceed a critical threshold. Conversely, many peat bogs in Ireland that have well-developed piping do not display the periodic 'bog burst' phenomenon found in those lacking pipes. The pipes may prevent the buildup of water pressure. Again, pipes can develop in mass movement cracks following landslides.

These relationships indicate that shallow subsurface erosion processes can be significant agents in landscape development, sometimes with cyclical collapse and renewal. The term 'pseudo-karst' has been used for such landscapes, because the KARST-like features are predominantly formed by mechanical erosion rather than solution. The term SUFFOSION aptly evokes the mechanical winnowing and scouring. Clearly, the role of subsurface erosion is still undervalued in geomorphology, particularly in modelling.

The mechanics of pipe initiation were first defined by Karl Terzaghi in the founding years of the science of soil mechanics, because piping was the cause of many failures of earth dams. The process begins when seeping water produces sufficient force to entrain material at the seepage outlet point, which may be on a hillslope, a cliff face, riverbank or dam toe slope. The essential feature of Terzaghi's mechanism is that the *water pressure* renders the soil particles or aggregates weightless by counterbalancing gravity, so that the soil seems to 'boil' away (Terzaghi and Peck 1966). Seepage is increasingly focused on the head of the hole as it eats back into the bank. This may be called 'true' piping in the engineering sense. It is quite possible, however, for piping to begin at the upslope end, even on earth dams where it has been called 'rainfall erosion tunnels'. This second process involves progressive expansion of an existing conduit mainly through the shear stress exerted by the flowing water. Cracks caused by desiccation are commonly exploited by rainwater, which preferentially selects and erodes those cracks that run downslope and keeps them open underground even after re-expansion closes the cracks at the surface. Mass movement cracks, tree roots and animal holes can also be exploited. This process has often been distinguished as 'tunnel erosion'. However, the exact mode of initiation may not always be apparent from looking at the resulting landforms in the field. Indeed, in many cases the two processes interact in a very complex manner, as even those who have advocated distinguishing between them admit (Dunne 1990).

Reports of pipes and tunnels from almost every climatic region of the world clearly demonstrate that there is no single, unique set of initiating factors. One of the few truly universal contributory factors is a sufficient water supply to create the necessary pressure or shear stress. This water supply may be very spasmodic and some of the larger pipes occur in drylands, where short, intense rainstorms or the occasional rapid snowmelt event are the main cause.

Steep hydraulic gradients are also commonly needed, typically generated by a combination of water surplus and local relief. Piping is therefore most common where there is a high local relative relief, be it on upland hillslopes, in deeply incised badlands or simply in a riverbank.

Impeding layers within or beneath the soil, which concentrate vertical seepage and divert flow horizontally downslope, are also amongst the most universal preconditions. These are most effective where the layer conducting the flow is more erodible and/or contains pre-existing macropores, especially desiccation cracks, to speed the flow. There is a clear distinction here between the most common initiating factors in humid lands and those in the drylands. Whilst prior cracking aids pipe development in both climatic realms, most dryland piping is linked with dispersible soils, which are not common in humid lands.

Individual soil particles are more easily washed away than soil aggregates. Soil dispersal is predominantly a chemical process. It generally depends on the three-way balance between the total concentration of dissolved cations in the water flowing through the soil, the percentage of soluble or exchangeable sodium in the cation content of the soil itself and the type of clay minerals. In soils bonded by clays, concentrated salt solutions or weakly soluble salts like gypsum, dispersion may be accomplished by the dilution of the bonding salts by seepage water with low salinity. Alternatively, it may be produced by the chemical exchange of cations between the water and the soil, especially the replacement of divalent cations like calcium with monovalent cations like sodium in the soil. Sodium increases the repulsive forces between mineral particles and is the main dispersing or deflocculating agent. However, deflocculation does not necessarily increase erodibility and pipe development. Deflocculated soils may be stable if the sodium content of the soil is high enough to so thoroughly disperse the soil that permeability is reduced below a critical level, and to so reduce the structural stability of the soil that any voids that do develop quickly collapse and fill in. Experimental studies have identified an area of potential instability, swelling clays and deflocculation where piping is likely to be initiated in the transition zone between stable dispersion (high soil sodium and low salinity water) and stable flocculation (low soil sodium and saline seepage water). The exact boundaries between these zones vary according to the predominant clay mineral species. Montmorillonite is generally the least stable, with the broadest zone of instability. Nevertheless, evidence from earth dam engineers in America has revealed that the situation can be rather more complex, and many cases of piping have been observed in earth dams that fall well within the so-called stable zones (Sherard and Decker 1977).

Clay minerals affect the mechanical properties of the soil as well as dispersibility. Minerals of the smectite group like montmorillonite are noted for their higher rates of expansion and contraction. This increases susceptibility to desiccation cracking and infiltration rates. However, the expansive and dispersive properties of montmorillonites depend on the variety of the mineral. Sodium montmorillonite is more expansive than calcium or aluminium montmorillonite, and montmorillonites with cation substitution in their tetrahedral layer rather than their octahedral layer will not display the usual dispersive properties.

The minimum rate of seepage required to initiate piping depends on many properties of the soil, e.g. a threshold of 0.1 mm $s^{-1}$ in non-dispersive silts against only 0.001 in dispersive clays. Soils that contain impeding horizons, are subjected to periodic desiccation and intense rainstorms, have high susceptibility to cracking, especially clayey or organic soils, and/or contain highly erodible, especially dispersible, layers are the most prone to piping.

Human interference has often increased the incidence of piping. Deforestation and devegetation decrease evapotranspirational losses, increasing water surplus. They expose soils to more desiccation and reduce the stabilizing effect of roots. In these situations, piping generally combines with other processes of accelerated erosion causing land degradation and increasing flood hazard (Jones 1981).

Considerable research has been undertaken into methods of rehabilitating farmland damaged by piping in Australia and New Zealand (Crouch *et al.* 1986). These have successfully included planting trees and grasses with high evapotranspiration rates and deep, stabilizing root systems, as well as adding soil conditioners to improve crumb structure. Even so, agricultural land is still being lost to piping, especially in drylands and through over-irrigation, e.g. in Arizona and Spain. Nearly half the farmland in the San Pedro valley, Arizona, has been lost to piping (Masannat 1980).

Research on the hydrology of piping has shown that pipes can contribute significant amounts of water, especially to upland rivers. Nearly half of floodflow and baseflow in one Welsh headwater tributary is derived from ephemeral or perennially flowing pipes (Jones 1987b). The pipes generally flow when the water table rises to pipe level and their discharge is a variable mix of new rainfall and 'old' water pushed out by the new. Monitoring in Canada, Japan, China, India and Britain generally confirms their role in speeding runoff and increasing floodflows. Most contributions fall between 20 and 50 per cent of streamflow, though amounts vary considerably in both time and space, even within the same basin, and in some cases are insignificant (Jones and Connelly 2002). Comparison with the response characteristics of other hillslope drainage processes suggests that pipeflow falls between diffuse throughflow and saturation overland flow in both timing and volume.

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J. ANTHONY A. JONES

## PLANATION SURFACE

Topographical surfaces which are nearly flat over longer distances are called in geomorphology 'planation surfaces'. Some relief is allowed, especially in the form of isolated residual hills, but otherwise slope gradients should be very low and drainage lines should not be incised. Ideally, planation surfaces should cut across bedrock structures.

Thus, the descriptive definition is a simple one, but the issue of planation surfaces is one of the more controversial in geomorphology. As the name implies, the state of low relief has been achieved in the course of planation of a formerly higher relief by means of various exogenic agents of destruction (Plate 90). There are at least a few points persistently disputed, including the nature of process, or processes, leading to planation, the meaning of planation surfaces in long-term landscape evolution, and the possibility of producing and maintaining flat surfaces without recourse to erosional processes acting over protracted time spans (Figure 123).

Several mechanisms have been proposed to account for the origin of near-level surfaces. Accordingly, specific types of planation surfaces are distinguished. These include PENEPLAINS formed by peneplanation (Davis 1899),

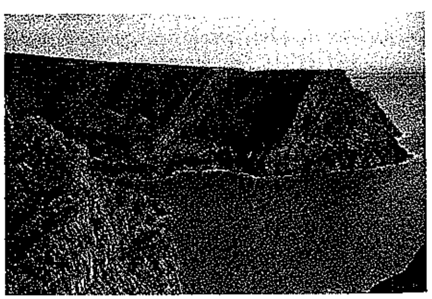

Plate 90 Planation surfaces ideally should truncate geological structures, as it is in the case of a coastal surface in the Gower Peninsula in south Wales. The actual origin of the surface, whether wave-cut or subaerial, is the subject of debate



pediplains formed by pediplanation (King 1953), etchplains formed by etchplanation (see ETCHING, ETCHPLAIN AND ETCHPLANATION) (Büdel 1957). Other, more localized modes of formation are planation by sea waves in the coastal zone, by frost processes in the periglacial environment (see CRYOPLANATION), by areal glacial erosion, or by ubiquitous salt weathering in some desert and coastal situations.

The discussion between protagonists of peneplanation and pediplanation is now to much extent historical. In short, the principal difference between the two models resides in the way of slope development. Peneplains develop primarily through downwearing, i.e. slope lowering. In the course of peneplanation divides are lowered, slopes become gentler with time, and the landscape is progressively graded towards BASE LEVEL. By contrast, in the pediplanation model slope retreat away from drainage lines, i.e. backwearing, plays a crucial part. Higher ground may persist for much longer than the peneplanation model would imply, but their areal extent diminishes in time as bounding escarpments retreat. In front of scarps gently inclined surfaces of PEDIMENTs form and then coalesce to form a regional, ever-growing planation surface of the pediplain type. Another difference is that pediplains are not necessarily graded towards base level and they may form stepped landscapes and develop simultaneously at different altitudes. In both theories, planation surfaces are the ultimate products of long-term landform development and need a long time to form, perhaps of the order of $>10^7$ my. Neither peneplanation nor pediplanation are geomorphic processes per se; rather, they include a variety of superficial processes, including fluvial erosion, surface wash and various categories of mass movement.

Etchplanation was initially seen as a specific variant of peneplanation, applicable to low latitudes, where deep chemical weathering is ubiquitous. However, it was shown later that the mechanism of planation is fundamentally different. Etchplains form in the subsurface, through rock decay which is intense enough to overcome local differences in rock resistance against weathering. This leads to the development of a planar boundary between weathered material and solid rock beneath (see WEATHERING FRONT). Subsequent removal of weathering products exposes the planar 'etched' topography, which now forms an etchplain. Etchplains are thus two-stage features, as opposed to peneplains and pediplains. Although almost featureless etchplains, fulfilling the descriptive criteria for a planation surface, have been described from several areas, the view seems to prevail now that long-term etching leads to diversification rather than to planation of relief. Many low-latitude surfaces of low relief are cut across the weathering mantle, but the hidden topography of the weathering front is much more varied.

As the residual topography rarely gives a clue to the mode of formation of surfaces of low relief, it is preferable to call them simply 'planation surfaces', without genetic connotations. Moreover, it is likely that plains of long geomorphic history have been shaped by various processes, alternating over time, hence they would be 'polygenetic' surfaces rather than any 'monogenetic' peneplains, pediplains, or etchplains (Fairbridge and Finkl 1980). For example, pedimentation may be a means of stripping products of deep weathering to expose an etched surface.

Marine action used to be a favoured mode of planation, and in the nineteenth and early twentieth centuries many flat surfaces, especially in Britain, were identified as 'abrasion surfaces', even if no marine sediments could have been demonstrated. Later studies have shown that wave-trimmed surfaces of regional extent are unlikely to exist, and abrasion platform would have only limited extent (King 1963). Demonstration of a subaerial origin for many surfaces previously claimed as of marine origin, undermined the concept even further.

Periglacial planation is supposed to be achieved by means of simultaneous action of frost weathering of bedrock, rock cliff development and retreat, and mass movement, chiefly solifluction. The resultant cryoplanation is essentially a variant of pediplanation, applied to high latitudes and the Pleistocene. As with marine planation, cryoplanation is now generally seen as unlikely to account for the origin of more extensive surfaces, which are mostly inherited from pre-Pleistocene times.

Extensive level terrains in the Canadian and Fennoscandian Shield were long believed to have been shaped by powerful glacial erosion exerted by consecutive ice sheets during the Quaternary. Later research has shown that areal glacial erosion is not as common as formerly thought (Sugden and John 1976). By contrast, remarkable flatness of basement surfaces is inherited from protracted pre-Quaternary subaerial development, whilst much of these surfaces are exhumed Precambrian features (Lidmar-Bergström 1997).

The recognition of the powerful role of salt weathering in low-lying desert environments has led to the proposal that this process may have been crucial in producing flat topography of some coastal plains or closed depressions such as Quattara in Egypt. The term 'haloplanation' has been suggested (Goudie and Viles 1997).

Planation surfaces occupied a central position in geomorphology in the days, when establishing DENUDATION CHRONOLOGY of a given area was considered the main objective of geomorphology and cyclic development of landforms served as a paradigm. The first step in geomorphic research was to identify planation surfaces in the present-day landscape, or more often their remnants surviving on divides after the landscape had been dissected. It was followed by recognition of

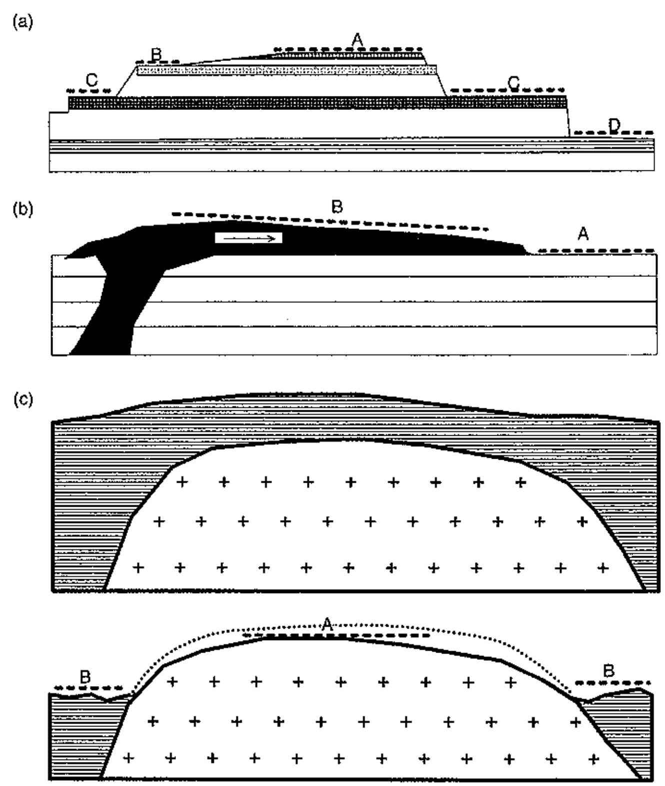


Figure 123 A diagram to show that not all plains are products of protracted planation. In areas built of flat-lying sedimentary rocks benches are distinctively controlled by structure (a). Flatness of surfaces underlain by lava flows may be largely primary and relate to the low viscosity of lava (b). The surface 'B', although located at higher altitude, is actually younger than the true planation surface 'A'. Surfaces of low relief common for many granite batholiths may be related to the original dome form assumed during emplacement (c). Different altitudes of surfaces cut in granite ('A') and country rock ('B') reflect different resistance of each rock complex rather than being indicators of different ages



altitude range of their occurrence, correlation of surfaces over wider areas and classification by height. The number of planation surfaces indicated the number of erosional cycles experienced by the landscape. As this approach frequently ignored influences differential tectonics may have had on landform development, relied on highly uncertain correlation procedures, and suffered from deficiency of accurate dating of surfaces, it began to be severely criticized in the 1960s and lost much of its popularity. Thus, whereas the issue of planation surfaces, their origin and chronology, features prominently in older regional studies, its importance has been greatly reduced in modern geomorphology. In addition, preoccupation with planation surfaces in historical geomorphology created an incorrect view that a search for ancient landscapes is essentially a search for planated relief. Recent work indicates that many palaeosurfaces preserved in geological and geomorphic record had a very complex topography, far from any state of advanced planation.

However, planation surfaces have retained significance in morphotectonic studies and are widely used as indicators of uplift and subsidence histories (Ollier and Pain 2000). By analysing spatial and altitude patterns of their distribution one can infer magnitudes of surface uplift, recognize direction of tilting and amount of warping, or locate fault zones in areas where conventional geological evidence is not at hand. In this context, the origin of a planation surface is usually not critical to the argument.

It is important to remember that the surface morphology alone may not give the clue to the reasons of flatness. Examination of geological structure underlying a flat topography can suggest origins alternative to the one implied by the name 'planation surface'. Moreover, if geological control can be demonstrated, tectonic quiescence is not a prerequisite any longer as inherent in the 'classic' modes of formation.

In many platform areas sedimentary strata lie horizontally over very long distances and topography may be adjusted to this negligible dip, especially if resistant rock layers occur at the surface. Likewise, backslopes of CUESTA ridges may show close adjustment to the dip of strata. In elevated plateaux there may exist several levels of 'planation surfaces' separated by escarpments, but in reality these are structural surfaces, each following a more resistant layer.

In formerly volcanic areas, topography may be adjusted to the geometry of lava flows. Basaltic lava, because of its low viscosity, may extrude in sheets of more or less uniform thickness over large areas. Flatness of the top surface of a flow will be then inherited from the time of extrusion and cooling. In case of multiple flows, denudation may expose top parts of each major lava flow and produce stepped topography, reminiscent of a generation of planation surfaces of various ages. In fact, all benches have structural foundation and may all have similar ages.

Remnants of ancient planation surfaces have also been sought in areas underlain by igneous rocks, especially by granite, which indeed often display a gently rolling topography or resemble laterally extensive, low-radius domes. A variety of structural controls is possible, including adjustment of form to the original roof of the intrusion, or to flat-lying joints.

That planation surfaces exist on Earth is indisputable. Examples are known from throughout the geological record, from Precambrian up to the present. In places such as the Fennoscandian Shield, Laurentide Shield or the Middle East, extensive surfaces of extreme flatness truncating various bedrock structures and disregarding differential rock resistance existed by the end of Precambrian. They were subsequently buried by Cambrian sediments and form sub-Cambrian planation surfaces, nowadays partially exhumed. Another generation of extensive planation surfaces evolved in the Mesozoic. Many upland areas are typified by the occurrence of surfaces of low relief at different altitudes, which probably have formed during the Cenozoic. It is the meaning, the mode of origin and age range of their formation which remain contentious and, paradoxically, underresearched issues.

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PIOTR MIGOŃ

# PLATE TECTONICS

Plate tectonics is a unifying theory that explains many of the major features of the Earth's lithosphere such as VOLCANOes, rift (see RIFT VALLEY AND RIFTING) zones and mountain belts. The theory asserts that the lithosphere – the crust and uppermost mantle – is divided into rigid bodies or 'plates' that move horizontally and interact at their boundaries to produce these features. The theory evolved from the earlier concepts of continental drift and SEAFLOOR SPREADING.

In the eighteenth century it was noted that the coastlines of the Atlantic Ocean fit together like a jigsaw puzzle. In 1915 Alfred Wegener published several geological arguments to hypothesize that the continents had drifted apart from a supercontinent named Pangaea. He pointed to features that could be aligned by closing the Atlantic including Palaeozoic fold belts like the Appalachian and Caledonian mountains (Figure 124), metamorphic shields like northern Scotland and Labrador, major faults, palaeoclimatic indicators such as Carboniferous subtropical coal beds and tropical evaporites and desert sandstones, tillites from a Carboniferous ice cap with radiating striations and glacial erratics, and unique fossils of ferns and reptiles. Shortly thereafter he noted that seismic velocities in oceanic rocks were faster than in continental rocks, indicating that the less dense continents would float on top of oceanic rocks.

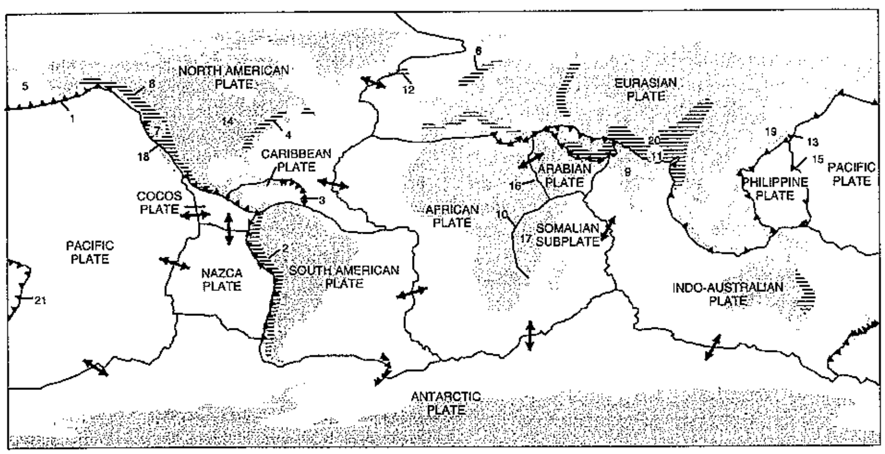


Figure 124 Plate boundaries of the Earth's crust. Features mentioned in the text: 1: Aleutian island arc; 2: Andes mountains; 3: Antilles island arc; 4: Appalachian mountains; 5: Bering Sea; 6: Caledonian mountains; 7: Columbia River basalts; 8: Cordilleran mountains; 9: Deccan basalts; 10: East African rift; 11: Himalayan mountains; 12: Iceland; 13: Japanese island arc; 14: Keweenawan aulacogen; 15: Mariana trench; 16: Mount Kilimanjaro; 17: Red Sea; 18: San Andreas fault; 19: Sea of Japan; 20: Tibetan plateau; 21: Tonga trench



Wegener's arguments were widely discredited for several decades. One exception was Alexander Du Toit who observed in 1937 that Pangaea split into the supercontinents of Laurasia (North America and Eurasia) and Gondwanaland (South America, Africa, India, Australia and Antarctica) in Triassic time, and that they split into the present continents in Jurassic time.

Geologic mapping increased dramatically worldwide in the 1950s, adding abundant evidence in support of continental drift. Critical geophysical arguments were added. For example, the advent of computers and related statistics enabled Bullard *et al.* (1965) to show the high probability of fit along the Atlantic continental margins. K-Ar radiometric age DATING METHODS and aeromagnetic anomaly maps confirmed the match of exposed and buried lithologies across the South Atlantic. Seafloor exploration confirmed that seafloor and continental rocks differed. Palaeomagnetic studies proved most persuasive. By measuring a rock unit's fossilized remnant magnetization in many samples, its location can be calculated relative to the Earth's palaeomagnetic pole with a known error. By comparing poles from coeval rock units on two continents, the relative rotation and displacement between them can be determined. Such studies statistically proved continental drift (Irving 1958).

Arthur Holmes in 1928 realized that radioactive decay heat would cause the Earth's mantle to convect and push continents along the Earth's surface. Dietz (1961) suggested that seafloor spreading occurred when convecting mantle magma intruded into the seafloor crust at mid-oceanic ridges, causing ACCRETION of new seafloor and carrying older seafloor away in both directions. Tests of the hypothesis followed, relying heavily on sonar maps of the submarine geomorphology and aeromagnetic maps of the seafloor that had been produced to track nuclear submarines. Quickly it was shown, on going perpendicularly away from a mid-oceanic ridge, that the youngest rock ages on oceanic islands increases linearly, that the K-Ar radiometric and fission track (see FISSION TRACK ANALYSIS) ages of seafloor basalts increase linearly, that the seafloor cools, and that the microfossils and tuffs in seafloor sediments record increasing ages – all supporting seafloor spreading. Most critical was the recognition that reversals of the Earth's magnetic field every $10^4$ to $10^7$ Ma were recorded in new lavas at the mid-oceanic ridges, which then spread away equally in both directions to create linear rift-parallel anomalies of increased and decreased magnetic field intensity. This analysis both confirmed the seafloor spreading hypothesis and provided a way to easily map and date the seafloor, which represents about two-thirds of the Earth's surface (Heirtzler *et al.* 1968). The plate tectonic theory evolved to explain how the Earth's crust is moving everywhere systematically.

The Earth's surface has seven large plates – the Pacific, Eurasian, North American, South American, African, Indo-Australian and Antarctic plates, five medium-sized plates (Figure 124), and an uncertain number of small plates in between. Plates move at rates up to about 20 cm yr$^{-1}$ although the norm is only 5–10 cm yr$^{-1}$. This motion can be directly measured using very long baseline interferometry, satellite laser ranging, and satellite radiopositioning using the Global Positioning System's satellites. Past rates are measured using the spacing of seafloor magnetic anomalies and palaeomagnetic methods. Further, because crustal plates are fitted around a sphere, each plate actually rotates about its own Euler pole with a varying rate of motion across it. Plate boundaries have three types of ACTIVE MARGINS: rift zones, subduction zones and transform faults. Each may involve oceanic, continental or both types of crust and each combination of boundary and crust has its own typical topographic expression, dynamics and geologic characteristics.

## Rift zones and ridges

Rift zones are linear zones of extension where new crust is added to the Earth's surface and where the motion is perpendicularly away from the lineament. Topographically, they have a central depression, typically 20 to 50 km wide and 1½ to 3 km deep, between uplifted plateaux that extend outwards for hundreds of kilometres on either side. Below the depression, convecting upwelling magma splits at the crust–mantle interface, pushing older crust upward and outward on inward-dipping normal faults to form the plateaux. Decreased lithostatic loading in the depression causes melting of the uppermost mantle to form a mafic magma. It intrudes along the rift lineament, particularly along its central axis, to form sheeted gabbroic intrusions at depth and basaltic lavas on the depression's floor, creating a new crust about 5 km thick. Because the crust is thin and the forces tensional, rift zones generate numerous relatively weak, shallow earthquakes.

On the seafloor, the rift zones are called 'ridges'. They form a great continuous submarine mountain system through the world's oceans that extends for about 60,000 km (Figure 124), but is exposed in only a few places such as Iceland. Typical mid-oceanic ridge basalts (MORB) are olivine tholeiites with only minor compositional variations. Erosion is minimal in the oceans, so that only a thin veneer of mostly chemical and biochemical sediments precipitate on the seafloor volcanics as siliceous and calcareous oozes. These sediments increase slowly in thickness progressively away from the ridge, typically at a rate of about 1 mm/$10^2$ yr. In the ridge depression, high heat flow from the Earth causes abundant geothermal activity. The hot hydrothermal fluids can precipitate elegant chimney structures around their vents that are rich in metal sulphides to form ore deposits. Also the fluids chemically alter the basalts and gabbros as they pass through and they create a marine microenvironment with a diverse range of exotic flora and fauna. Heat flow, seismic and gravity studies show that the seafloor crust progressively cools, increases in density and sinks, and thickens by underplating as magma freezes on its lower surface as it is pushed further away from the rift.

The East African rift zone is about 5,000 km long and exemplifies how the continental crust of a CRATON is split (Figure 124). Such rift zones are topographically and seismically similar to oceanic ridges. Because the bounding plateaux are exposed to WEATHERING, minor chemical sediments occur around hot springs only but EROSION produces abundant terrestrial clastic sediments that are deposited in the rift valley and the volcanic plateaux are progressively attacked by surficial weathering away from the valley. Tholeiite basalts sometimes fill and overflow the rift, producing large plateau lava sequences such as the Columbia River or Deccan basalts. Also, alkali basalt volcanism may occur in the adjacent plateaux through partial melting of continental crust, creating intrusive plugs, extrusive cinder cones and spectacular stratovolcanoes such as Mt. Kilimanjaro with its equatorial GLACIER peak. If the rift zone is truly active, the continental crust is entirely split in 10 to 20 Ma, the intervening crust becomes a mid-oceanic ridge, the sea invades as has happened in the Red Sea, and the continental edges become PASSIVE MARGINS.

## Subduction zones

Subduction zones are convergent margins where the Earth destroys old crust to make room for the new crust created in rift zones. As two plates converge, one plate bends downward and is driven back into the Earth's mantle at an angle to be remelted. A subducted slab can reach depth extents of 1,400 km long and reach a depth of 700 km. The plates may converge head on or at a substantial angle, resulting in strong compressive forces that produce numerous powerful earthquakes with foci along the depth extent. In fact, about 90 per cent of all earthquake energy is released in subduction zones so that their locations are well known from seismology studies. Three types of crustal collisions are possible: ocean-to-continent, ocean-to-ocean, and continent-to-continent.

When oceanic crust collides with continental crust, as is happening around most of the Pacific Ocean (Figure 124), the thin (5–10 km) denser (~2.9 g/cc) oceanic plate dives under the thicker (30–50 km) less dense (~2.7 g/cc) continental plate at a 30° to 70° angle. At the surface a trench is formed that may be several thousand kilometres long in plan with typical widths of 50 to 100 km and depths of 7 to 9 km. Thus the epicentres for shallow earthquakes occur at the trench. Most trenches are arcuate with their concave side facing the continent like the Aleutian trench, but some are straight like the Tonga trench. Any sediment from erosion on the exposed continent that reaches the trench's bottom is subducted down into the Earth, so the trench never fills. However, such sediments do accumulate on the CONTINENTAL SHELF as TURBIDITY CURRENTS carry them into a fore-arc basin between the continental margin and the continent's shoreline. Commonly the turbidites are compressed, deformed and metamorphosed in the plates' collision to form a flysch belt along the coastline. Sometimes seafloor sediments and crust are obducted onto the shelf and preserved in the flysch. Inbound of the flysch, an ISLAND ARC forms. Heat, generated mostly by friction in the subduction zone, melts the overlying sialic and simatic rocks of the upper and lower continental crust and the mafic to ultramafic uppermost mantle, causing large diorite to granite intrusions to be emplaced into the old continental crust and mafic to felsic volcanics to be extruded onto it. Thus the world's most spectacular volcanoes are found mostly around the Pacific Ocean. Where there is a wide continental shelf, this intrusive–extrusive rock complex forms a mountainous island arc with a submarine back-arc basin behind, like the Aleutian arc and Bering Sea or the Japanese arc and Sea of Japan. Alternatively, if the shelf is narrow, the complex forms a range of high coastal mountains. Further compression deforms and stacks the rocks behind the arc into sheets on thrust faults, forming high mountains like the Andes and Cordillera with their photogenic MOUNTAIN GEOMORPHOLOGY.

Collisions between oceanic plates are less common than oceanic–continental collisions but the process is similar except that both crusts are relatively thin and dense. Consequently their subduction zones descend a steeper angle approaching 90° and go deepest into the Earth, producing most of the world's deepest earthquakes. Also their trenches, like the Mariana trench, are deeper and reach depths of about 11 km. Further, only the peaks of the arc volcanoes reach the surface as a chain of basaltic islands, such as the Antilles, on the margin of the non-subducting plate. With so little land above sea level, very little clastic sedimentation occurs, but chemical sediments and CORAL REEFS often form around such islands in tropical climates to form barrier (see BARRIER AND BARRIER ISLAND) reefs and atolls.

Collisions between continental plates create the world's greatest mountains with spectacular TECTONIC GEOMORPHOLOGY, like the 10-km high Himalayan Mountains where the Indian subcontinent rides on the Indo-Australian plate and butts against the Eurasian continent and plate. The mountains are high because both crusts are relatively thick but less dense than the Earth's mantle and so they float to high elevation. Their combined 70-km thick crust sinks deeper into the Earth's mantle because of isostasy. Prior to collision, the continents had oceanic crust between them with either a passive margin like most of the Atlantic coastline or more commonly with an oceanic–continental collisional margin as described above. As the continents collide, some seafloor sediments, volcanics and gabbroic intrusions are often squeezed up and trapped between them where they are complexly deformed, and metamorphosed to serpentinites. Finding serpentinites defines the suture between the plates. On both sides of the suture zone are the deformed and metamorphosed remains of island arcs and back-arc basins first, then very high thrust-faulted mountains of stacked sedimentary sequences, followed by high foothills of folded sedimentary rocks, and then high plateaux such as the Tibetan plateau, with gently deformed strata that are deeply dissected by canyons. The suture zone and thrust-fault belts are tectonically active, producing strong shallow earthquakes mainly. Below the suture, subduction stops on closure. However, it takes tens of millions of years for the relict slab of descending crust to melt so that some minor earthquakes still occur at intermediate depths. Closure also terminates most volcanism because frictional heating ceases in the subduction zone.

## Transform faults and triple junctions

J. T. Wilson's (1965) description of a new type of fault, a transform fault (Figure 125), was the key to understanding the dynamics of plate tectonics. The motions of rift and subduction zones were well understood, but the seafloor scarps were thought to be transcurrent faults where the two sides slid in opposite directions to offset a marker. Noting that these scarps cut the ridges at right angles and offset them, Wilson reasoned that the seafloor was moving away from both offset ridge segments in both directions so that the fault's sides were moving: (a) towards each other between the offset ridges as an active plate boundary, and (b) in the same direction and speed outside of the two ridges where they are tectonically

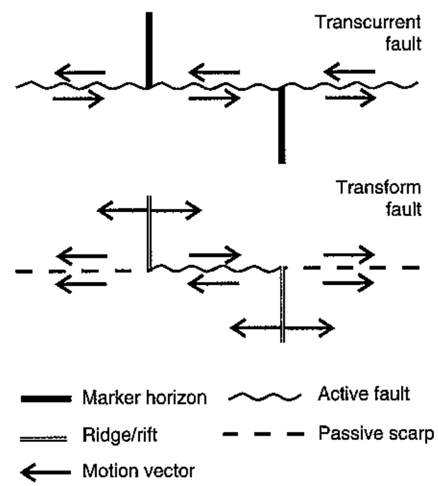


Figure 125 Dynamic comparison of transcurrent and transform faults

passive and inside the plates (Figure 125). He further explained how the transform fault motions interacted with ridge and subduction zone motions to create plate boundaries.

On the seafloor, transform fault lineaments can be over 3,000 km long and 3 km in height, being highest on the side of the closest ridge segment. Excluding very minor amounts of active volcanism, the Earth's crust is neither created nor destroyed. Between the ridges, horizontal shear generates weak to moderate, shallow earthquakes that are rarely destructive. In a few places, such as the San Andreas fault, the active part of a transform fault cuts much thicker continental crust, causing periodic powerful and highly destructive shallow earthquakes as the two sides grind past each other. Minor alkalic volcanics, sag ponds and other geomorphic features may occur along such faults.

Active plate boundaries form triple junctions where three plates meet, separated by three boundaries or arms. Each arm may be a rift, subduction zone or transform fault to make sixteen possible combinations. For example, the Pacific–Cocos–Nazca triple junction has three radiating rifts as arms whereas the Pacific–Cocos–North American triple junction has a rift, a subduction zone and a transform fault as its arms. The geometrically important fact is that the velocity and direction of each plate at a triple junction must form a spherical vector triangle. Thus, if the rotational motions of two plates are known from seafloor magnetic anomalies or palaeomagnetism, the rotational motion and Euler pole of the third plate can be calculated.

## Importance of the theory

Understanding plate tectonics has led to a revolution in the Earth sciences. Virtually all of its supporting evidence relates to the last 200 Ma or so but, invoking UNIFORMITARIANISM, it has provided the basis for geographers to understand how the Earth's landscapes developed and for geologists to interpret the preceding 4,000 Ma of the Earth's rock record. For example, the Caledonian and Appalachian mountain belts are now recognized to be the deformation product of three Palaeozoic collisional events with oceans opening and closing in between in WILSON CYCLES. Similarly, geologists and geophysicists are using the theory to fit Precambrian supercontinents together such as Rhodinia, to identify failed rift valleys such as the 1,100 Ma Keweenawan aulacogen, to explain how geologic terranes that formed in radically different tectonic settings are now abutting, and to explain epeirogenic vertical motions in the continental interiors to form basins and plateaux. Finally, knowing how plates have moved over the past 200 Ma is enabling geophysicists to constrain models to investigate the Earth's magnetic field and to understand convective motions in its interior.

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D.T.A. SYMONS

# PLOUGHING BLOCK AND BOULDER

Ploughing blocks and boulders are individual boulders that move downslope faster than their surrounding material by processes related to seasonal frost. Their movement ranges from millimetres to a few centimetres a year and is restricted to the annual freeze–thaw cycle (Ballantyne 2001; Berthling *et al.* 2001a). Due to the differential movement, the boulder pushes up a mound against its downslope side while leaving a depression along its upslope track. Ploughing boulders belong to the area of PERIGLACIAL GEOMORPHOLOGY. They are developed on slopes in the warmer part of the periglacial belt, commonly together with solifluction lobes. Only a few detailed process studies exist, but these indicate that boulder movements are caused by the same processes that occur in SOLIFLUCTION. During autumn and winter, the boulders protrude above the snow cover for some time. Combined with differences in thermal conductivity, this causes more intensive heat loss through the boulder than through ground elsewhere. This results in favourable conditions for ice segregation (Ballantyne 2001) and causes FROST HEAVE of the boulders. A heave of up to 7.5 cm has been demonstrated from southern Norway (Berthling *et al.* 2001b). It was shown that boulder heave stopped in midwinter regardless of snow conditions. Depletion of soil moisture might explain this behaviour. During spring and early summer, the boulders melt out of the snow and the soil beneath the boulder starts to melt earlier than the surrounding ground. Consolidation rates of up to 4.2 mm/day through a six-day period were measured by Berthling *et al.* (2001b). If melting of ice is more rapid than the ability of the released water to escape, water is trapped beneath the boulder so that the porewater pressure rises. This causes the soil beneath the boulder to lose strength and downslope deformations may occur. The process is referred to as gelifluction, and is the main cause for ploughing boulder movement. A second process that has been invoked is frost creep. Frost creep results from frost heave normal to the freezing plane and settlement along the vertical. This results in a net downslope movement in the cases where the freezing plane is essentially parallel to the sloping ground surface. The frost creep model in its simple form should be abandoned, as the boulders heave and tilt in directions determined by variations in heat removal, frost susceptibility and soil water content beneath the boulder, not slope. Yet, instability caused by tilting during frost heave might induce some displacements during thaw.

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SEE ALSO: frost heave; periglacial geomorphology; solifluction

IVAR BERTHLING



# PLUVIAL LAKE

Bodies of water that accumulate in basins as a result of former greater moisture availability resulting from changes in temperature and/or precipitation. The study of pluvial lakes developed in the second half of the nineteenth century. Jamieson (1863) called attention to the former greater extent of the great saline lakes of Asia: The Caspian, Aral, Balkhash and Lop-Nor and Lartet (1865) pointed to the expansion of the Dead Sea. The term pluvial appears to have first been applied to an expanded lake by Hull (1885), but was originally applied by Tylor (1868) to valley fills in England and France. A major advance in the study of pluvial lakes came in the western USA with the work of Russell (1885) on Lake Lahontan and of Gilbert (1890) on Lake Bonneville. A discussion of these early studies and their bibliographic details is given in Flint (1971: Chapter 2).

The Great Basin of the USA held some eighty pluvial lakes during the Pleistocene, and they occupied an area at least eleven times greater than the area they cover today. Lake Bonneville (Plate 91), was roughly the size of present-day Lake Michigan, about 370 m deep and covered 51,640

Plate 91 A group of high shorelines that developed in the Late Pleistocene around pluvial Lake Bonneville, Utah, USA

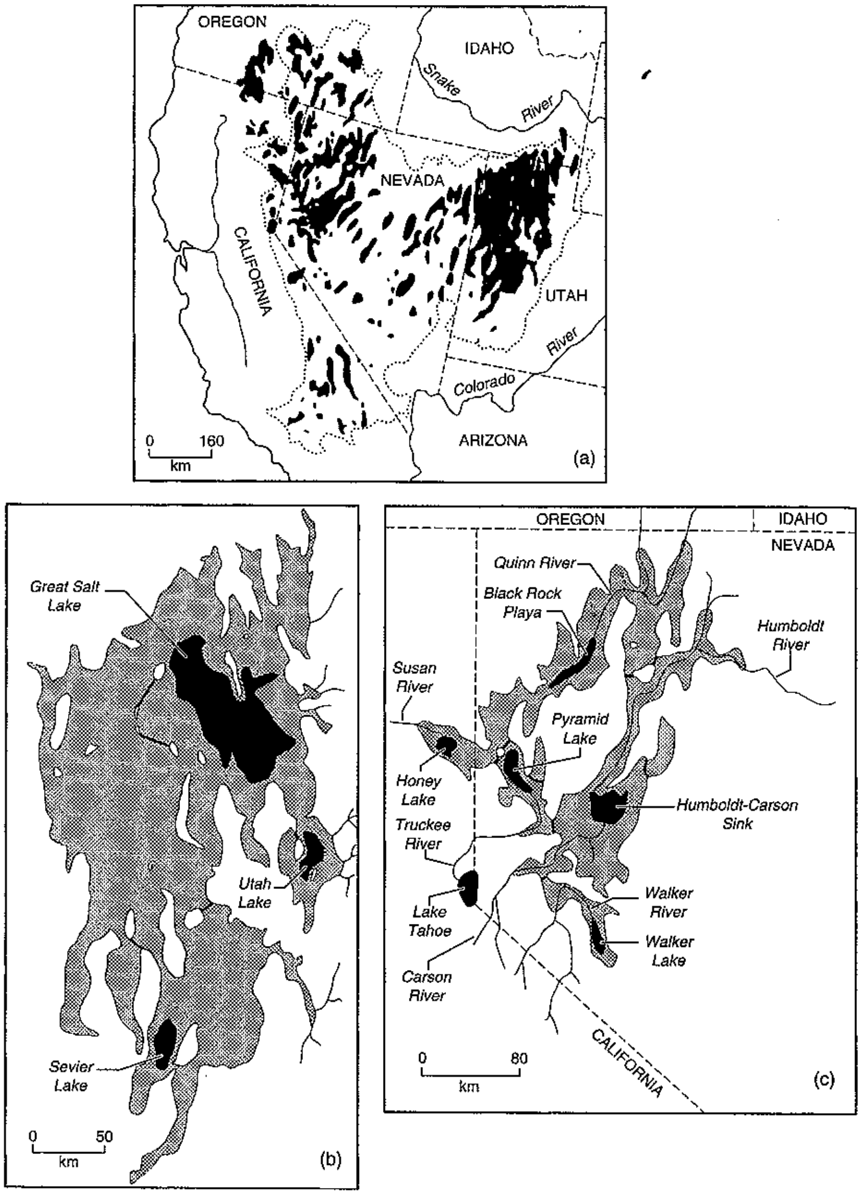


Figure 126 (a) The distribution of Pleistocene pluvial lakes in the southwestern USA; (b) Lake Bonneville; (c) Lake Lahontan



km$^2$. Lake Lahontan was rather more complicated in form, covered 23,000 km$^2$, and reached a depth of about 280 m at the site of today's Pyramid Lake (Figure 126). It covered an area nearly as great as present-day Lake Erie. River courses became integrated and lakes overflowed from one sub-basin to another. For example, the Mojave River drainage, the largest arid fluvial system in the Mojave Desert, fed at least four basins and their lakes in pluvial times: Lake Mojave (including present-day Soda and Silver lakes), the Cronese basin and the Manix basin (which includes the Afton, Troy, Coyote and Harper sub-basins; Tchakerian and Lancaster 2002). Also important was the Owens Lake–Death Valley system.

A very large amount of work has been done to date and correlate the fluctuations in the levels of the pluvial lakes. Much of the early work is reviewed by Smith and Street-Perrott (1983). They demonstrated that many basins had particularly high stands during the period that spanned the Late Glacial Maximum, between about 25,000 and 10,000 years ago. More recently there have been studies of the longer term evolution of some of the basins, facilitated by the study of sediment cores, as for example from Owens Lake, the Bonneville Basin, Mono Lake, Searles Lake and Death Valley.

The high lake levels during the Last Glacial Maximum may well be the result of a combination of factors, including lower temperatures and evaporation rates, and reduced precipitation levels. Pacific storms associated with the southerly branch of the polar jet stream were deflected southwards compared to the situation today.

Other major pluvial lakes occurred in the Atacama and Altiplano of South America (Lavenu *et al.* 1984). The morphological evidence for high lake stands is impressive and this is particularly true with regard to the presence of algal accumulations at high levels (as much as 100 m) above the present saline crusts of depressions like Uyuni (Rouchy *et al.* 1996). There is a great deal of variability and confusion about climatic trends in the Late Quaternary in this region, not least with respect to the situation at the Late Glacial Maximum and in the mid-Holocene (Placzek *et al.* 2001). Nonetheless, various estimates have been made of the degree of precipitation change that the high lake stands imply. Pluvial Laguna Lejíca, which was 15–25 m higher than today at 13.5 to 11.3 Kyr BP and covered an area of 9–11 km$^2$ compared to its present extent of 2 km$^2$, had an annual rainfall of 400–500 mm, whereas today it has only around 200 mm. Pluvial Lake Tauca, which incorporates present Lake Poopo, the Salar de Coipasa and the Salar de Uyuni and which had a high stillstand between 15 and 13.5 Kyr BP, had an annual rainfall of 600 mm compared with 200–400 mm today.

In the Sahara there are huge numbers of pluvial lakes both in the Chotts of the north, in the middle (Petit-Maire *et al.* 1999) and in the south (e.g. Mega-Chad). In the Western Desert there are many closed depressions or playas, relict river systems and abundant evidence of prehistoric human activity (Hoelzmann *et al.* 2001). Playa sediments contained within basins such as Nabta Playa indicate that they once contained substantial bodies of water, which attracted Neolithic settlers. Many of these sediments have now been subjected to radiocarbon dating and they indicate the ubiquity of an early to mid-Holocene pluvial phase, which has often been termed the Neolithic pluvial. A large lake formed in the far north-west of Sudan, and this has been called 'The West Nubian Palaeolake' (Hoelzmann *et al.* 2001). It was especially extensive between 9,500 and 4,000 years BP, and may have covered as much as 7,000 km$^2$. If it was indeed that big, then a large amount of precipitation would have been needed to maintain it – possibly as much as 900 mm compared to the less than 15 mm it receives today.

In the Kalahari of southern Africa, Lake Palaeo-Makgadikgadi encompassed a substantial part of the Okavango Delta, parts of the Chobe–Zambezi confluence, the Caprivi Strip, and the Ngami, Mababe and Makgadikgadi basins. At its greatest extent it was over 50 m deep and covered 120,000 km$^2$. This is vastly greater than the present area of Lake Victoria (68,800 km$^2$) and makes Palaeo-Makgadikgadi second in size in Africa to Lake Chad at its Quaternary maximum. Dating it, however, is problematic (Thomas and Shaw 1991) as is its source of water. Some of the water may have been derived when the now DRY VALLEYS of the Kalahari (the *mekgacha*) were active and much could have been derived from the Angolan Highlands via the Okavango. However, tectonic changes may also have played a role and led to major inputs from the Zambezi.

In the Middle East expanded lakes occurred in the currently arid Rub-Al-Khali and also in Anatolia (Roberts 1983). In Central Asia the

Aral–Caspian system was hugely expanded. At several times during the late Pleistocene (Late Valdai) the level of the lake rose to around 0 m (present global sea level) compared to −27 m today and it inundated a huge area, particularly to its north. In the early Valdai glaciation it was even more extensive, rising to about +50 m above sea level, linking up to the Aral, extending some 1,300 km up the Volga River from its present mouth and covering an area in excess of 1.1 million km$^2$ (compared to 400,000 km$^2$ today). At its highest it may have overflowed through the Manych depression into the Black Sea. In general, transgressions have been associated with warming and large-scale influxes of meltwater (Mamedov 1997), but they are also a feature of the glacial phases when there was also a decrease in evaporation and a blocking of ground water by permafrost. Regressions occurred during interglacials and so, for example, in the Early Holocene the level of the Caspian dropped to −50 to −60 m below sea level.

Large pluvial lakes also occur in the drylands and highlands of China and Tibet and levels appear to have been high from 40,000 to 25,000 BP (Li and Zhu 2001). Similarly the interior basins of Australia, including Lake Eyre, have shown major expansion and contractions, with a tendency for high stands in interglacials (Harrison and Dodson 1993).

As can be seen from these regional examples, pluvial lakes are widespread (even in hyper-arid areas), reached enormous dimensions, and had different histories in different areas. Pluvials were not in phase in all regions and in both hemispheres (Spaulding 1991). In general, however, dry conditions during and just after the Late Glacial Maximum and humid conditions during part of the Early to Mid Holocene appear to have been characteristic of tropical deserts, though not of the south-west USA.

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A.S. GOUDIE

## POINT BAR

Point bars are a form of river bar. They are located along the convex banks of river bends. They typically have an arcuate shape that reflects the radius of curvature of the bend. The cross-sectional slope of the bar is inclined towards the centre of the channel, reflecting the asymmetrical channel geometry at the bend apex. Textural attributes of the bar reflect patterns of secondary helical flow over the bar surface as the thalweg shifts to the outside of the bend at high flow stage.

Point bars are most commonly found along meandering rivers where there are clear genetic links between instream processes that form and maintain pool–riffle sequences, the channel morphology that results, the formation of point bars



on the insides of bends, and resulting channel planform attributes (the meandering behaviour of the channel).

At bankfull stages, helical flow in bends carries sediment up the convex slope of point bars, while the concave bank is scoured. Sand or gravel bedload material is moved by traction towards the inner sides of channel bends via this helical flow. In this lateral accretion process, bedload materials are deposited on point bar surfaces. Lateral accretion deposits are detectable in the sedimentology of point bars by their oblique structures dipping towards the channel. Differing patterns of sedimentation are imposed by the radius of bend curvature (bend tightness) as well as the flow regime and sediment load.

Grain size typically fines down-bar (around the bend) and laterally (away from the channel). This produces a longitudinal 'around the bend' set of sedimentary structures comprising bedload material at the head of the bar, where the thalweg is aligned adjacent to the convex bank (at the entrance to the bend). As the thalweg moves away from the bend down-bar, lower energy suspended load materials are deposited. A mix of bedload and suspended load is generally evident at the tail of the bar. The most recently accumulated deposits are laid down as bar platform deposits at the bend apex. Typically these unit bar forms are largely unvegetated.

In many instances, point bars are compound features. These bank attached compound point bars comprise a mosaic of geomorphic units. In gravel situations, the bar platform is a relatively flat, coarse feature atop which a range of features are deposited or scoured. In the centre of the point bar is a gravel lobe. This likely represents the position of the shear zone during high flow stage. At high flow stage, compound point bars are dissected by chute channels, often with associated ramp deposits (McGowen and Garner 1970). Flow around vegetation produces a series of depositional ridges that have distinct grain size distributions (Brierley and Cunial 1998). The bar apex is a shallower feature inclined towards the channel.

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SEE ALSO: bar, river; channel, alluvial; fluvial geomorphology

KIRSTIE FRYIRS

## POLJE

A form of large, flat-floored closed depression formed in KARST regions. They are a distinctive feature of limestone geomorphology. The term comes from the Slav word for field. Poljes are often covered with alluvium, are subject to periodical flooding, and they may have sinks, called *ponors*, into which streams may disappear. There is much debate as to the origin of poljes. They are probably polygenetic, with solution and tectonics playing important roles.

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A.S. GOUDIE

## POOL AND RIFFLE

Pools and riffles are one of the most common and recognizable bedform sequences in channels. In the most basic sense, pools are classified as deep areas with low velocities at low stage, while riffles exhibit higher water-surface slopes and faster velocities. Pools and riffles exist in both alluvial and bedrock channels, but are best developed in gravel-bed substrates and meandering channels. Pools and riffles occur in moderate gradient channels with a transition to step-pool morphologies at higher gradients. These undulations in bed topography provide the primary framework for aquatic habitat in channels, and are of great interest because of their importance for macro invertebrates and fish species. The features also create tremendous form drag and flow resistance that may help rivers achieve equilibrium.

Because basic flow characteristics in pools and riffles change with stage (Richards 1976), limitations in the definition of pools and riffles exist. The zero-crossing method, where pools and riffles are recognized as residuals above a calculated mean bed profile, and spectral analysis provide robust methods for identifying the bedforms provided minimum size criteria are specified (Carling and Orr 2000). The residual-depth criteria or control-point method is used to define a pool based on the idea that riffles pond water in pools if flow ceases (Lisle and Hilton 1992). The residual pool is easily recognized in the field as the area that would be inundated, but it is still necessary to use minimum depth and width criteria to avoid subdividing a channel into extreme numbers of small morphologic units. Complications can arise because the residual depth of a pool is also influenced by sediment deposition, which can fill much of the pool at low flow. However, this fine sediment infilling can be used to estimate the sediment supply in a particular channel based on the idea that the residual volume of a pool prior to low-flow deposition is represented by the elevation of the coarse substrate below the overlying fines (Lisle and Hilton 1992). This assessment of pool sedimentation is particularly useful to evaluate land-use impacts on channels. It is also worth recognizing an important distinction between different types of pools where pools formed by some obstruction to flow are called forced pools and the remaining pools are termed free-formed pools (Montgomery *et al.* 1995).

The depth of pools and height of riffles is clearly maintained by some process, but the nature of the process is still debated. One idea that receives considerable attention is the velocity-reversal hypothesis. Keller (1971) used near-bed velocity measurements to show that velocities in pools are initially lower than those in riffles but increase at a faster rate and may exceed riffle velocities near bankfull stage. The water-surface elevations also change so that water-surface slopes equalized at high stage and may be steeper over riffles above bankfull level. The idea is closely related to the concept of two-phased bedload transport where low-flow deposition occurs in pools and high-flow deposition occurs in riffles (Jackson and Beschta 1982). Although the velocity-reversal hypothesis is often cited, it is frequently criticized based on continuity of mass concerns. However, recirculating eddies can form in some channels, increase velocities in pools and maintain continuity (Thompson *et al.* 1999). Meanwhile, the constrictions create backwater and locally elevated water-surface slopes in pools that can exceed water-surface slopes in riffles at bankfull stage. Evidence for recirculating-eddy enhanced velocity reversals exists, but flow routing of sediment around pools may dominate in other locations (Brooker *et al.* 2001). The combination of sedimentological properties and turbulence characteristics also is used to explain pool formation and maintenance. According to this idea, an obstacle to flow temporarily creates turbulence fluctuations, the perturbations to the flow generate the pool-riffle morphology, and differences in turbulence intensities and sediment characteristics between pools and riffles help to maintain disparities in sediment movement along the sequence (Clifford 1993). Much of the remaining work on pool formation and maintenance focuses on meandering channels and draws links between meandering and pool formation because pools tend to form at bends. Yang (1971) used this linkage and the idea that channels would adjust to minimize unit stream power along a channel in a theoretical approach with an equalization of water-surface slopes over pools and riffles at high stage. He concluded that the resulting formation process was a combination of dispersion and sorting of sediments. Studies also draw links between helical flow development and pool and riffle formation, but these processes are so clearly linked it is difficult to determine a casual relationship between them.

Pools and riffles exert an important influence on sediment sorting along a channel, especially the downstream end of pools, which create an uphill climb for particles moving downstream (Thompson *et al.* 1999). Sediment size, packing density and relative protrusion can all differ between pools and riffles. However, the combination of low-flow and high-flow sediment deposition can create a large variability in the size of bed sediments within a small area, and make it difficult to recognize distinct differences between pools and riffles (Richards 1976). Although channel-bed sediments in pools are often reported to be smaller than those in riffles (Clifford 1993), the opposite trend is reported in sediment supply-limited channels (Thompson *et al.* 1999). The disagreement in the general sorting trend probably results from two-phase bedload transport. During low flows, fine sediments can cover coarse substrate in pools along channels with high sediment loads, while supply-limited channels generally preserve the sediment-sorting patterns established at high flow.

Another fundamental characteristic of pools and riffles is the distance between successive pools or riffles, a measure termed the pool and riffle spacing. Values between five and seven average bankfull widths are often reported for pool and riffle spacing (Keller and Melhorn 1978), and this spacing has been attributed to reach-scale influences related to meander wavelengths in sinuous channels (Carling and Orr 2000). For example, variations in bed topography follow second-order autoregressive models as a result of a combination of periodic and random effects that may be related to meander wavelength (Richards 1976). Variation in spacing occurs because the channel-bed slope exerts a control on average spacing between pools (Wohl *et al.* 1993). Average spacing also varies due to the influence of obstructions to flow and variations in how pools are defined (Montgomery *et al.* 1995). In channels dominated by forced pools, local-scaling effects related to recirculating eddies behind randomly spaced channel constrictions can build morphologies with spacing values that agree with published values for a range of channel conditions (Thompson 2001). Therefore, it is unclear if reach-length or local-scaling effects create the semi-rhythmic spacing reported in natural channels.

Channel slope, channel-bed resistance and drainage area influence pool and riffle dimensions. Pool length and depth both tend to decrease with increased channel-bed slope (Wohl *et al.* 1993), and riffles become smaller in deeper water (Carling and Orr 2000). Given the fact that stream power increases with slope, the inverse relation between pool size and slope reflect changes in channel-bed resistance with more resistant beds associated with higher slopes. Pools also increase in size on larger channels, presumably because of the simultaneous increase in stream power with an increase in discharge on these larger systems. The relative magnitude of the bed undulations also tends to decrease with increased sediment supply because pools begin to fill and lose their distinct characteristics (Lisle and Hilton 1992).

As demonstrated by past research, pools and riffles will continue to be a central focus of research in fluvial geomorphology because of their important influence on both the physical and biological characteristics of natural channels.

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SEE ALSO: bedform; meandering; point bar; rapids; step-pool system

DOUGLAS M. THOMPSON



# PORE-WATER PRESSURE

REGOLITH and highly fractured rock at Earth's surface (here termed soil) contain voids (pores) that are variously wetted or filled with water (pore water). Forces acting on pore water establish gradients of fluid potential, the work required to move a unit quantity of fluid from a datum to a specified position, and pore-water flows in response to these gradients. The concept of hydraulic head usefully describes pore-water potential. Total hydraulic head, or potential per unit weight of fluid, is usefully described in terms of gravitational, pressure and kinetic energy potential. The total hydraulic head ($h$) for an incompressible fluid (fluid having a constant density; $\rho_w$ for water) is given by (Hubbert 1940):

$$h = \xi + \frac{p}{\rho_w g} + \frac{u^2}{2g}$$

where $\xi$ is the gravitational, or elevation, potential; $p/\rho_w g$ is the pressure potential, in which $p$ is the gauge pressure of the water relative to atmospheric pressure and $g$ is gravitational acceleration in the co-ordinate direction; and $u^2/2g$ is the kinetic energy, or velocity, potential, where $u$ is water velocity. Flow velocity in soil is usually very small, so calculations of hydraulic head in soils typically neglect velocity head. Pore-water pressure therefore constitutes one of two dominant components of the fluid potential in many soils.

Pore-water pressure is isotropic, but it varies with position relative to the water table (the depth horizon where pore-water pressure is atmospheric, which defines the zero-pressure datum) and with the proportion of soil weight carried by intergranular contacts. Below the water table, pore-water pressure is greater than atmospheric and positive; above the water table pore-water pressure is less than atmospheric and negative owing to tensional capillary forces exerted on pore water (e.g. Remson and Randolph 1962). If intergranular contacts carry all of the soil weight and water statically fills pore space, then the hydrostatic pore-water pressure ($p_h$) at a depth $z$ normal to the soil surface (Figure 127) is given by

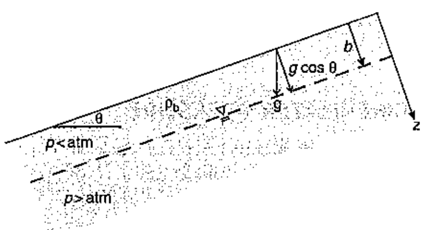


Figure 127 Schematic profile and definition of geometric parameters in an infinite slope having a water table at depth. Above the water table pore-water pressure is less than atmospheric; below it is greater than atmospheric

$$p_h = \rho_w g \cos\theta\,(z - b)$$

where $b$ is depth to the water table and $\theta$ the slope of the soil surface. Pore-water pressure can exceed or fall short of hydrostatic under hydrodynamic conditions or if a soil collapses or dilates under load. If a soil collapses, it compacts. Below the water table this will cause a transient increase in pore-water pressure, the duration and magnitude of which are governed mainly by the rate of collapse and the permeability of the soil. The gauge pressure ($p$) at depth $z$ can then be written as $p = p_h + p_e$, where $p_e$ is a nonequilibrium pressure in excess of hydrostatic. If collapse thoroughly disrupts intergranular contacts, then the pore fluid may bear the entire weight of the solid grains, and the soil will liquefy. In that case, gauge pressure can be written as

$$p = \rho_w g \cos\theta(z-b) + \rho_b g \cos\theta\, b + (\rho_s - \rho_w)(1-\phi) g \cos\theta(z-b)$$

where $\rho_b$ is the moist bulk density of soil above the water table, $\rho_s$ is grain density and $\varphi$ is soil porosity. When a soil liquefies, the excess water pressure equals the sum of the unit weight of soil above the water table and the buoyant unit weight of the soil below the water table:

$$p_e = \rho_b g \cos\theta b + (\rho_s - \rho_w)(1-\phi) g \cos\theta(z-b)$$

In unsaturated soil, such as occurs above the water table or (occasionally) when a saturated soil dilates (expands) under load, water does not fill pore space completely, and pore-water pressure locally is less than atmospheric. In that case capillary and electrostatic forces cause water to adhere to solid particles. As soil water content decreases, tensional forces increase and negative pore-water pressure bonds solid particles, increasing soil strength. The magnitude of negative pore-water pressure depends on soil texture and physical properties as well as on water content. Fine soils have a broader pore-size distribution and larger particle-surface area than do coarse soils. As a result fine soils have a greater range of negative-pressure potential because they hold more water than coarse soils, and the water bonds more tightly to particle surfaces. Piezometers are used to measure positive pore-water pressure; tensiometers commonly are used to measure negative pore-water pressure (e.g. Reeve 1986).

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SEE ALSO: debris flow; effective stress; landslide; liquefaction; mass movement; piezometric; undrained loading

JON J. MAJOR



# POSTGLACIAL TRANSGRESSION

The postglacial transgression (see TRANSGRESSION) has had many names, depending from the area and the time period studied. The name of Flandrian stage was first proposed in the late nineteenth century to indicate the 'Campinian sands' of Flanders and of the Anvers' Campine. Dubois (1924) included in the Flandrian stage all sediments, marine and continental, that characterize the displacement of the shoreline from the sea-level minimum, corresponding to the last glacial maximum, to the present situation.

In the British Isles, on the other hand, the deposits called Flandrian correspond more or less to those of the Holocene period, i.e. to the last 10 ka (Shotton 1973). An Irish variant for the Flandrian is the Littletonian, the base of which is placed near the base of a peat dated about 10,130 BP in a core, or at the maximum of *Juniperus* pollen in another core.

In the Mediterranean, Blanc (1936) described in the Versilia Plain the stratigraphy of deposits, *c.*90 m deep, covering all epiglacial and postglacial times, that he considered equivalent to the Flandrian transgression. The term Versilian was subsequently applied to other Mediterranean deposits by several authors, following the chronostratigraphic meaning proposed initially by Dubois (1924).

In Japan the postglacial transgression has received several local names: Yurakucho transgression, after the name of a marine Holocene bed in the Tokyo area, Numa transgression, after the raised coral bed of the Numa Terrace in the Boso Peninsula (Naruse and Ota 1984), or Umeda transgression, corresponding to deposits found in the Kinki area. In uplifted Japanese regions, the higher part of the marine deposits are also called Jomon transgression, after the age of the older stage of the Japanese Neolithic culture (between *c.*9,400 and 2,800 BP) (Takai *et al.* 1963).

In west Africa, the Nouakchottian episode, defined by Elouard (1966), consists of shell deposits, a few decimetres to over 2 m thick, which are found along most of the coast of Sénégal and Mauritania, at elevations close or slightly higher than the present sea level (between −2 m and +2 or +3 m). These RAISED BEACH deposits date from 5,500 to 1,700 BP and follow a shoreline that formed several gulfs. In the Ndrhamcha Sebkha area, where the coastline is now rectilinear, the Nouakchottian gulf extended into the continent about 90 km. Other similar transgression stories, each with a different name, could be added from other coastal regions of the world.

The use of many names to identify the same phenomenon in various areas, in order to attempt correlations, may have been useful in the past, when precise dating tools were rare or unavailable, but seems not to be justified today. Radiochronology has shown that the last maximal glacial peak occurred about 18,000 radiocarbon years BP, i.e. after calibration, *c.*21 ka ago (Bard *et al.* 1990). A more general use of the terms 'postglacial transgression' (for the last 21 ka) or 'Holocene transgression' (for the last 10 ka) would certainly contribute to clarify and unify the international terminology.

A marine transgression may occur, however, not only with a rising sea level, but even with a falling sea level if sediment supply is depleted and erosion can occur; conversely, a regression of the sea often results from a sea-level fall, but may also occur with a rising sea level in the case of high sediment supply and coastal progradation. Transgression–regression sequences, i.e. lithostratigraphic evidence of marine deposits inter-fingered with freshwater or terrestrial sediments, usually correspond to major changes in sea level. However, when the inter-fingered layers are not continuous in space and time, interpretation should be careful, especially in the case of Late Holocene sediments deposited near river mouths or near plate boundaries where tectonic displacements may have taken place.

In postglacial times, especially in the mid- to late Holocene, interpretation of transgression

-regression sequences has been reported from several coastal localities, again with various names. This was the case, e.g. on the southern coasts of the North Sea, where within the upper part of the Holocene transgression, three Dunkerquian transgressions were distinguished above Calais deposits (Tavernier and Moorman 1954). For each transgression a former sea-level position was deduced from the present elevation of the deposits. Correlations between the Flemish Dunkerquian stratigraphic sequences and other local deposits were subsequently attempted by various workers in France, the Netherlands and Germany, giving rise to the reconstruction of sea-level histories showing oscillations of varying amplitudes. Nevertheless, some fifty years later, the precise amount of these Dunkerquian sea-level oscillations, as well as their existence, remains to be demonstrated.

As SEA LEVEL is concerned, the last deglaciation seems to have occurred mainly in two eustatic (see EUSTASY) steps, with a first warming period that peaked at the Bölling (about 13–12 ka BP) and a second warming period after about 11.6 ka BP, separated by a temporary cooling (Younger Dryas). The melting histories of the various ICE SHEETS were non-synchronous and the last deglaciation ended in each place when the former ice sheet had completely melted. This seems to have occurred around 10 ka BP in Scotland and for the Cordilleran ice sheet, close to 9 ka BP in the Russian Arctic, around 7.5 ka BP in Scandinavia and around 6 ka BP for the Laurentide ice sheet. In Antarctica and Greenland, only the outer part of the ice domes melted.

The melting of the ice sheets caused considerable glacio-isostatic (see ISOSTASY) vertical movements of the Earth's crust: mainly uplift in unloaded areas, and subsidence in wide peripheral areas around the former ice sheets. The load of melted water on the ocean floor caused the latter to subside (hydro-isostasy). This is expected to have caused a flow of deep material from beneath the oceans to beneath the continents, with possible reactivation of seaward flexuring at the continental edges. Part of the above isostatic movements were elastic, i.e. contemporaneous to loading and unloading. Part continued, at gradually decreasing rates, for several thousand years after loading or unloading had stopped, because of the viscosity of the Earth's material. Part of such isostatic vertical displacements is still going on.

When the subsidence in areas peripheral to former ice sheets took place under the sea, it increased locally the volume of the oceanic container. This caused hydrostatic imbalance, with an indraught of water from other areas, decreasing the sea-level rise or even producing a slight sea-level fall in oceanic regions far away from the influence of former ice sheets. It is generally towards 7 to 6 ka BP, with the ending of ice melting, that slight emergence of isostatic or tectonic origin started to occur in many areas. Combination of the above processes produced a variety of relative sea-level and postglacial transgression histories along the world's coastlines. According to Mörner (1996), during the last 5 ka relative sea-level changes have been affected also by the redistribution of water masses due to changes in oceanic circulation systems.

In former ice-sheet areas, where marks of past shorelines on ice vanished with ice flowing and melting, local sea-level histories can be reconstructed only after deglaciation. They usually show continuing regression/sea-level fall, though at decreasing rates, up to the present.

In peripheral areas to the former ice sheets, the maximum postglacial sea-level peak has not yet been reached and the postglacial transgression is more or less still going on. Finally, in areas remote from the ice sheets and in most uplifting regions, a sea-level maximum, now emerged at variable elevations, has generally been reached towards the mid-Holocene.

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P.A. PIRAZZOLI

# POT-HOLE

Pot-holes are vertical, circular and cylindrical erosion features in consolidated rock of various lithologies. They are common in fluvial, fluviglacial and shore environments. Their sizes (diameter and depth) range from a few dm to a few m; but mega pot-holes many metres deep and wide also occur in formerly glaciated areas. Pot-holes are produced by abrasion, CAVITATION, dissolution and/or corrosion. A tool (pebble, sand) is necessary for abrasion whereas cavitation is a mechanical process of wearing created by a turbulent flow. Dissolution is active in carbonate rocks whereas corrosion, a more complex process, is manifested in most rock types, particularly in warmer climates. Many pot-holes have a complex origin. Shallow cavities made by cavitation or dissolution are often subsequently eroded through abrasion. Coastal pot-holes are less common than fluvial and fluviglacial. A few anthropogenic pot-holes on shore platforms have been reported in Brittany. The use of the term pot-hole for kettle and moulin should be avoided.

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JEAN CLAUDE DIONNE

# PRESSURE MELTING POINT

The melting point is the temperature at which a solid changes into a liquid. The application of pressure to a solid depresses the melting point, and the melting point under a given pressure is referred to as the pressure melting point. For ice, the lapse rate of melting point with pressure is $-0.072\,°\mathrm{C\,MPa^{-1}}$, the effect of which, for example, is to depress the melting point to $-1.28\,°\mathrm{C}$ under 2,000 m thickness of ice.

The pressure melting point is significant in glacial geomorphology because of its effect on the interaction of glacier ice with its substrate. Ice exists on the surface of the Earth at temperatures very close to, and sometimes at, its melting point. The application of pressure to Earth surface ice, as a result of either the hydrostatic pressure of ice overburden or the interaction of moving ice with undulations in the substrate, can lead to the depression of the melting point sufficient to allow melting. The process of pressure melting on the upstream over-pressured side of bedrock bumps, and refreezing of water on the downstream under-pressured side, plays a major role in glacier motion, and is also significant in the entrainment and transport of subglacially eroded debris. This REGELATION process allows material generated by subglacial abrasion and quarrying to be entrained in re-frozen layers, and is one process by which debris-laden ice can be added to the basal layer of a glacier or ice sheet.

WENDY LAWSON

# PRESSURE RELEASE

Many rocky outcrops show sets of horizontal or curvilinear fractures (SHEETING or EXFOLIATION structures) that are roughly parallel to the topographical surface. They are known with different names, some of them equivalent, such as pressure release, relief of load or offloading. It is obvious that all rock fractures are an expression of erosional offloading because only through the release of vertical and/or lateral pressure can the closed discontinuities become opened. But the gist of the pressure release concept is that rocks which cool and solidify deep in the Earth's crust (e.g. magmatic rocks), do so under conditions of high lithostatic pressure, i.e. loading by overlying materials (either rocks, sediments or even water or ice). So, many people suppose that when the rock outcrops in the Earth's surface it suffers a pressure release and this causes the development of stress and subsequent fractures parallel to the land surface. That is why the form of the land surface in broad terms could determine the geometry of the so-called pressure release fractures. According to this the fractures so generated would be secondary features (i.e. developed after the topography). But that is not always true because it is generally accepted that many plutons have been emplaced at shallow depths and the related structure was generated by the stresses imposed on magmas during injection or emplacement and, hence, so was the shape of the original pluton. Moreover, the so-called pressure release structures (i.e. sheet jointing) are well developed in rocks such as sedimentary and volcanic, which have never been emplaced, and even in granites, the magnetic orientation contemporaneous with the emplacement is clearly discordant to the pressure release structures. The pressure release theory may be questioned on several other grounds. Unloading appears to be mechanically incapable of producing fractures because if expansive stress developed during erosion, it would be accommodated along pre-existing lines of weakness and does not need to generate new ones, namely the sheet fractures. Another reason is that in fact several morphological and structural features developed on and in granitic rocks are incompatible with the tensional or expansive conditions implied by offloading. It is the case of structural domes, wedges and overthrusting associated with sheeting and is impossible to explain in terms of an extensional regime. Furthermore, evidence of dislocation and mylonitization along sheet fractures suggests that they are true tensional faults. Thus the pressure release structures may be better interpreted as primary features of the rock and accordingly the joints (see JOINTNG) were first developed in the bedrock and the shape of the land surface is a response to this previous internal structure.

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JUAN RAMON VIDAL-ROMANI

# PRIOR STREAM

Prior streams are Late Quaternary PALAEOCHANNELS of the semi-arid Riverine Plain in southeastern Australia. These ancient rivers were first described in the scientific literature by Butler (1958), who was given the task of producing soil maps in a region set aside for expanded irrigated agriculture in the period following the Second World War.

The 77,000 km$^2$ Riverine Plain consists of the coalescing floodplains of westward-flowing rivers of the southern Murray–Darling system. Despite its exceedingly subdued topography and low surface elevation (the great majority is less than 100 m above sea level), the Plain displays a complex pattern of sediments, soils and micro-topography. At first, the apparently featureless nature of the landscape frustrated Butler's attempts to make sense of his field observations. However, with the aid of aerial photographs, it became clear that well-drained sandy linear depressions that stood a little above the adjacent plain marked the locations of ancient aggraded palaeochannels that Butler called prior streams.

Soil variation on the Plain was controlled by proximity to a prior stream. Well-drained calcareous soils on prior stream levees graded laterally into heavy clays on the distal floodplain. Beneath the prior stream channels were thick beds of pebbly sand. As Butler mapped the regional soil landscape in more detail he discovered that the prior streams formed a complex distributary pattern (Plate 92) that petered out to the

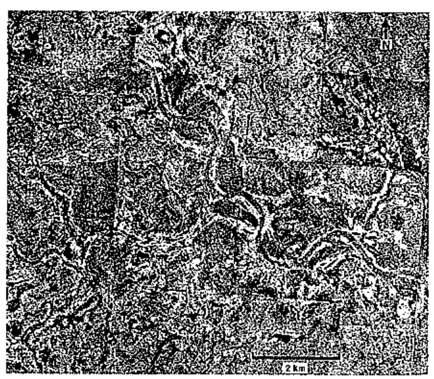

Plate 92 Air photograph mosaic of distributary prior stream channels on the Riverine Plain, Australia



west. Because the channels often intersected one another it was clear that there had been more than one period of prior stream activity. Butler invoked a cyclic model to interpret the different prior stream phases. Channel incision was thought to occur during more humid conditions when an absence of deposition permitted the development of well-organized soil profiles. The stable surfaces on which soils developed were called *groundsurfaces*. Channel aggradation occurred during more arid conditions when copious amounts of bedload sediment from upland catchment regions resulted in channel aggradation and extensive deposition.

## Phases of prior stream activity

The oldest groundsurface described by Butler was the Katandra. It was thought to represent a long period of soil formation under more humid conditions than exist at present. A switch to more arid conditions resulted in a new phase of fluvial deposition (Quiamong) that progressively buried the Katandra surface. As aridity intensified vegetation breakdown in the region to the west led to widespread clay deflation by westerly winds and the deposition of an extensive blanket of pelletal clay (Widgelli PARNA) which mantled the earlier Quiamong deposits. As the peak of aridity passed parna deposition waned and renewed deposition by Mayrung prior streams occurred. This final phase of prior stream deposition was followed by a long humid phase when stream incision occurred and well-developed soils formed across the surface of the Plain. Soils of the Mayrung groundsurface developed on fluvial and aeolian parent materials of Quiamong, Widgelli and Mayrung age. A generalized summary of Butler's stratigraphic units is shown in Figure 128.

The modern channels of the Riverine Plain developed in the post-Mayrung period. They occupy narrow floodplain trenches incised two to three metres below the Mayrung surface and are characterized by high sinuosity, low width to depth ratio and a dominance of suspended sediment. According to Butler, these younger Coonambidgal deposits display very weak soil organization.

## Post Butler

Not all workers agreed with Butler's interpretation of the prior stream deposits. The geomorphologist Langford-Smith (1960) argued that large meander wavelengths of the prior stream channels demanded greater discharges associated with late glacial pluvial, rather than arid, conditions. However, the absence of absolute dates on the prior streams (they all appeared to be beyond the radiocarbon limit of about 30,000 years) precluded any secure correlation with the glacial and interglacial episodes of the Late Quaternary.

Butler's early ideas were extensively revised during the latter part of the twentieth century. In the 1960s, Pels (1971) concluded that Butler's youngest stratigraphic unit, the Coonambidgal, was more complex than previously supposed. The early Coonambidgal phase was characterized by distinctive ancestral rivers that post-dated the prior streams and were the immediate precursors

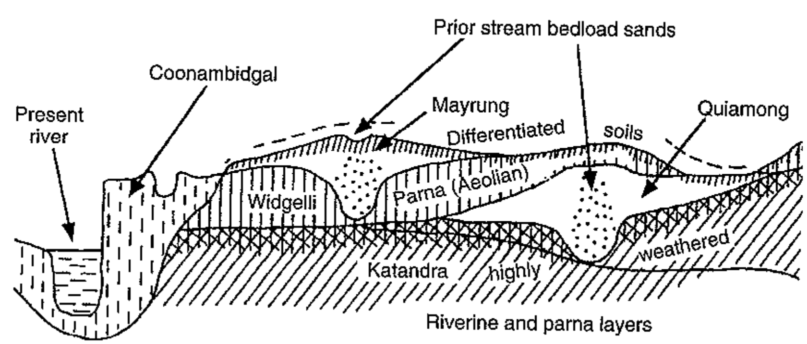


Figure 128 Generalized cross section of Riverine Plain showing Butler's (1958) prior stream sediments and soils (groundsurfaces)

of the modern drainage. The ancestral rivers were deep, sinuous, without levees and dominated by suspended load. They were much larger than the present rivers and maintained their courses across the Riverine Plain. Pels' model of Riverine Plain ancestral rivers and prior streams gained wide acceptance in the 1970s.

However, Bowler (1978), questioned the classification of ancient channels into two exclusive sequential types. He noted that both prior and ancestral attributes sometimes occur in different reaches of the same palaeochannel. In addition, Bowler found that some ancestral channels carried appreciable quantities of bedload, were bordered by sand dunes and mantled by well-developed soils similar to those of Butler's Mayrung groundsurface. Bowler concluded that Pels' separation of palaeochannels into two genetically different categories was unjustified.

Despite Bowler's misgivings, further progress on the nature and chronology of fluvial deposition awaited the development of thermoluminescence dating (Page *et al.* 1996). In brief, it was shown that four phases of palaeochannel activity occurred between approximately 100,000 and 12,000 years ago. Channel activity was characterized by alternations between sinuous, laterally migrating, mixed-load and straighter, vertically aggrading, bedload channel modes. Although the laterally migrating and vertically aggrading channel modes respectively approximate ancestral and prior streams, the sequence of channel activity was more complex than envisaged by Pels.

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KEN PAGE

# PROGLACIAL LANDFORM

The literal meaning of 'proglacial' is 'in front of the glacier'; it is the area that receives the products of glaciation. The proglacial environment is complicated, especially where warmer glaciers produce more meltwater. It can include terrestrial environments, streams, lakes and the ocean. The deposits include moraines, large outwash fans, deltas, marine fans and thick packages of sediment deposited in the marine realm. Other than where streams rework and erode previously deposited material, this is a depositional environment. The proglacial zone moves with the ice edge. As the ice advances, the zone of proglacial deposition moves forward as well. As the ice retreats, the proglacial setting follows the retreating margin 'backwards' and former subglacially deposited sediment-landform assemblages (see SUBGLACIAL GEOMORPHOLOGY) will be partially eroded, redeposited and/or buried.

The use of the term 'proglacial' has not been consistent in the literature. It is clear that there is a transition between ice-marginal and proglacial fluvial or lacustrine/marine processes. Some have suggested a further transition between proglacial fluvial processes and paraglacial processes, defined as any non-glacial processes conditioned by glaciation (Ryder, 1971). On this definition PARAGLACIAL processes strictly subsume proglacial fluvial processes. Thus the terms 'ice-marginal', 'proglacial' and 'paraglacial' have been used in overlapping senses and there is no universally agreed set of definitions. Table 37 classifies proglacial environments according to distance from the ice margin and lists the associated landforms. As most of these landforms are covered by separate entries (see cross references in the fourth column of the table) proglacial landform assemblages in the following will be viewed at the larger scale of ice sheet systems, mountain valley systems and subaquatic landsystems. A detailed review of these three landsystems can be found in Benn and Evans (1998).

## Ice sheet systems

The processes and patterns of proglacial deposition in front of ice sheets and ice caps are strongly influenced by glacier thermal regime. Temperate glacier margins are wet-based for at least part of the year. Meltstreams are the main agents of sediment transport and deposition. Ice-marginal forms are produced by alternating glacifluvial



Table 37 Classification of proglacial landforms according to different land-forming environments

| Environment | Process | Landform | See also: |
|---|---|---|---|
| Terrestrial ice-marginal | Meltwater erosion | Ice-marginal meltwater channels | Meltwater and meltwater channel Urstromtäler |
| | Mass movement/ meltwater deposition | Ice-marginal ramps and fans<br>Dump and push moraines<br>Recessional moraines | Moraine<br>Glacitectonics<br>Kame<br>Kettle and kettle-hole |
| | Glacitectonics | Composite ridges and thrust block moraines<br>Hill-hole pairs and cupola hills | <br><br>Ice stagnantion topography |
| | Meltwater deposition | Kame and kettle topography | |
| Subaquatic ice-marginal | Mass movement/ meltwater deposition | Morainal banks<br>De Geer moraines | Glacilacustrine<br>Glacimarine<br>Moraine |
| | Meltwater deposition<br>Debris flows | Grounding-line fans<br>Ice-contact (kame-) deltas<br>Grunding-line wedge | |
| Transitional from ice-marginal to fluvial | Meltwater erosion | Scabland topography<br>Spillways | Meltwater and meltwater channel<br>Outburst flood |
| | Meltwater deposition | Outwash plain (sandur)<br>Outwash fan<br>Valley train<br>Pitted outwash<br>Kettle hole/pond | Glacifluvial<br><br><br>Kettle and kettle-hole |
| Transitional from ice-marginal to lacustrine and marine | Meltwater deposition/mass movement | Deltas | Glacideltaic<br>Glacilacustrine<br>Glacimarine |
| | Deposition from suspension settling and iceberg activity | Cyclopels, cyclopsams, varves<br>Dropstone mud and diamicton<br>Iceberg dump mounds<br>Iceberg scour marks | |

and gravitational processes, locally modified by glacitectonic deformation. In addition, extensive proglacial rivers may rework glacigenic sediments. In some cases, virtually all the evidence of a temperate ice lobe is in the form of glacifluvial sediments. At temperate glacier margins with moderate debris cover deposition typically produces small dump moraines, or push and squeeze moraines derived from sediment exposed on the glacier foreland. During deglaciation suites of recessional moraines are commonly formed. The areas between recessional moraines often exhibit well-preserved subglacial landforms. At temperate glacier margins covered by considerable quantities of glacifluvial sediment uneven ablation of sediment-covered ice can lead to the development of a karst-like topography with a relative relief of up to several tens of metres. Sediment deposited in supraglacial outwash fans and lakes produces a complex assemblage of landforms including ESKER systems, kame ridges and plateaux, and pitted outwash. Distinctive spatial associations of glacifluvial landforms deposited during ice wastage can be recognized: proglacial outwash passes upvalley into kame and kettle topography. Kame and kettle topography is locally refashioned into suites of river terraces by proglacial and postglacial streams.

Subpolar margins are characterized by cold-based conditions near the snout and an upglacier wet-based zone. Subpolar margins are commonly affected by glacitectonic processes. Meltwater is available on the glacier surface but not on the bed. It can have some impact on ice-marginal landforms, but in general these are only locally reworked into outwash deposits. Where thick accumulations of unconsolidated sediments are present, terminal moraines in the form of composite ridges are constructed by proglacial tectonics. Where proglacial thrusting does not occur, ice margin positions are recorded by frontal aprons built up from fallen debris. Upvalley end moraines or frontal aprons pass into ice-cored moraines of chaotic hummocky or transverse ridges and/or kames. In the cases of totally cold-based glaciers and where bedrock is close to the surface the amount of available debris reaches a minimum. Former glacier margins in such areas are often marked by lateral meltwater channels cut into bedrock.

Entirely cold-based polar-continental glacier margins leave very little imprint on the landscape. Features that appear to be terminal and hummocky moraines often constitute a thin veneer of debris overlying buried ice.

The margins of surging glaciers typically have a very high debris content. Surging glaciers also tend to be associated with widespread subglacial and proglacial glacitectonic deformation. Near the margin there is often extensive tectonic thrusting, particularly if the substratum has been weakened by high pore-water pressures. In addition, large discharges of meltwater and sediment associated with glacier surges are responsible for major changes in deposition rates in proglacial lakes and sandar. Sharp (1988) described the geomorphic effects of a glacier surge cycle. During the advance (surge) phase, fluted tills and thrust moraines are formed. At the termination of the surge, crevasse-fill ridges form at the bed, then hummocky moraine and outwash are deposited during glacier recession.

Under circumstances where glacially dammed lakes are breached, exceptionally high magnitude discharges are generated and the proglacial environment will be characterized by extensive erosional landforms as well as outwash deposits. The Channeled Scablands region of Washington State for example derives its name from the dramatic erosional forms generated by the draining of glacial lake Missoula.

## Mountain valley systems

The majority of the debris transported by valley glaciers is derived from mass wasting of valley walls. Valley glaciers in high relief settings typically have extensive covers of supraglacial debris. Following ice retreat, high-relief mountain environments are subject to paraglacial reworking of ice-marginal sediments and landforms.

The margins of valley glaciers in mountain areas are commonly delimited by latero-frontal dump and push moraines. Meltwater deposition produces ice-marginal ramps and fans. In low relief mountain areas (e.g. the Scottish Highlands or the Norwegian mountains) Neoglacial end moraines are typically 2–5 m high. In debris-rich high relief settings the latero-frontal moraines can be much higher, so that repeated glacier advances may terminate at the same location and contribute to moraine-building, resulting in large landforms which can exceed 100 m in height.

Glacier retreat in mountain environments is normally recorded by recessional moraines. However, debris-mantled glaciers tend to remain at the limits imposed by their latero-frontal moraines until advance or retreat is triggered by significant climatic change. The landform record of a retreating debris-mantled glacier often consists of major moraine complexes, separated by extensive tracts of hummocky moraine deposited during episodes of ice-margin wastage and stagnation.

Glacial lakes are common features in mountainous environments. In low relief mountains they are mainly ice dammed as a result of the blocking of side or trunk valleys by expanded glacier tongues. Good examples of landform and sediment assemblages formed in Pleistocene lakes are found in the Highlands of Scotland and the water levels of former Glen Roy lake, for instance, are recorded by very prominent shorelines known locally as the 'Parallel Roads'. In high-relief mountain environments proglacial lakes dammed by moraines or by rockfalls, landslides and debris flow fans are more important than ice-dammed lakes. Some of the modern proglacial lakes impounded by latero-frontal moraines in the Andes and the Himalaya present a high risk of outburst floods to downvalley settlements.

Glacifluvial deposits are commonly well preserved in low to moderate-relief glaciated valleys. The focusing of meltwater flow by valley sides results in the erosion of gorges or the deposition of ribbon-like valley trains along valley axes. Staircases of terraces occur along the floors of many glaciated valleys, and the highest members may show signs of ice-marginal kame deposition. In valleys of high mountain environments mass movement features, lacustrine and fluvial sediment accumulations, and river terraces may be much more widespread than glacigenic landforms soon after their deglaciation. Paraglacial reworking of glacial landforms and sediments is less effective where glaciers advanced from high-relief mountainous regions to the foothills, and in such settings the preservation potential of the substantial ice-marginal landforms is greater. The margins of the Pleistocene piedmont glaciers in the northern foothills of the European Alps are delimited by high semicircular terminal moraines surrounding excavational basins. The ice proximal flanks of the moraines are characterized by kame and kettle topography and the central parts of the basins are either occupied by lakes or exhibit well-preserved DRUMLIN fields. The outer flanks of the moraines are skirted by large outwash fans, in which flights of several terraces were incised by postglacial rivers. For this spatial arrangement of landforms the term 'glacial sequence' was coined by Penck and Brückner (1909).

## Subaquatic systems

Given the extent of water-terminating glacier margins today and during the past, the subaquatic proglacial environment is an important one. More than 90 per cent of the Antarctic ice sheet margins terminate in the sea. The Pleistocene northern hemisphere ice sheets were bordered in many places by lakes hundreds of kilometres across, ponded between the ice and topographic barriers, or by epicontinental seas. In Europe, the largest proglacial lake was the Baltic ice lake which, around 10,500 years BP, stretched for some 1,200 km along the southern margin of the Scandinavian ice sheet. The most extensive of the North American lakes, inundating 2,000,000 km$^2$, has been named proglacial Lake Agassiz (Teller 1995).

The type of sediment-landform association deposited adjacent to the glacier grounding line depends on ice velocity and calving rate, sediment supply, input of meltwater from the ice, and water depth and salinity.

Sediment supply and subglacial discharges of meltwater are highest at temperate glaciers with a tidewater front. Large amounts of coarse debris are deposited in morainal banks and grounding-line fans, and fine-grained sediments are carried away in turbid plumes to form a distal zone of laminated mud deposits. At the grounding line of temperate glaciers ending as an ice shelf meltwater-related processes tend to be less important and grounding-line fans are less common. Grounding-line deposits pass into drapes of dropstone muds, released by meltout of sediments embedded in the basal zone of the ice shelf and further out into dropstone muds derived from icebergs. Iceberg-rafted debris is generally more important in the vicinity of ice shelves where icebergs are often trapped close to the ice margin by sea ice, whereas in the forefront of grounded temperate ice margins sea ice does not restrict iceberg drift.

In contrast to wet-based ice bodies, all glaciers and ice sheets, which are frozen to their beds, provide little debris and little meltwater. Grounding lines below ice shelves are associated with grounding-line wedges, composed of mass flow deposits. In the case of ice margins ending with a tidewater front hardly any sediment will be released to the lacustrine/marine environment and proglacial landforms are rare.

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CHRISTINE EMBLETON-HAMANN

# PROTALUS RAMPART

A ridge, series of ridges or a ramp of debris formed at the downslope margin of a perennial or semi-permanent snowbed, which is located typically near the base of a steep bedrock slope in a periglacial environment. Observations of active examples indicate that constituent sediments range from diamicton to accumulations of coarse rock fragı Roundness of rock fragments can vary from subangular to very angular, depending on the source of sediment supply. In planform, ramparts range from curved, to sinuous and complex. Typically, they have thicknesses (measured perpendicular to the slope) of up to 10 m. Examples on relatively steep slopes tend to have short proximal (i.e. adjacent to the snowbed) and long distal slopes.

The term 'pronival rampart' was preferred by Shakesby *et al.* (1995) on the basis that all examples lie at the foot of a snowbed (as 'pronival' indicates) but not all lie at the foot of a TALUS slope (as 'protalus' indicates). Until the early 1980s, there had been few observations of active forms. Suggested origins were based almost entirely on circular reasoning linking logical but hypothetical processes to supposed fossil 'ramparts', which might easily have been mistaken for other landforms (e.g. ROCK GLACIER, LANDSLIDE, MORAINE, avalanche impact ridge). It was reasoned that ramparts were formed entirely by coarse rockfall debris rolling, bouncing or sliding down a snowbed surface with very little if any fine debris reaching the rampart (White 1981). Any fine sediment in the ramparts, it was suggested, had been derived by *in situ* weathering or by the impacts of transported rock fragments. During the mid- to late 1980s, other processes (AVALANCHES and DEBRIS FLOWS) were found to be supplying fine as well as coarse material across snowbed surfaces to actively forming ramparts (Ono and Watanabe 1986; Ballantyne 1987). During the 1990s, the range of processes was expanded to include those operating beneath snowbeds, both as regards sediment supply (snowmelt, debris flow, SOLIFLUCTION) and modification of pre-existing sediment (bulldozing by a moving snowbed) (Shakesby *et al.* 1995, 1999). Because of confusion with other upland depositional landforms, there have been a number of attempts to identify diagnostic criteria, which have included morphological and sedimentological characteristics. In particular, attention has focused on the distinction between ramparts and moraines formed by small GLACIERS. Since, however, many 'rampart' characteristics have been based on (1) conjectural fossil examples, and (2) assumed formation at the bases of static snowbeds (although a rampart origin by a mobile snowbed has been demonstrated (Shakesby *et al.* 1999)), such diagnostic criteria must be viewed with extreme caution.

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SEE ALSO: nivation; periglacial geomorphology

RICHARD A. SHAKESBY

# PSEUDOKARST

A term first employed by von Knebel in 1906 (Bates and Jackson 1980), which has been widely used to describe topography, landform assemblages or features developed on non-carbonate rocks which exhibit a morphology similar to those characteristic of carbonate KARST terrain.

Such a lithology based definition excludes landforms in non-carbonate rocks from genuine karst. This classification is still in use by some geomorphologists and speleologists. However a more recent, all-encompassing definition of karst is now becoming increasingly widely accepted (Jennings 1983). Jennings's designation is less restrictive and he argued that karstic processes and landforms may be found on any rock type where the 'process of solution is critical, although not necessarily dominant'. Pseudokarst landforms should therefore be considered as those that morphologically resemble karst, but have formed through processes that are not dominated by solutional weathering or solution-induced subsidence and collapse.

Pseudokarst includes landforms morphologically similar to those commonly associated with carbonate or GYPSUM KARST landscapes, and include subterranean drainage, CAVES, DOLINES, BLIND VALLEYS, grikes, SPELEOTHEMS and surface KARREN.

Examples of pseudokarst fulfilling the conditions of Jennings definition, that is, where solution is not a critical formative process, include (1) caves in glaciers, or topographic depressions in permafrost regions (thermokarst), caused by a change in phase (i.e. from solid to liquid water) rather than dissolution (Otvos 1976); (2) VOLCANIC KARST, comprising tunnels within lava, formed where molten lava continued to flow inside an already partially solidified lava bed (i.e. caves are formed at the same time as the host rock) (Anderson 1930), and also depressions associated with the mechanical collapse of such caves; and (3) caverns and karst-like features caused by predominantly mechanical erosion of rock by animals such as abrasion caused by molluscs in the tidal zone of limestone outcrops on tropical and temperate coasts (Sunamura 1992), or moving water, wind or ice. Some workers also class as pseudokarst depressions and pipes (see PIPE AND PIPING) formed in soils or other unconsolidated sediments by the mechanical erosion of unconsolidated material (piping) (Otvos 1976) as, for example, often found within loess deposits.

In view of its original definition and long-term usage, the term pseudokarst can also be widely found in the literature referring to any karst-like features in rocks other than limestone (or gypsum) (including rocks such as basalt, granite or diorite) regardless of their mode of formation. Examples of these so-called pseudokarst features include basins, runnels, caves, underground drainage and even small speleothems. Provided these features can be ascribed to a range of physical or chemical weathering and erosive processes that do not rely on solution to any significant extent, the use of the term pseudokarst is appropriate.

The term pseudokarst has, however, also often been applied to landforms on rocks of relatively low solubility such as quartzites or highly siliceous sandstones, which consist almost entirely of silica ($SiO_2$) (e.g. Pouyllau and Seurin 1985). Such usage has been based on the widely held but incorrect assumption that quartzose rocks are practically immune to chemical weathering (Tricart 1972). This belief is based on the fact that the equilibrium solubility of many carbonate rocks ranges ween 250 and 350 $mg\,l^{-1}$ at normal temperatures, whilst under the same conditions the solubility of crystalline silica (quartz) does not exceed 15 $mg\,l^{-1}$, and even that of amorphous silica is less than half that of many carbonates. Quartzose rocks were thus generally considered not to develop solutional and therefore 'genuine' karst, but rather pseudokarst (e.g. Pouyllau and Seurin 1985).

However, during the past few decades features of considerable dimensions and striking morphological similarity to dissolutional karst have been identified in quartzose rocks in Africa, South America and Australia. For example, in Africa solutional landforms and caves are found in the quartz sandstones and quartzites of Tchad, Nigeria, South Africa and the Transvaal; the great South American quartzite landscapes of Brazil and the Venezuelan Roraima display numerous large and small, remarkably carbonate-like, surface forms, silica speleothems and many cave systems with lengths exceeding 2.5 km and depths of 350 m; and the quartz sandstones of the Arnhem Land and Kimberley regions of northern Australia and even the Sydney region of south-eastern Australia displays many caves, tower karst, smaller surface karren and speleothems (see Wray 1997 for a detailed review).

A range of studies carried out in these highly quartzose regions has now either argued or directly shown that the prime process leading to these 'pseudokarst' features is the direct dissolution of silica. Where quartz grains are held together by amorphous silica cement, the dissolution of this comparatively soluble material (up to about 150 $mg\,l^{-1}$) may isolate individual quartz grains from the parent rock (arenization) (Jennings 1983), which may then be removed by flowing water. However, arenization also occurs in rocks with very little amorphous cement when individual quartz grains and crystalline overgrowths are dissolved despite their low solubility (see especially Jennings 1983 for northern Australia; Wray 1997 for south-eastern Australia; and Chalcraft and Pye 1984 for South America). In a study investigating cave passages in the well-developed quartzite karst in Venezuela, Doerr (1999) has even argued that, under specific conditions, such karstforms may develop largely through dissolution, with arenization playing only a minor role.

A range of workers conclude that in many areas with quartzose rocks, dissolution is the key process in the formation of karst-like features, and argue that genuine karst may develop in highly siliceous rocks, where very long periods of weathering offset slow rates of dissolution (e.g. Jennings 1983; Chalcraft and Pye 1984; Wray 1997; Doerr 1999). Following the earlier urgings of Jennings (1983), Wray (1997) argued in a wide-ranging and comprehensive analysis of the worldwide karst-like features in quartzites and quartz sandstones, that in these features, the critical role of solution clearly identifies these forms as true karst (i.e. quartzite or sandstone karst) and not pseudokarst.

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SEE ALSO: biokarst; chemical weathering

STEFAN H. DOERR AND ROBERT WRAY

# PULL-APART AND PIGGY-BACK BASIN

Pull-apart and piggy-back sedimentary basins are typically associated with convergent plate tectonic settings (see PLATE TECTONICS). Pull-apart basins are topographic lows developed by rifting along strike-slip faults in areas of transtension (i.e. areas subjected to both transform and extension tectonics). The term 'pull-apart' was first used by Burchfiel and Stewart in 1966 to describe features in the Death Valley region of the USA and was later used by Crowell in 1974 to describe features along the San Andreas fault. Pull-apart basins have also been referred to as 'rhombochasm' and 'rhombograben' for the largest features (several kilometres by tens of kilometres in dimensions) and 'sag pond' for the smallest features (a scale of tens to hundreds of metres) (Seyfot 1987). The areas of transtension that develop pull-apart basins are typically associated with either (1) bends in the fault system (known as releasing bends) or (2) fault off-sets. These bends or off-sets need to step over to the left for a left lateral fault system or right for a right lateral fault system in order to generate the required transtension for basin development. The resulting basin is bounded on two sides by the strike-slip faults (which also have a significant normal component to the fault movement) and on the other two sides, approximately perpendicular to the main strike-slip faults, by normal faults. With continued extension of the basin the floor can be stretched and thinned to the extent that volcanism may occur and thus may cover the floor of the basin. Sedimentary fill of the basins may be developed in one main depocentre (area of maximum subsidence) in the central part of the basin or two depocentres, each adjacent to the bounding normal faults (Deng *et al.* 1986). Modelling by these authors suggests that the number and position of the depocentres is dependent on the geometry of the basin which in turn is dependent on three main factors (1) separation between the overlapping lateral fault strands; (2) degree of overlap between the main lateral faultstrands; and (3) the depth to the basement. Basins elongated parallel to the main lateral faults (overlap is more than separation) tend to have two depocentres, whereas 'shorter' basins where the separation is more than the overlap tend to have one depocentre. In most cases the depth of the basins typically tends to be greater than typical rift basins developed in divergent plate tectonic settings, and tends to be dominated by alluvial or lacustrine sedimentary fill.

Piggy-back basins, in contrast, are typically associated with thrusted terrain where basin development is complicated by deformation of earlier basin deposits by more recent thrusting. The term 'piggy-back basin' was first used by Ori and Friend in 1984 to describe minor sedimentary basins that rest on moving thrust sheets. Such basins have also been termed 'thrust-sheet-top basins' (Ori and Friend 1984)', 'satellite basins' (Ricci-Lucchi 1986) and 'detached basins' (Steidmann and Schmitt 1988). These basins are typically a few tens of kilometres across and are physically separated from the foredeep (the basin in front of all the active thrusts). Classic examples of piggy-back basins are found throughout the Alpine mountain chains of Europe. The basin fill comprises sediment sources from all basin margins, with a dominant provenance from the uplifted ramp of the older thrust behind the basins. Sedimentary environments range from coarse submarine fan and fan delta to fluvial deposits. Fluvial systems typically comprise a transverse drainage from the thrust ramps on both sides of the basin and a longitudinal drainage which enters the basin from the topographic lows that develop above lateral fault terminations (Miall 1999).

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SEE ALSO: fault and fault scarp; plate tectonics; rift valley and rifting; tectonic geomorphology

ANNE E. MATHER



# PUNCTUATED AGGRADATION

The theory that the long-term aggradation of sediment (through geological time) has been via episodic SEDIMENTATION. This is in contrast with the traditional concept of UNIFORMITARIANISM and the continual and gradual build-up of sediments through time. Early studies such as that by Barrell (1917) provided the initial challenge to the long-held paradigm of gradual aggradation. The theory of punctuated aggradation began to gather momentum once more in the early 1980s. Ager (1980: 43) fuelled the debate by referring to sediment stratigraphy as having 'more gaps than record', and argued that the large disparities between modern sediment deposition (for a specific environment) and ancient calculated deposition was a result of the episodic nature of aggradation. The theory of punctuated aggradation treats each bedding plane as a pause in sedimentation, whereas continual aggradation considers bedding planes as merely signifying a change in diagenesis or texture, and treats the formation as the basic stratigraphic unit, each one a product of a particular environment.

The term punctuated aggradational cycle (or PAC) was coined by Goodwin and Anderson (1985), within their hypothesis for episodic stratigraphic accumulation. The hypothesis argues that, allowing minor exceptions, the stratigraphic record consists of thin (1–5 m thick), basin-wide, shallowing-upward cycles. These are sharply defined by surfaces produced by geologically instantaneous relative BASE-LEVEL rises (termed punctuation events). Deposition occurs during intervening periods of base-level stability. A host of depositional environments can be included in the PAC hypothesis (e.g. fluvial, deltaic, shelf, slope, etc.), as PACs are assumed to exist in all depositional environments influenced by rapid base-level rises.

The PAC hypothesis proposes that allogenic processes such as sea-level change are responsible for changes in the stratigraphic record, rather than autogenic processes (e.g. channel migration, etc.) that are held as responsible in continuous aggradation. Autogenic processes are not dismissed entirely, but are treated as localized stratigraphic influences, superimposed on the allogenic processes. The bounding surfaces between the PACs are often traceable laterally for vast distances since they are formed by large-scale allogenic processes. This allows them to be accurate stratigraphic markers in the field. Base-level rise during a punctuation event can be rapid (reaching 1 m per 100 years) whereas stratigraphical analysis indicates that the recurrence of such punctuation events can be as frequent as 50,000 years, thus reflecting the rapidity of the base-level rise. Thickness of PACs, though generally thin, varies considerably though long-term aggradation rates remain similar. Goodwin and Anderson suggest that the most likely mechanisms responsible for PACs would include episodic crustal movement, episodic movement of the geoidal surface and global eustatic sea-level changes.

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STEVE WARD

# PYROCLASTIC FLOW DEPOSIT

Pyroclastic flow deposits are the products of fragmental material transported laterally by gas-charged, concentrated flows (sometimes called NUÉES ARDENTES). Pyroclastic flows are generated in many ways, with a spectrum from the 'passive' collapse of oversteepened lava-flow or dome margins, through the gravitational collapse of high eruption columns, to powerful overpressured blast-like events. In contrast to gravity-controlled lava flows (see LAVA LANDFORM) and water-charged LAHARS, pyroclastic flows may possess considerable momentum and cross substantial obstacles (sometimes >1-km high mountains). Pyroclastic flow deposits are so diverse that they are here described in terms of five spectra.

The first spectrum is in densities of the juvenile (newly erupted) component, which reflect the relative importance of expansion of dissolved volatiles in frothing and fragmenting the magma. Densities are higher (1.0–2.7 Mg m$^{-3}$) in many small-volume deposits, particularly those associated with composite VOLCANOES and/or the collapse of lava domes, where the magma is fragmented by external means such as crushing or shattering by interaction with water. In larger deposits, clast densities reduce to <1 Mg m$^{-3}$ (i.e. pumice), commensurate with an increasing role for expansion of dissolved volatiles. Simultaneously, contents of fine ash (<1/16 mm) increase; deposits with dense juvenile clasts have low contents (typically <2–5 per cent), those containing pumice have higher contents (>10–15 wt per cent). Dense clast-rich deposits often contain abundant large (dm–m-sized) juvenile clasts, and are labelled as, e.g. 'block-and-ash flow deposits' or 'dome-collapse avalanche deposits'. Deposits where the juvenile component is pumice are termed 'ignimbrites' or 'ash-flow tuffs'; collectively they represent by far the greatest volume of pyroclastic flow deposits worldwide.

The second spectrum is size. Distances travelled range from a few hundred metres in lava-collapse flows, to >150 km for prehistoric large ignimbrites. Areas range from a few thousand square metres to >30,000 km$^2$. Volumes range from about 1,000 m$^3$ for individual dome-collapse events to >1,000 km$^3$ for large ignimbrites. Observed pyroclastic flow eruptions generated only relatively small examples, with distances travelled up to 30–40 km, areas up to 400 km$^2$, and volumes up to ~15 km$^3$. Small pyroclastic flow deposits (<1 km$^3$) are generated from vents on composite volcanoes or from collapse of lava flows/domes. Intermediate-sized deposits (up to a few tens of km$^3$) can be generated from composite volcanoes or CALDERA volcanoes, often associated with caldera collapse. Larger deposits are associated with eruptions of gas-rich, evolved magmas (particularly rhyolite) from caldera volcanoes.

The third spectrum is deposit morphology. Individual, small-volume pyroclastic flows form tongue-like deposits, often with surface ridging, marginal levees and lobate flow fronts akin to those developed on DEBRIS FLOWS. However, most deposits form during many (tens to hundreds of) individual flow events, and so the gross deposit morphology then reflects the energetics of flow emplacement and deposit volume. The energetics are represented by the 'aspect ratio', which is the ratio of the average deposit thickness to the diameter of a circle with the same area as the deposit. Sluggishly emplaced deposits have a high aspect ratio (as high as 1 : 200), that is, the material is relatively thick for its extent. Energetically emplaced deposits have low aspect ratios (>1 : 10,000), that is, the material is very widespread for a given volume of material. The volume of the pyroclastic deposit coupled with its aspect ratio then yields three major morphologies: landscape-mantling, landscape-modifying, and landscape-forming (Figure 129). The largest deposits can create wholly new land surfaces over areas of >1,000 km$^2$, forming fan- or pediment-like surfaces around the source volcano.

The fourth spectrum is in the internal structure of the deposits. Single pyroclastic flows generate single flow units, that may be composed of a number of layers and facies that in turn reflect the mechanics of flow emplacement. Deposits of multiple flows should, in principle, show multiple flow units, but the clarity with which flow-unit boundaries can be discerned within such deposits is very variable. Thick stacks of ignimbrite may show no stratification, or only vague bedding or fluctuations in grain size, to suggest that they are the product of multiple flows. Grading structures within individual flow units vary widely also, and can reflect both migration of coarse clasts (regardless of density) under shearing forces, and

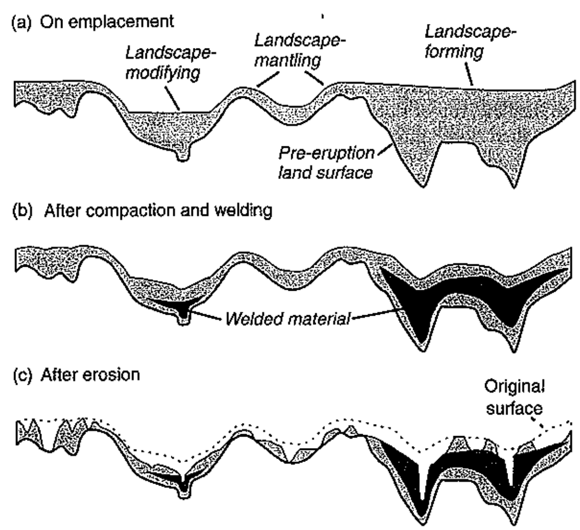


Figure 129 Schematic diagram to illustrate the morphologies of pyroclastic flow deposits. (a) On immediate emplacement, showing how the pre-eruptive landscape may be mantled, modified, or buried, depending on the thickness of the deposit, the topographic relief and the energetics of emplacement. (b) After consolidation and welding; note how compaction is greatest where the deposits are thickest, thus new valleys are generated along the line of pre-eruptive buried valleys. (c) After erosion; valleys are re-cut along their old courses. Non-welded deposits are preferentially removed, but summit heights may still be concordant, reflecting the original surface

flotation/sinking of lighter/denser coarse clasts, respectively, under buoyancy forces.

The fifth spectrum is in the lithologies of the deposits. Pyroclastic flows are efficient conservators of heat, and so many deposits are emplaced at temperatures above those at which the juvenile material can flow plastically (e.g. >550–600 °C for rhyolitic pumice). The combination of retained heat and load stresses imposed by overlying deposits causes the juvenile fragments to adhere and flatten (weld) to form a coherent rock. At its most extreme, welding can eliminate all initial pore space and the rock may be so hot as to continue to flow plastically as a kind of lava flow. Welding can only occur as long as the juvenile phase is glassy, but in most welded deposits the glass has subsequently devitrified. In addition, gases released from the juvenile material can cause further crystallization and vapour-phase alteration of the deposit, either along discrete pathways ('fossil fumaroles') or pervasively through the porous rock mass. Non-welded deposits show little or no JOINTING, but welding (and any other causes of induration) is generally accompanied by formation of jointing in the rock mass. The orientation and spacing of the joints can vary, but columnar joints, spaced at decimetres to metres apart, are characteristic of the interior of thick ignimbrites. Closer to the base, top or sides of the deposits, or in places where local fluxes of hot gases have occurred, the jointing can be more closely spaced and fan-like in disposition.

The morphologies of freshly emplaced pyroclastic flow deposits (Figure 129) are generally very rapidly modified by erosion, as loose pyroclastic-flow material is readily eroded, generating syn- and post-eruptive debris flows, lahars and HYPERCONCENTRATED FLOWS. Incision by streams often occurs so rapidly that interaction may occur between water and the still-hot interior of the deposits, leading to 'rootless' phreatic explosions. In non-welded deposits, incision rates of metres to tens of metres per rain event are known. Incision tends to recur along the lines of the pre-eruption valleys; the greatest thicknesses of deposits (and hence the greatest compaction) occur there and so the pre-eruptive topography is mirrored in subdued fashion on the surface of the deposits, controlling the paths of re-established streams. Erosion slows considerably when hard (welded) material is reached, or the non-welded deposits are stabilized by regrowth of vegetation.

Landscape morphologies seen in areas covered by pyroclastic flow deposits reflect a complex interplay between the initial depositional morphology, the presence or absence of welding or induration to create hard rock, and the local climate. A characteristic feature in dissected large ignimbrites is a concordance of ridge or summit heights, defining a surface parallel to the original deposit surface. Slopes in non-welded deposits are typically at or close to the angle of rest, except along streams or river where undercutting leads to vertical cliffs. Slopes in welded deposits are often cliffed, as the removal of material is controlled by vertical jointing that allows toppling of columnar masses as they are undermined by erosion.

Although pyroclastic flow deposits are volumetrically important in many volcanic terrains, the enormous variety of characteristics these deposits can display, and the hazards associated with flow emplacement, mean that there is still much to be discovered about the processes and products of pyroclastic flows.

## Further reading


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# Q

## QUICK FLOW

Hydrologists generally separate streamflow into two operationally defined components: event flow, considered to be the direct response to a given water-input event (also called direct runoff, storm runoff or stormflow), and base flow, which is water that enters from persistent, slowly varying sources and maintains streamflow between water-input events (derived largely from groundwater circulation). Quick flow is simply another term for event flow. The mechanisms involved may be one, or a combination, of Hortonian overland flow, saturation overland flow, and near-stream subsurface storm flow via groundwater mounding. In the latter case, at least some of the water identified as quick flow is 'old water' that entered the basin in a previous event. Quick flow can also be 'delayed', which involves storm runoff from distal sources via predominantly subsurface routes.

SEE ALSO: runoff generation

MICHAEL SLATTERY

## QUICKCLAY

The quickclays (quick clays, quick-clays, Swedish: *kvicklera*) are clay-sized postglacial marine sediments of very high sensitivity (see SENSITIVE CLAY). The term relates to the old Nordic *queck*, meaning living. They are found in Norway, Sweden and Canada, and to a much lesser extent in Alaska, Finland and Russia, and they have been defined as having a sensitivity of greater than 50. The original definition was: a clay whose consistency changed by remoulding from a solid to a viscous fluid. Very high sensitivity values have been found – up to 200 for the Champlain clays of east Canada. The literature is dispersed; there are reviews by Bentley and Smalley (1984), Cabrera and Smalley (1973), Maerz and Smalley (1985), McKay (1979, 1982), Brand and Brenner (1981) and Locat (1995). The high sensitivity value means that the clays lose most of their strength on remoulding, and this can lead to catastrophic landslides, which progress rapidly as flowslides. Soderblom (1974) proposed that two types of quickclays should be recognized: rapid quickclays and slow quickclays. The rapid materials lose their strength very quickly on reworking; but the slow materials require the input of a fairly large amount of energy before they convert to a liquid. The strength parameters of the remoulded clays can be difficult to measure.

The classic quickclay explanation by I.Th. Rosenqvist (1953) depended on postglacial uplift, and leaching. The clay material was deposited in shallow salty seas in immediate postglacial times. As postglacial uplift occurred these deposits became dry land and were exposed to rainfall and groundwater flow. This had the effect of leaching out the salts and changing the electrochemical environment of the soil particles. The loss of the soil cations meant that the system became more metastable and responded to stress via soil structure collapse, LIQUEFACTION and flowsliding. The Rosenqvist theory appeared to work for the rapid Scandinavian clays, but not to be so suitable for the slower Canadian clays.

As mineralogical analysis became more sophisticated it became apparent that in many quickclays the actual clay mineral content was quite low and that they were perhaps better described as very fine silts. This fitted in rather well with their observed distribution on the fringes of glaciated