

Chapter 7

The Relationship Between Continental Landscape Evolution and the Plant-Fossil Record: Long Term Hydrologic Controls on Preservation

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Abstract Continental depositional environments preserve the majority of the macrofloral record since the advent of land-plant colonization in the mid-Paleozoic, and wetland representatives are encountered more commonly than those that grew under more seasonal conditions. It has been assumed that preservation potential and future recovery of plant debris are high once detritus is introduced into any appropriate environment of deposition (e.g., fluvial-lacustrine or paludal setting), regardless of prevailing associated climate, sediment load, or geochemistry at the time of emplacement or interval thereafter. If a plant fossil is identified in any part of a stratigraphic interval, even if it occurs solely as an impression, it has been presumed that favorable conditions persisted over time to facilitate this record. Conversely, the

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absence of fossil plants in a stratigraphic sequence commonly has been interpreted as the result of catastrophic perturbation across the landscape, rather than the ascribing their absence to taphonomic filters that may have operated millennia after burial. Terrestrial landscapes are affected by aggradational, equilibrium, and degradational processes that control not only the local or regional water table, but also the long-term fossilization potential of organic debris entombed within these sediments. Fossil plants have the highest preservation potential when high water tables are maintained long-term within soils (e.g., histosols, entisols, gleyed soils), or in settings that are maintained below the maximum draw down of the regional water table (e.g., channel barforms, abandoned channels, lakes) of aggradational landscapes. When landscapes reach equilibrium, extensive pedogenesis ensues and the development of deep mature soils (e.g., calcisols) results in the bacterial degradation of any previously buried plant debris due to extreme penetration of atmospheric gases. When sediment is removed during landscape degradation, the local and/or regional water table is reset lower in the unconsolidated stratigraphy, once again promoting rapid decay of previously buried detritus at depth. These processes, operating under time frames of centuries to millennia and longer (lakh), control the ultimate preservational mode of plants recovered from the fossil record.

This chapter reviews the factors influencing the preservation of terrestrial plants in both subaqueous and subaerial environments based on actualistic studies, and develops a conceptual framework for landscape evolution in continental regimes. A model is presented in which preservational mode is related to the taphonomic and sedimentary history of the landscape in which plant detritus is buried. Case studies of the plant-fossil record, ranging from the Triassic to the Eocene, in exclusively aggradational and in aggradational/degradational landscapes are presented.

1 Introduction

Plant communities form the base of terrestrial ecosystems, serving multiple functions including, but not limited to: acting as the primary food resource for life; biogeochemical cycling and carbon storage; development and enrichment of soils; moderation of local and regional temperature; and animal habitats and shelters. Colonization of land may have occurred very early in the Phanerozoic, with evidence of cryptogam and bryophyte-grade plants found within nearshore marine deposits (Strother 2000; Baldwin et al. 2004) earlier than fragmentary debris of multicellular plants preserved in fluvial siliciclastic and associated environments (Pratt et al. 1978; Gensel and Edwards 2001). As plant clades evolved various architectures imparting more robust growth statures beginning in the Silurian, both aerial and subterranean plant parts become more prevalent in the stratigraphic record. Early preservational modes range from adpressions (Shute and Cleal 1987) to pyritization (Grimes et al. 2001) and charcoalification (Glasspool et al. 2004). With the advent of higher vascular plants in the Late Silurian (Edwards and Feehan 1980; Rickards 2000) and the appearance of the seed habit in the Late Devonian

(Fairon-Demaret and Scheckler 1987; Rothwell et al. 1989), plants evolved strategies for successful colonization under wide climatic regimes within various enriched or depauperate soil types. Individual clades developed a broad range of adaptations that allowed taxa to grow under moderate to extreme climates, although fossil assemblages rarely are preserved or encountered in these parts of the landscape for a variety of reasons (DiMichele and Gastaldo 2008). The principal preservational mode where such assemblages are found usually is permineralization (Demko et al. 1998) or authigenic cementation (Schopf 1975), promoted by physico-chemical groundwater conditions interacting with entombed vascular tissues during early diagenesis (e.g., Drum 1968; Allison and Pye 1994).

There is no doubt that the terrestrial plant record consists of an over-representation of wetland assemblages (Greb et al. 2006; Wing and DiMichele 1995). But, it is not unreasonable, and more parsimonious, to hold that deep time landscapes outside of the wetlands also were vegetated to some degree at least as far back as the latest Devonian, when the evolution of the seed habit allowed for biogeographic expansion of clades into more inhospitable regions. Of course, there does exist the possibility that community representatives could be preserved in these extrabasinal areas (e.g., Beraldi-Campsei et al. 2006), but their general absence in the pre-Tertiary stratigraphic record generally is construed to reflect their true absence in the landscape at that moment in geologic time. Lazarus taxa are known, although they are envisioned as having been wetland-centered species (Mamay 1992) and not representative of the remaining coeval landscape. Hence, the prevalence of wetland assemblages appears to have resulted in a prevailing paradigm that when plant-part debris is buried within a suitable depositional regime, early diagenetic processes generally will promote preservation. Conversely, when there is an absence of plant fossils in strata, it is assumed that the landscape was hostile to their successful colonization and they were extirpated from the region (often, in spite of paleopedological evidence) or that this absence marks a major extinction event. But, it is equally probable that the processes that promote preservation also will promote degradation and recycling via fungal and bacterial activity (Gastaldo 1994; Gupta and Pancost 2004). One must remember that the majority of organic matter is recycled for reuse, with a very small proportion of biomass sequestered in the rock record. Hence, concepts tying the presence or absence of the plant-fossil record to the evolution of continental landscapes generally have been overlooked or neglected.

The plant-fossil record plays a major role in understanding and interpreting the response of ecosystems to changes in climate, evolution, and crises in Earth systems. The physical presence (e.g., Gastaldo et al. 2009) or absence (Gastaldo et al. 2005) of terrestrial plants within any particular depositional regime at any specific point in time is integrally tied with the packaging of continental sedimentary successions. Recent workers have suggested methods and frameworks to subdivide continental rocks based upon differences in rates of accommodation and their relationship to allogenic and autogenic factors in basin fill (e.g., Shanley and McCabe 1994; McCarthy and Plint 1998; Etheridge et al. 1998). And with these models in mind, the preservation of terrestrial plant-fossil assemblages may be more a function of longer term processes operating within the landscape than any other factor.

It is essential to understand the relationships between the short-term (facies) and long-term (landscape) controls on the taphonomic biases controlling this record before using these data in paleoecological, paleoclimatic, and macro-evolutionary studies. This contribution will provide a model for continental sedimentary successions constraining the physico-chemical conditions within which terrestrial plant-fossil assemblages, and their preservational mode(s), can be understood and utilized.

2 Factors Influencing Plant-Part Preservation

Vegetated landscapes vary enormously in systematic composition, plant density, and vegetational architecture on continental spatial scales that are controlled by climate, topography, and edaphic conditions. Yet, the prerequisites that allow for preservation of the continental plant-fossil record are met under relatively restricted sedimentological and geochemical conditions (Krasilov 1975; Spicer 1989; Gastaldo 1992, 1994). These prevent incorporation of many types of plant assemblages in the deep-time record. In addition, there are a limited number of potential depositional sites where plant parts accumulate under conditions that may promote their long-term (10^1 to 10^{4-5}) preservation prior to deeper burial within any landscape. Aerial debris must accumulate within a depositional regime where (1) dysoxia and/or anoxia prevails (i.e., at the sediment–water interface in a lake system), (2) micro-environmental geochemical gradients are strong (e.g., fluvial channel-bar troughs; Gastaldo et al. 1995), (3) resistant and diagnostic phytoclasts persist unaltered long after decay has removed all volatiles (e.g., phytoliths – Strömberg 2004), and (4) sedimentary entombment maintains (preventing degradation) or enhances (through pore-water interactions with organic ligands) the geochemical environment within the facies. In the majority of instances, biomass is fated to be reused within various biogeochemical cycles by the living biota, which is the rule rather than the exception.

2.1 *Plant-Part Decay Rates*

The loss of vegetative and reproductive structures, either through physiological or traumatic disarticulation (Gastaldo 1992, 1994), subjects them to decay through a myriad of potential interactions including those with saprotrophs, fungi and bacteria, as well as autocatalytic cellular and subcellular breakdown. Although microbial films may promote preservation under specific environmental conditions (Dunn et al. 1997), phytoclasts have a tendency to degrade instantaneously in a geological sense. Subterranean rooting structures, already pre-entombed, may remain less affected for longer intervals of time, particularly when influenced by a change in pore-water chemistries promoting the precipitation of carbonate of various mineralogical compositions (calcite, siderite, pistomesite, etc.; Retallack 2001).

There is an abundance of actualistic data in the ecological and plant taphonomic literature focused on rates of forest-litter decay across the latitudinal (climate) spectrum (e.g., Bray and Gorham 1964; Gastaldo and Staub 1999 and references therein). In general, the refractory nature of the original biochemical composition of the plant organ, or part thereof, will influence the rate of decomposition which is calculated as the decay constant k (Perry 1994; Fig. 7.1). A leaf with a decay constant of $k = 1$ will be completely degraded in 1 year's time. But, complete decay can proceed within time frames on the order of weeks (e.g., flowers and leaves), months (leaves), or years to several decades (wood, gymnospermous cones, fruits, and seeds; Burnham 1993). It is well documented from neocological and actuo-paleontological studies that decay rates not only differ within taxa of a single clade or between various clades, but also under different climatic conditions (Gastaldo and Staub 1999). Rates even may vary within microhabitats under the same general climate (Bray and Gorham 1964). Hence, all plant assemblages essentially provide a geologically instantaneous (T_0) snapshot of preservable landscape constituents. It is true that certain resistant phytoclasts may be reworked, such as woody debris, heavily lignitized cones, fruits, seeds, charcoal (Scott 2000), and palynomorphs. Labile plant parts, such as leaves, flowers, and less reinforced reproductive structures, will sustain physical abrasion when re-entrained into bedload and reduced to unidentifiable phytoclasts (Gastaldo et al. 1987). But, published criteria allow for recognition of such recycled parts (e.g., wood-clast rounding – Gastaldo 1994; change in palynomorph fluorescence – Traverse 1994) within an allochthonous

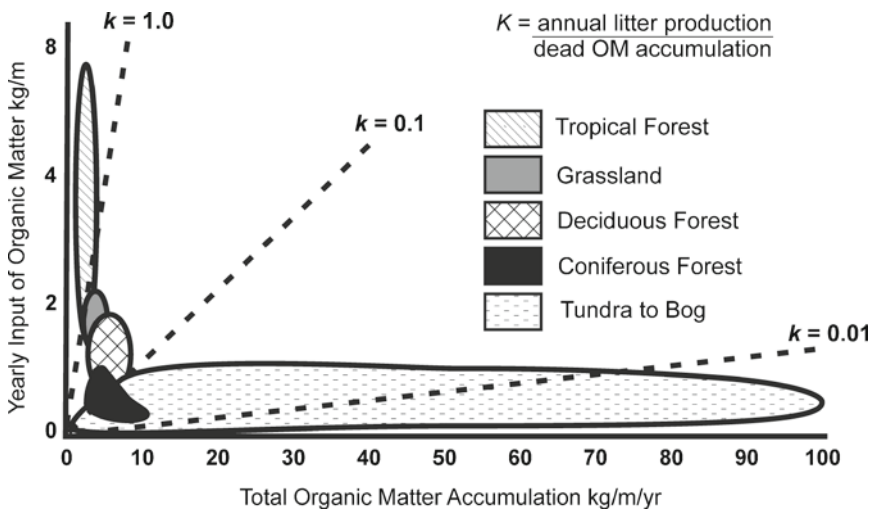


Fig. 7.1 The relationship between the yearly production of plant biomass and the total organic accumulation in various terrestrial biomes is described by the k constant (Perry 1994). A plant part with a decay constant of $k = 1$ will be completely degraded in 1 year's time. Note that decay constants differ between the most labile (leaves, fruits, flowers, etc.) and refractory (xylem elements, wood, amber/dammar, etc.) plant parts

assemblage. And, in reality, most phytoclasts do not possess structural attributes that allow them to be buried, exhumed, and recycled, if at all, more than once or twice (in the case of lignitized plant parts) before being reduced to palynofacies-grade debris.

2.2 Relationship Between Rates of Decay and Sedimentation

Inasmuch as decay rates of the most labile plant parts are, at best, on the order of months to only a few years, this rate exceeds average sedimentation rates in most instances, precluding potential preservation anywhere in the landscape. Sedimentation rate usually is expressed in cm/ka, which is an insufficient rate to promote plant-part preservation. Hence, a convergence must exist in nature where the sedimentation rate at some point in time exceeds the decay rate of plants for there to be any potential preservation of terrestrial vegetation in the stratigraphic record. In addition, the geochemical conditions associated with entombment that promote preservation must be maintained in both the short and long term for that organic debris to be identifiable.

2.2.1 Subaqueous Environments

Plant parts accumulate at the sediment–water interface within discharging and standing water bodies either when their specific gravity exceeds that of water (Gastaldo 1994) or when flow rate is reduced sufficiently to allow for settling from suspension (Spicer and Greer 1986; Spicer 1990). Assemblages within active channels include lag deposits of wood and carpological (fruit and seed) remains (Gee et al. 1997; Gee 2005). Dense accumulations occur within both channel bottoms and various barforms, as well as isolated coarse woody debris (CWD) scattered within the system (e.g., Fielding et al. 1997; Alexander et al. 1999; Gastaldo 2004). In contrast, coarse woody assemblages also have been recognized at the top of fluvial channel fills preserved as log jams (Gastaldo and Degges 2007). Such relationships require an understanding of the contextual taphonomic framework before interpreting Late Devonian (Meyer-Berthaud et al. 1999) to Recent dense woody assemblages in the fluvial stratigraphic record. Troughs within and between fluvial barforms, particularly point bars and lateral barforms, are sites where an admixture of aerial plant parts tend to accumulate when conditions allow for suspension-load settling. This may be in response either to a decrease in discharge velocities following seasonal changes in water supply, the lowering of river stage following a high discharge event (either bankfull or flood stage; e.g., Scheihing and Pfefferkorn 1984), or interactions with meso- to macro-tidal processes transforming a free flowing river to a standing body of water at tidal-bore turnaround (e.g., Gastaldo et al. 1996a). Similarly, organic drapes consisting of various phytoclast components often are found at bounding surfaces separating foreset laminae created by bedform migration. Preservation potential of all

these assemblages increases when they are buried by continued bedform migration and maintained below the air–water interface. Geochemical properties inherent within the accumulation, such as the release of organic acids and tannins, may promote long-term preservation in spite of the fact that pore-waters are in chemical equilibrium with the water column. But, when water stage falls to a level below that of the buried organic debris, sediment oxygenation and fluctuating redox conditions promote bacterial and fungal activity that reduces most identifiable plant parts to palynofacies-grade detritus (Gastaldo, 1989). This results in an organic residue in which only the most resistant phytoclasts (e.g., palynomorphs, cuticle, and structured organic matter = mesofossils fraction) may be recovered.

Standing bodies of water have the highest preservation potential for plant-assemblage preservation. These include settings that one envisions as stereotypical lakes and ponds, although the blockade of drainage systems within watersheds either through mass wasting or effects of volcanogenic activity within active tectonic settings also will result in a standing body of water equivalent in scale to lakes (Spicer 1989). In addition, plant parts will accumulate within inactive and abandoned fluvial (oxbow; Gastaldo et al. 1989) and tidal (Gastaldo and Huc 1992) channels. Plant parts transported through feeder channels into lake bodies often are sequestered in shallow water, Gilbert-type deltaic deposits (Spicer and Wolfe 1987) where preservation may result if lake levels are maintained. Lake margins vegetated by aquatic and semi-aquatic plants may act as filters, trapping organic debris in the shallows (Gastaldo 1994). But, once lake level falls and sub-erially exposes these areas, buried organic matter shares the same fate as accumulations noted above within barforms. And, if lake levels fall significantly, pedogenesis will overprint these to some degree (e.g., Wing 1984; Gastaldo et al. 1998). Assemblages that accumulate at marked water depth in more distal parts of the water body have a higher probability of preservation if several physical (associated high sedimentation rate) and chemical (i.e., redox conditions operating at and below the sediment–water boundary) conditions are met. Otherwise, debris that settles to the sediment–water interface will be recycled via microbial, invertebrate, or vertebrate activity.

Abandoned (or blocked) channels remaining in connection with an active fluvial (Gastaldo et al. 1996b) or tidal (Gastaldo and Huc 1992) regime provide very localized sites in which primarily parautochthonous assemblages accumulate in association with high sedimentation rates resulting in their entombment. The maintenance of high water levels in a largely confined and restricted setting promotes redox states wherein reducing conditions are not controlled, necessarily, by acidic waters but, rather, by the development of strongly negative Eh values in the sediment. The pH in these water bodies actually may be near neutral (e.g., Gastaldo and Huc 1992), yet the rate of decay is retarded and even labile tissues are conserved in the subsurface. In such settings the promotion of bacterial films, as identified by Dunn et al. (1997), developed at the sediment–water interface in conjunction with sediment influx, may be controlling the preservation potential in these assemblages. Hence, there is a localized geochemistry with its own internal equilibrium very different than the surrounding landscape that controls the taphonomic character of these regimes.

2.2.2 Subaerial Environments

In subaerially exposed floodplain settings, it is generally held that plant-fossil assemblages have the highest preservation potential when buried by overbank flood deposits. High discharge, sediment-laden flood events allow for the spread of fine-grained clastics away from the trunk channel over a short time interval that may last a few weeks before flood waters recede and suspension load settles to what previously was the soil-air interface (Fig. 7.2). During flooding, the regional water table is elevated to above the soil interface and standing water may promote site-specific dysoxyia and a change in redox conditions. But, once flood waters recede and the regional water table is reestablished at its previous level, normal decay processes proceed within the buried litters as groundwater levels fluctuate in response to meteoric water supply, evaporation, and plant growth (transpiration). Fluctuations in the groundwater table introduce oxygenated waters promoting biochemical and biogeochemical processes, reducing plant parts to meso- and micro-detrital sizes without physical abrasion. In effect, the return to the local or regional equilibrium resets the degradational processes, reducing to null the potential for preservation as

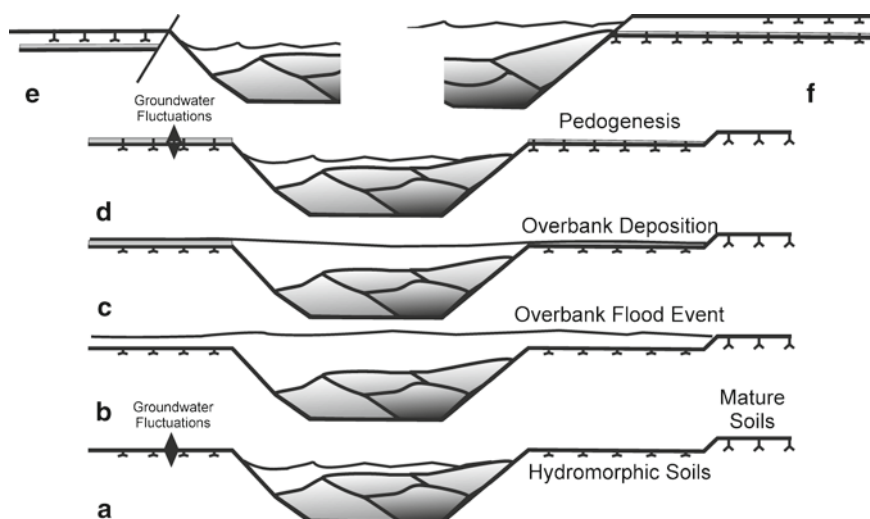


Fig. 7.2 The fate of organic matter at the soil-air interface within interfluvies. (a) Groundwater fluctuations in hydromorphic (wetland) soils are responsible, in part, for decay of organic matter at and immediately below ground level. (b, c) During overbank flooding, fine-grained sediment is transported into the interfluvies where it is deposited from suspension load as a thin to thick blanket over the former soil. (d) As flood waters subside and return to the original river stage, groundwater table also returns to its previous placement in the landscape. Pedogenic and biological activity, along with yearly-centennial fluctuations in groundwater table, promote decay of buried plant debris. To isolate the buried litter (O-) horizon and increase the probability of preservation, the surface must be placed beneath the regional watertable. This may be done either in response to tectonic subsidence (e) or rapid aggradation of the landscape (f)

pedogenesis becomes the primary mechanism operating within the landscape. This scenario also holds for landscapes influenced by avulsion unless a perched water-table is established, preventing root penetration, oxygenation, and microbial decay of buried debris. Hence, it is only possible to preserve subaerial plant assemblages that accumulate at the soil–air interface by not only (1) entombing the debris within overbank (and avulsion) deposits that are relatively thick, but also (2) in an area where the post-depositional groundwater table has been elevated above the former soil profile to insure that subsequent pedogenesis and associated biological decay processes are eliminated (Fig. 7.2e, f).

Plants in volcanogenic regimes (Spicer 1989, 1991) may be encountered in a variety of settings ranging from autochthonous litters, preserved at the contact between the soil and ashfall deposits (Burnham and Spicer 1986; Wing et al. 1993; Opluštil et al. 2007), to allochthonous assemblages encased within lahars and debris flows (Fritz 1980; Fritz and Harrison 1985) or reworked tuffaceous sediments (Jacobs et al. 2005). Besides the possibility for the presence of charcoal, either the result of temperatures experienced during the blast event (Spicer 1989) or following burial (Scott and Glasspool 2005), very early diagenetic interactions as a result of reactive pore-water chemistries may promote sulfide precipitation enhancing preservation potential of more labile parts (Burnham and Spicer 1986). Such rapid mineralization may allow for identification of those plants in the stratigraphic record subsequent to the reestablishment of regional groundwater table and promotion of degradation in the buried assemblages. In other instances where mineral-charged groundwaters are transported through xylary conducting tissues (tracheids in plant steles and tracheids and/or vessel elements in wood), reactions with organic ligands at the boundaries of cell walls may result in silicification (Sigleo 1978, 1979). Subsequently, abiotically mineralized plant parts will be retained in the stratigraphic record even when others may be removed by degradation. Hence, even in explosive volcanogenic regimes where sedimentation rates exceeds decay rates the control on potential preservation is linked with maintaining the assemblage within the phreatic zone.

3 Models of Stratigraphic Frameworks and Landscape Evolution

With the advent of the sequence-stratigraphy paradigm and its primary application to marine-influenced sedimentary successions, extension of sequence boundaries onto continental terrains became necessary to understand how these were expressed in subaerial environments. Initial models for continental regimes relied upon base-level changes tied to eustacy (e.g., Posamentier and Vail 1988; Miall 1991; Shanley and McCabe 1991, 1994), although Wright and Marriott (1993) note that models controlled by simple base-level changes can not be applied to fluvial systems where complexity is controlled, in part, by climate. They also dispute the concept of available accommodation in continental settings by noting that floodplain sedimentation

not only is controlled by the elevation of the channel but also by its bankfull depth. More recently, Muto and Steel (2000) extended the concept of accommodation, *sensu lato*, when they argued that the term “potential accommodation” is equivalent to the maximum possible accommodation that essentially coincides with water-column height at a specific place in time. Hence, the depth and extent to which sediment can accumulate within any continental regime will be controlled by both the water level within fluvial conduits (including avulsion channels) and at over-bank stage when sediment is distributed between channel systems that, in effect, increase the height of confining levees.

3.1 Continental Sequence Stratigraphy

As originally conceptualized, the sequence stratigraphy of continental successions was placed within the context of base-level change within coastal plain and deltaic, marginal marine settings (Posamentier and Vail 1988). Posamentier and Vail (1988) differentiated between fluvial deposition in the coastal plain, at or just above sea level, and in the alluvial plain, above sea level. In the coastal plain, fluvial deposition was constrained to incised valleys during lowstand, and during the latter stages of highstand progradation of the shoreline depositional system. Several authors were critical of this framework because of the fact that fluvial architecture is related to concepts of equilibrium-profile change over time (Miall 1991; Wright and Marriott 1993). With this in mind, Wright and Marriott (1993) modified the model relating base level and accommodation to alluvial architecture and soil development. Although still visualized within the context of eustasy (e.g., Blum and Törnquist 2000), fluvial sediment accumulation was constrained to the coastal plain in transgressive and highstand systems tracts when there is sufficient accommodation to store sediment within the system. But, as Wright and Marriott (1993) note, their model is only one of many possible scenarios.

The range of possible controls on stratigraphic base level within continental successions was discussed by Shanley and McCabe (1994) wherein they noted an interdependency of climate, tectonism, and eustasy that may determine fluvial architectural patterns. Their and subsequent concepts of a continental sequence followed that of Mitchum et al. (1977) in which genetically related strata are bounded at the top and base by unconformities or their correlative conformities. Such boundaries mark a depositional hiatus, the origin of which may have been controlled more by climate and/or tectonic processes within continental sequences than in marine ones, and are reflected in the character of interfluvial paleosols developed within coastal plain settings (e.g., McCarthy and Plint 1998, 2003; Plint et al. 2001). Debate continues as to how to confidently identify the boundaries of marine systems tracts within correlative alluvial strata (e.g., Etheridge et al. 1998), as well as the application of terminology devised for marine sequences used in continental settings. General models for continental sequences continue to be proposed (e.g., Boyd et al. 2000) although all models, to date, tend to focus on eustatically influenced

coastal plain deposits to a significant degree (Atchley et al., 2004). In this paper, we use the term *continental* to refer to settings in the alluvial plain, in the sense of Posamentier and Vail (1988), which are beyond the reach of marine influences.

3.2 *Graded Profiles, Paleosols, and Landscape Evolution*

As many authors have noted (e.g., Bull 1991; Quirk 1996; Posamentier and Allen 1999; Muto and Steel 2000), fluvial systems are in a state of disequilibrium when they are doing stratigraphic “work” (depositing or eroding). Fluvial equilibrium can be defined as a state when there is no significant erosion or deposition occurring along the course of a river that would permanently changes that system’s overall profile (Machin 1948; although see arguments about theory and reality discussed by Muto and Steel (2000)). This hypothetical graded state results in a rate of bedload transport equivalent to that of sediment supply; hence, there is a balance between the energy required to carry the bedload and the amount of bedload transported (Quirk 1996). Disequilibrium occurs when either there is: (1) insufficient discharge or a decrease in alluvial gradient that results in a decrease in stream power to carry bedload through the system, resulting in aggradation of the alluvial plain (positive fluvial accommodation of Posamentier and Allen 1999), or (2) higher discharge than that required to move bedload, or an increase in alluvial gradient that results in an increase in stream power causing degradation (downcutting) of the alluvial plain by incising channels (negative fluvial accommodation of Posamentier and Allen 1999). Hence, fluvial systems only contribute additional strata or surfaces of erosion to the stratigraphic record when in disequilibrium, although, as Schumm (1993) pointed out, reworking of the landscape occurs due to channel and channel belt migration during times of relative stasis. Disequilibrium is promoted by changes in allogenic forcing factors such as tectonic subsidence or uplift in the basin, climate and precipitation regimes (Cecil and Dulong 2003), relative height of ultimate base level (sea level or, in the case of closed basins, interior lakes), and processes operating in both the provenance area and coastal zones (baselevel change; Schumm 1993). Quirk (1996) remarks that when climate or the position of relative sea-level are influencing factors, large parts of the drainage basin are likely to be affected. And, as such, intervals of aggradation and degradation may provide a means of chronostratigraphic correlation in alluvial strata. With this in mind, Quirk (1996) introduced the concept of base profile, an idealized graded profile of a drainage basin constrained in time to a potential chronostratigraphic datum. Rivers aggrade up to, or degrade down to, this base profile (referred to as the “base level of erosion” by Bull (1991)).

Substantial attention has been focused on sandstone-prone channel facies in alluvial plain stratigraphy due to their economic importance in hydrocarbon and groundwater exploration and exploitation. Yet, interfluvial areas (floodplain, overbank, etc.) make up the majority of any alluvial plain, with the trunk drainage channels of major rivers covering a relatively small component of the overall landscape. Hence,

the character of paleosols adjacent to, and coeval with, within-channel deposits provides critical information about the prevailing climate (e.g., Sellwood and Price 1993; Retallack 2001) and the relative timing of their formation. Because paleosols that form under a distinctive climatic signature represent an amount of time that is roughly equivalent to, or less than Milanković-scale cycles, they, in effect, allow for chronostratigraphic control in alluvial strata (McCarthy and Plint 1998, 2003; Plint et al. 2001).

When fluvial systems are in equilibrium, paleosols in adjacent areas are not provided with any significant new sediment supply, outside arid and semi-arid areas where aeolian deposition may dominate. During this time, climate is the predominant influence across the landscape and the result is the development of a mature, complexly overprinted, or amalgamated paleosol with features reflecting the physio-chemical processes of formation (e.g. histosols – high water tables and organic production, vertisols – seasonal wetting and drying, aridisols – accumulation of salts, etc.). More mature paleosols record longer durations over which the fluvial regime remained in equilibrium providing for increased depth of pedogenic alteration. When fluvial systems are in disequilibrium, one of two general scenarios will ensue. Landscape degradation will result in the loss of alluvial plain stratigraphy due to increased downcutting and floodplain scavenging. Potential evidence for the conditions that prevailed during equilibrium may be restricted to channel lag deposits (e.g., reworked soil nodules: Pace et al. 2005) or aerially restricted sites that were unaffected by, or resistant to, erosion (e.g., Gastaldo et al. 1998). Conversely, landscape aggradation will result in the rapid buildup of interfluves with a concomitant rise in watertable. These conditions promote the development of immature and wetland paleosols (Kraus 2002).

In summary, continental landscapes can be envisioned as existing within one of three different states at any point in time during their histories – degradation, aggradation, and equilibrium (stasis). During degradation, fluvial systems are downcutting, removing previously deposited material or bedrock because of excess stream power. The regional water table will follow the downcutting, and floodplain sediments that may have been stored under saturated or submerged conditions will be subject to deeper drainage, infiltration, and more oxidizing conditions. During aggradation, fluvial systems deposit material, filling up potential accommodation in order to increase the gradient to a point that is adequate to carry the bedload that is available. Once this gradient is reached, the fluvial system is in a state that will carry the bedload provided. As the system aggrades, the regional watertable will rise with the increased elevation of the fluvial/overbank system. Previously deposited channel-fill and floodplain sediments will be buried below this rising water table and will be cut off from further connection to vadose-zone weathering and/or oxidation. If this aggradation is taking place within a valley cut during a previous period of degradation, the fluvial/overbank system will have to first fill the confined space within the valley before being able to aggrade the regional landscape surface. This initial period of valley filling may produce a landscape that has both well-drained oxidized soils on interfluve areas and poorly-drained areas with relatively wet edaphic conditions within the incised valley. Since only the valley floor is available for

aggradation and storage of channel-fill and overbank material, relative sedimentation rates may be high while the valley topography is annealed. The volume of valley accommodation also may increase because of lateral retreat of the valley walls due to cutbank erosion, slope wash, and mass-wasting during aggradation. If the potential equilibrium profile of the system is at a higher elevation than the interfluvial areas, the fluvial system will overtop the valley walls and will have a greater area over which to deposit channel and overbank sediments. At this time, relative sedimentation rates will apparently decrease, and there will be evidence of a related increase in floodplain soil maturity. Systems subject to repeated or cyclic changes in the forcing factors that control landscape aggradation and degradation (discharge, bedload sediment supply, tectonically-controlled gradient, etc.) may have successive valley fill and post valley-fill strata that have been subject to superimposed pedogenic and paleohydrologic regimes. The preservation of plant-fossil assemblages with identifiable remains within these successions will be controlled by the magnitude and timing of these depositional and hydrological processes.

4 A Model for Plant-Part Preservation in Continental Landscapes

The potential for encountering fossil plants within a continental framework is related directly not only to whether or not the organic detritus accumulates within a potential preservational site, but also whether or not it is maintained in a burial site where pore-water geochemistry retards or halts degradation. The highest probabilities for preservation exist within bodies of water in which fine-clastic sediments accumulate at rates greater than those for organic-matter decay. Conversely, the lowest probabilities exist within subaerially exposed sites, including wetlands (Gastaldo et al. 1989), where groundwater flux and meteoric water input vary over diverse time scales promoting pedogenesis and carbon recycling. Hence, for plant parts to become recognizable (identifiable) biological remains in the stratigraphic record, they first must be confined to stratigraphic intervals where high A/S (Accommodation/Sediment Supply) ratios prevailed. And, once buried, these assemblages also have to be maintained at a stratigraphic level beneath the maximum draw down of the prevailing watertable to prevent subsequent decay, deterioration, and loss from the sedimentary succession. Envisioned within the framework of continental landscape evolution, there are limited potential instances where these criteria are met in space and time.

During periods when the landscape is static and fluvial systems are in equilibrium, plant-part preservation is confined to: (1) within channel, subaqueous deposits (CWD concentrated in basal lags or barforms, heteromeric [admixture of plant-part types and sizes; Krasilov 1975] hydrodynamically equivalent assemblages in barform troughs or abandoned channels), and (2) lakes where the depth-to-bottom exceeds the chemical and biological depth of reactivity (Fig. 7.3). Pedogenesis across the interfluvial and along river-and-lake margins (aquatic and semi-aquatic colonization)

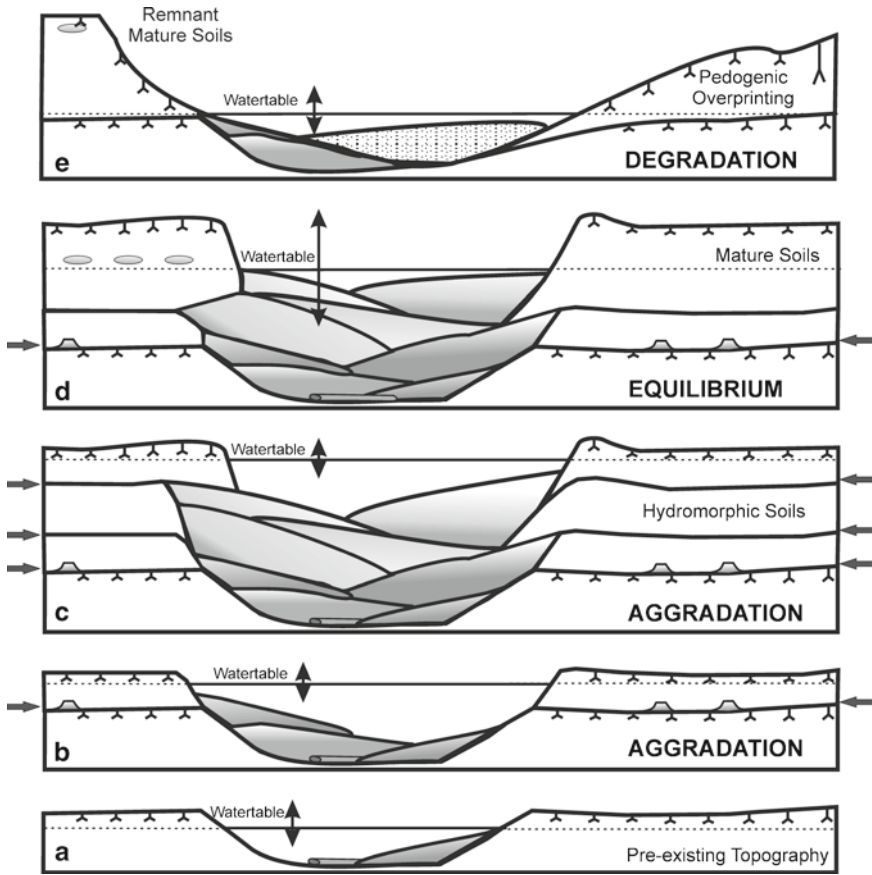


Fig. 7.3 A model for preservation of plant assemblages in continental regimes. See text for details

promotes decay and carbon recycling, preventing preservation of all but the most resistant aerial plant parts (e.g., palynofacies debris, palynomorphs; Gastaldo et al. 1996a). Evidence for subterranean rooting structures may be preserved depending upon the type of vegetation or soil, and the reactivity of pore waters with entombed organic material (e.g., drab halos, rhizoconcretions, etc.; Retallack 2001).

The onset of fluvial disequilibrium and landscape aggradation promotes sediment accumulation within the interfluves that, in turn, may bury organic litters once residing at the soil–air interface or shallow subsurface. As incremental bedload deposits accumulate within channels in response to discharge rates that are lower than needed to flush the system, CWD lags may occur within various parts of barforms (Gastaldo 2004). Incremental deposition of siliciclastics also occurs along channel margins (levees) which respond to overbank sedimentation or avulsion with an increase in their vertical height and lateral extent. The change in relative position of the river’s water surface is reflected in the watertable across the interfluve, and

it ascends in the section in response to sediment buildup (Fig. 7.3b, c). As the watertable continues to reestablish at a higher stratigraphic level in response to aggradation, buried plant debris is maintained within anoxic geochemical conditions that prevent decay past the state of the original organic matter when entombed. Organic matter that accumulates at the soil–air interface within the new (primarily wetland) landscape still is subjected to pedogenic activity and carbon recycling.

A phase of continuous aggradation without landscape degradation, accompanied by a stratigraphic rise in interfluvial watertable, will result in a sequence of stacked fine-grained clastic deposits wherein fossil-plant assemblages will be confined to the basal unit(s) immediately above the disconformity (previous soil horizon) if geochemical conditions are met for potential preservation. Where semi-continuous aggradation occurs without landscape degradation, the possibility exists, but rarely is met, for fossil-plant assemblages to be preserved at the contact between each soil horizon and the overlying aggradational, fining upwards sequence if subsequent aggradational intervals accumulate rapidly (Fig. 7.3). These plant assemblages will be preserved primarily as adpressions (coalified compressions and impressions) with carbon, lipid, resin (if present systematically), and cuticular residua. Depending upon the timing between soft-tissue decay and the rate of sediment influx and final entombment, casts of axes (e.g., pith casts) also may be found (Allen and Gastaldo 2006).

Once equilibrium is reestablished across the landscape, pedogenesis again dominates the interfluvial (Fig. 7.3d). Depending upon the prevailing climate and associated distribution of rainfall over the year, deep soil alteration can occur, buried organic matter recycled, and pedogenic concretions may form. Under more arid or highly seasonal and restricted precipitation regimes, stable-isotope signatures sequestered in carbonate concretions will reflect atmospheric gas concentrations because of formation within a geochemically open system (Tabor et al. 2007). But, carbonate concretions also may form in geochemically closed systems, wherein stable-isotope signatures reflect bacterially-mediated decay of organic matter within saturated soils (Tabor et al. 2007). Hence, the presence, alone, of carbonate concretions at some depth within a paleosol may not be the sole criterion for interpreting prevailing climate at the time of landscape equilibrium and, hence, soil formation (Gastaldo and Rolerson 2008). But, as long as entombed plant-part assemblages are maintained below the stratigraphic level of deepest watertable penetration, their preservation potential remains high (Fig. 7.3).

The onset of fluvial disequilibrium resulting in landscape degradation, either through tectonic uplift or tilting, increased river discharge and/or decreased sediment supply associated with a shift in climate, or base level change, leads to valley cutting and scavenging of the former landscape due to the downward shift in equilibrium profile. The removal of strata and establishment of a new base profile, resets the regional watertable (Fig. 7.3e). The outcome of such a regional watertable reset exposes any previously buried plant assemblages to renewed decay processes if they are positioned now (or at any time) in the vadose zone. Depending upon the original grain-size, mineralogical parameters, and degree of compaction of the entombing sediment, a number of different plant-part preservational modes may result. For example, if non-woody plant parts originally were entombed within fluvial sand,

sediment porosity and permeability will promote complete deterioration of the most labile organic matrix. The only remnants that may be left behind would be unrecognizable or unidentifiable debris and “poorly” preserved axial remains, which could be assigned to a major plant group if distinguishing morphological characters exist (e.g., sphenopsid axes). In contrast, if the confining sediments were fine-grained in nature, decay will result in loss of the organic matrix while retaining the overall impression of the features of the buried plant part in the mudrock. Impressions will be the result of sediment and organic-matter compaction within the confining pressures of the matrix (Rex and Chaloner 1983). Plant assemblages that remain at a stratigraphic depth below the new watertable have the potential for continued retention in the stratigraphic record, unless landscape degradation in the future again resets the level of regional watertable. Preservation potential for plant parts again increases once aggradational processes begin to fill these incised landscapes, usually resulting in wetland assemblages (Demko et al. 1998).

In summary, plant-fossil assemblages in continental regimes have the highest preservation potential when there is a turnover from landscape degradational to landscape aggradational processes. In systems where aggradation is continuous, the highest preservation potential exists at the contact above the disconformity between the degradational and aggradational landscapes and within standing bodies of water. There are decreasing probabilities for the preservation of plant debris higher in these sections depending upon the A/S ratio in effect at the time of plant-part entry into an aquatic (fluvial or limnic) environment. In systems where aggradation is semi-continuous without landscape degradation, the highest preservation potential exists at the contact between fining upwards intervals, each of which can be considered analogous to a parasequence in marginal-marine settings. In this case, instead of the genetically related strata being bounded by marine flooding surfaces, they are bounded by surfaces that mark an increase in relative accommodation rate filled nearly instantaneously by fluvial and overbank aggradation. As a new, long-term watertable is established at the top of the stratigraphic section in response to the buildup of the landscape, plant assemblages maintained stratigraphically within saturated sediments are buffered from decay and loss from the potential fossil record. When fluvial equilibrium and landscape stasis are attained, thick mature soils form that can be used as chronostratigraphic marker beds across the region.

5 Case Studies

The following case studies are provided as examples illustrating the basic premises between plant-part preservational modes within the landscape model context. By no means do these cover the full spectrum of possible scenarios over space and time. Rather, each sedimentary basin within the range of various tectonic settings must be evaluated independently to determine the relationships between presence or absence of fossiliferous plant beds within stratigraphic and regional context.

5.1 *Plant Assemblages in Aggradational/Degradational Landscapes*

5.1.1 Paleogene Weißelster Basin, Central Europe

The Paleogene Weißelster basin, Germany, is comprised of interfingering continental and marine sequences, and is paleogeographically in close proximity to the paleo-North Sea (Kruttsch 1992). The Eocene and Oligocene deposits were subdivided by Eissmann (1970) into three stratigraphic units. The Borna beds are predominantly terrestrial and of Middle and Late Eocene in age, the Middle Oligocene Böhlen beds are predominantly brackish-marine, whereas the overlying Middle Oligocene Bitterfelder beds are indicative of terrestrial to brackish-marine settings. The fluvial Borna beds (Halfar et al. 1998) consist of coarse sand and gravel, with intercalated thick clay deposits representing oxbow lakes and wetland paleosols. Several thick economic coals occur within the Eocene sequence, in addition to several well preserved fossil-leaf assemblages that have been used for paleoecologic (Gastaldo et al. 1998) and paleoclimatic (e.g., Mosbrugger et al. 2005) restorations of central Europe.

In the Bockwitz mine near Borna, Germany, a Middle Oligocene flora is preserved in close proximity to a Late Eocene coal – Oberflöz II – and an overlying interfluvial paleosol (Gastaldo et al. 1998; Fig. 7.4), which stratigraphically correlates with sediments below the Late Eocene Oberflöz III coal in the Schleenhain mine several kilometers to the west (Halfar et al. 1998). Taphonomic criteria (Gastaldo et al. 1996b) allow for the recognition of autochthonous floodplain assemblages, parautochthonous channel-fill sequences, and allochthonous accumulations either of coarse woody detritus (Gastaldo 2004) or heteromeric assemblages within trough fills of channel barforms.

Three autochthonous plant assemblages were exposed in the mine highwall in the early 1990s; more recently, mine reclamation has left little surface exposure of any plant-bearing intervals. Besides the lignite bed (histosol of Oberflöz II) of varying regional thickness, two additional paleosols occur in the stratigraphic sequence. The first is found terminating the overlying fluvial complex, and represents the culmination of alluvial plain aggradation following peat cessation. This Late Eocene paleosol is only a remnant of a laterally extensive, well-developed soil that attained a thickness of at least 1.7 m. A prominent A-horizon is underlain by a weathered and stained B interval, beneath which was a well-developed C zone (Gastaldo et al. 1998, Fig. 7.4). In situ woody tree bases are preserved in the A-horizon, with extensive and deeply penetrating roots found in the B and C horizons. Clay mineralogy and major element analyses show little variance. The second paleosol is a very thick, at least up to 17 m or more (due to missing section at ground level), kaolinitic massive and homogenized silt of Middle Oligocene age. Large diameter woody roots penetrate deeply throughout the interval, accompanied by slickensides and other features that were used to interpret pedogenesis of a vertisol. This soil formed across a middle Oligocene landscape characterized by

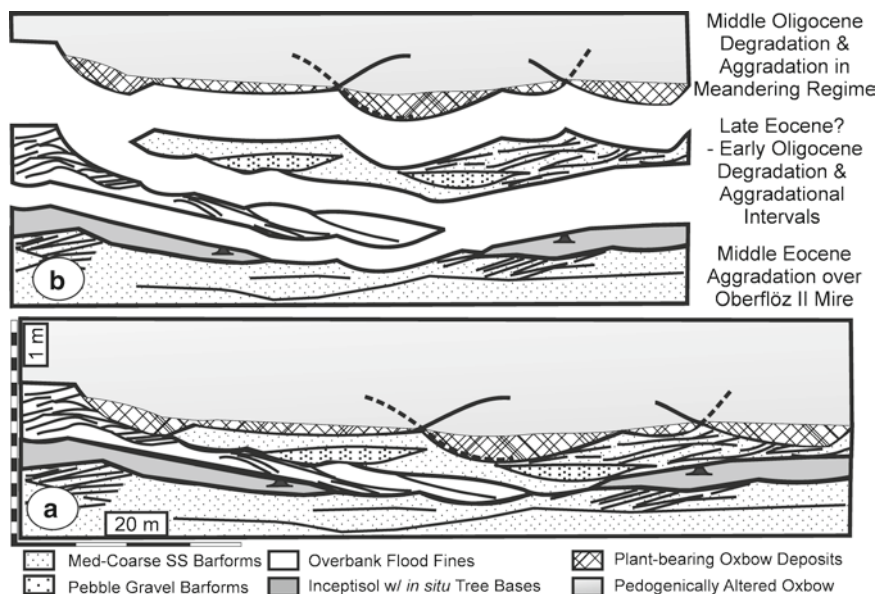


Fig. 7.4 An example of the effects of an aggradational/degradational landscape on plant-fossil assemblages in Eocene-Oligocene fluvial deposits of the Weißeelster Basin, central Europe. (a) Illustration of the mine highwall based on photomosaic taken in the Bockwitz mine near Borna, Germany (Gastaldo et al. 1998). The stratigraphic sequence, exposed above a thick Eocene coal (Oberflöz II), consists of aspects of both aggradational and degradational landscapes that encompass approximately 10 MY. (b) Aggradational sequences and boundaries marking the base of degradational phases in chronostratigraphic context are isolated in a Wheeler-type diagram depicting the relationship between the type of plant-fossil assemblage and these events. Erect, in situ, woody trees and an organic (O) horizon are preserved at the top of a Middle Eocene landscape, with much of this landform removed by subsequent degradation. Exceptionally well preserved leaf assemblages occur only in the base of Middle Oligocene oxbow lake deposits, the lowest meters of which were not subjected to pedogenic overprinting

the aggradation of one or more meandering fluvial regimes, each of which were abandoned and filled as oxbow lakes. Pedogenic alteration was subsequent to channel fill.

Identifiable parautochthonous plant assemblages only are restricted to the lowermost 1.5 m of Oligocene channel-fill sequences, although these probably were more extensive (Gastaldo et al. 1996b) prior to pedogenic overprinting (Fig. 7.4). The aquatic fern *Salvinia* is commonly interbedded with aerial leaf, branch, fruit-and-seed remains, representing autochthonous elements within this parautochthonous assemblage. Leaf accumulations occur as isolated leaves or clusters of leaves (mats) on bedding planes without any preferred abiotic orientation, either solely as impressions or inclusive of cuticular remains.

Allochthonous assemblages are restricted to channel lags as CWD above scour contacts (Gastaldo et al. 1998; Gastaldo 2004) and within channel barforms as admixtures of leaf, fruit, and seed debris. Due to the grain size of these channel deposits,

ranging from medium and coarse sand to pebble conglomerate, non-woody detritus is restricted to troughs of bedforms where most exhibit a very low quality of preservation.

The quality, quantity, and stratigraphic position of these assemblages in the Bockwitz mine are consistent within the context of the proposed preservation model. Autochthonous assemblages that include erect woody trees, deeply penetrating rooting systems, and differentiated soil horizons, mark intervals of landscape stasis and equilibrium of the fluvial regime. Phases of disequilibrium occurred when river avulsion or fluvial reactivation occurred in response to Alpine tectonic activity, climate change, or both. This resulted in the burial of paleosols and overall continental aggradation in the Late Eocene (Halfar et al. 1998). Oxbow lake deposits and the plant-fossil assemblages preserved therein were coeval with in-channel deposits of meandering fluvial systems elsewhere in the region.

The stratigraphic hiatus recognized between the Late Eocene and Middle Oligocene is a function of disequilibrium that resulted in landscape degradation. The first preserved degradation event (Fig. 7.4) emplaced fluvially derived fine-grained sediment above an erosional contact that removed the latest Eocene deposits (including Oberflöz III) and much of the Eocene autochthonous woody assemblage. Although the base of the channel form that marks the erosional contact occurs at places below that paleosol, the regional watertable was maintained higher in the landscape profile. A high regional watertable prevented decay and degradation of these buried trees, forest-floor litter, and subterranean rooting structures. If regional watertable had dropped at any time below the stratigraphic position of this paleosol, organic matter decay and pedogenesis would have proceeded in the same manner as that found in the overlying Middle Oligocene.

The middle Oligocene experienced renewed aggradation in the landscape, with the emplacement of at least two stratigraphically stacked fluvial systems (Fig. 7.4). Plant-fossil assemblages in coarse fluvial bedforms reflect the persistence of refractory CWD in the absence of other plant detritus, which underwent decay in oxygen-rich waters and/or exposure of bedforms during intervals of low discharge. Abandoned trunk channels filled with a mixture of fine-grained clastics and aerial plant parts derived from channel margin, riparian communities (Gastaldo et al. 1989, 1996b). Stratigraphic sequences in modern oxbow lakes show a pattern of interbedded intervals of leaf clusters (mats), representing annual leaf fall, and mud, deposited during overbank flood events from the base of the abandoned channel to the top of the channel fill. The Bockwitz fossil assemblage, though, is restricted to the basalmost 1.5 m of section because of channel-fill sequence overprinting by pedogenetic activity. The depth of pedogenetic activity is marked by the contact between the well preserved plant-fossil assemblages and the homogenized, kaolinite-rich and root penetrated paleosol. Hence, the regional watertable at the time of paleosol development dropped to within 1.5 m of the contact between the bottom of the oxbow-lake channel and the underlying fluvial barforms. If regional watertable had dropped below that level, there would have been no evidence of megafloreal remains at this stratigraphic horizon.

5.1.2 Upper Triassic Chinle Formation, Southwestern United States

The Chinle Formation in the Colorado Plateau region of the southwestern United States was deposited in a fully continental basin along the western margin of equatorial Pangaea (Dubiel 1994). The bulk of the Chinle Formation consists of a succession of fluvial, lacustrine, and minor aeolian strata that fill a dynamic basin subsiding between the old Permo-Carboniferous Ancestral Rockies Uplift to the east and the Mogollon Highlands continental volcanic arc to the south (Stewart et al. 1986), and separated from the Luning marine basin to the west by a back-arc bulge (Lawton 1994). The lower part of the Chinle is dominated by volcanic-sourced smectitic mudstone and mineralogically immature sandstone and conglomerate with gleysol- and aflisol-type paleosols, while the upper part has more detrital clay, aeolian sandstone with calcisol-, and aridosol-type paleosols (Dubiel et al. 1991, Demko et al. 1998). Paleosol types, paleobiologic and taphonomic data indicate that the paleoclimate during Chinle deposition was highly seasonal with respect to precipitation, characterized as megamonsoonal by Dubiel et al. (1991).

Plant fossils are locally abundant in the lower part of the formation, including the bulk of the described assemblages of adpressions and the cuticular remains of gymnosperms, cycadophytes, ferns, and other plant types (Daugherty 1941; Ash 1970, 1980) and the justly famous permineralized gymnosperm log assemblages (e.g., Petrified Forest National Park – Ash and Creber 2000; Creber and Ash 2004). The upper part of the formation is characterized by rare fossil plant occurrences, mostly all impressions of robust foliar material (Ash 1987).

A locality in Petrified Forest National Park in Arizona illustrates the relationship between plant preservation, paleosol type, and timing of landscape degradation and aggradation in the lower part of the Chinle Formation (Fig. 7.5). This locality, within the Tepees area of the central portion of the park, has produced a significant

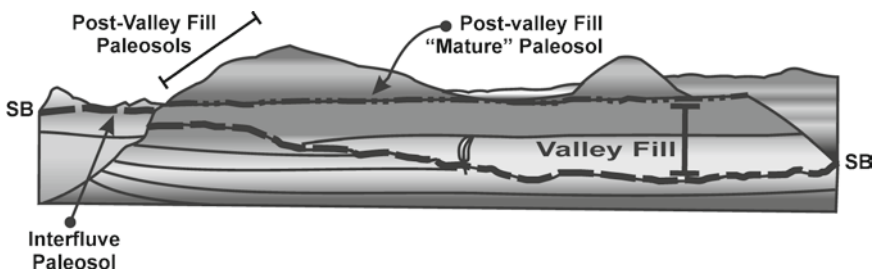


Fig. 7.5 Strata of the Chinle Formation at the Tepees locality, Petrified Forest National Park, Arizona, were deposited during two periods of fluvial aggradation separated by an episode of landscape degradation, resulting in a 12–13 m deep valley cut into the underlying aggradational succession. Fossil plants are preserved in the initial phases of landscape aggradation that overlap the paleovalley walls. Fossil-plant abundance in the lower Chinle, when compared to a sparse record in the upper part of the formation, is the result of deposition within aggrading fluvial and overbank systems confined to paleovalleys under a hydrologic setting characterized by perennially high water tables. SB = sequence boundary

number of the fossil plants that have been described from the Chinle (Ash 1972, 1991, 2001). These strata were deposited by two periods of fluvial aggradation separated by an episode of landscape degradation that cut a valley 12–13 m deep into the first aggradational succession (Demko 1995a). The entire succession at the locality can be divided into three units characterized by different depositional facies, paleosol types, and fossil-plant taphocoenoses: (1) an initial aggradational succession characterized by drab purplish-gray floodplain mudstones with discontinuous coarse-grained sandstone channel-fills; (2) an aggradational valley-fill succession comprised of light greenish-gray mudstone, and light red and brown siltstone with climbing-ripple cross-laminated very fine-grained sandstone, and lateral-accretion beds of fine- to medium grained current ripple cross-laminated and trough cross-bedded sandstone; and (3) a post valley-fill succession of dark reddish brown floodplain mudstones with moderately- to well-developed calcareous and vertic paleosols (Demko 1995a). The light greenish-gray mudstone within the aggradational valley-fill succession preserve abundant parautochthonous and autochthonous cycadophyte, fern, and other leaf material in distinct litter layers at bed boundaries. Autochthonous coalified gymnosperm logs are preserved in the base of, and erect, casted in situ *Neocalamites* trunks and leaf mats are preserved in the upper parts of time-equivalent lateral-accretion units.

The fossil plants at the Tepees locality were preserved in the initial phases of landscape aggradation after a period of fluvial incision and degradation. The purplish-gray floodplain mudstones and coarse-grained sandstones below the surface of incision, a sequence boundary, were deposited by aggradation of an avulsion-dominated fluvial system (Demko 1995a). The floodplain mudstones are characterized by pedogenic features such as drab root haloes and color mottles that indicate groundwater gleyization and fluctuating watertables associated with the aggradational succession. No fossil plants are preserved in either the partially-filled sandstone channel deposits or their associated levee and floodplain deposits. The surface of incision cuts 12–13 m into these deposits. At the base of this valley cut, this degradation surface is overlain by lateral-accretion beds of the Newspaper Rock sandstone and associated levee and overbank wetland deposits. These units onlap the valley walls and preserve the fossil plants noted above. The overbank wetland deposits are characterized by meter-scale coarsening-upwards units comprised of greenish-gray mudstone, siltstone, and climbing-ripple cross-laminated fine-grained sandstone. In areas near the facies change into levee and lateral-accretion beds, the tops of the coarsening-upwards units are pedogenically modified and have a slight reddish-gray coloration. At interfluvial areas along the margins of the valley cut, a well-developed dark reddish brown calcareous paleosol is developed on the underlying pre-incision units. A dark red calcareous vertic paleosol is developed over the valley fill and marks the overlap of the incisional topography, a return to unconfined fluvial and overbank conditions, and an apparent slowing of depositional rate.

As discussed by Demko et al. (1998), the distribution of coalified compression-impressions (adpressions) and preserved cuticle of foliar plant material in the lower part of the Chinle Formation was controlled by sedimentological and near-surface

geochemical conditions inherent to depositional settings within incised valley-fill successions. The abundance of plant fossils in the lower Chinle, as compared to their rarity in the upper Chinle, is the result of deposition within aggrading fluvial and overbank systems confined to paleovalleys. Evidence from lower Chinle paleosols and trace fossils in areas outside the valley-fills indicates a climatic and hydrologic setting characterized by seasonal fluctuations in recharge and groundwater table (e.g., Dubiel et al. 1991; Hasiotis and Mitchell 1993). However, during aggradation of the valley-fill succession, watertables in these deposits were perennially high, as they are in modern aggrading alluvial valleys (e.g., Gallaher and Price 1966; Runkle 1985). Burial of plant material within depositional environments in these paleovalleys (fluvial channels, crevasse splays, lakes, wetlands, and lacustrine deltas; Demko 1995b) increased the preservation potential because reducing conditions and low biological demand in these areas of high watertables (near and at the surface), and groundwater discharge were conducive to low rates of decay. Confined fluvial systems also have a relatively small area in which to store and deposit aggrading material, contributing to comparatively high rates of vertical aggradation (Posamentier and Allen 1999). Preservation potential for foliar material was dramatically lower in areas outside the paleovalleys during deposition of the lower Chinle, and in upper Chinle strata overall, conditions of fluctuating and low water tables, associated with both fluvial depositional styles (e.g., avulsion-dominated and ephemeral streams) and increasing aridification. Only rare robust (coriaceous) leaves and stems in areas outside channels, and permineralized wood in channels filled with, or laterally-adjacent to, facies with abundant volcanogenic material, can be found (Demko et al. 1998).

5.1.3 Lower Triassic Katberg Formation, South Africa

The Katberg Formation in the Karoo Basin represents deposition within Early Triassic fluvial aggradational and degradational successions (Pace et al. 2009) within a fully continental basin. These rocks are assigned to the Tarkastad Subgroup of the Beaufort Group (Johnson et al. 1997), and were sourced by sediments transported several hundred kilometers from the Cape Fold Belt (Smith et al. 1993). The distance from the provenance and overall grade resulted in fluvial regimes dominated by very fine, feldspathic sand to silt, with overbank interfluves consisting of silt and thin sheet sand bodies. Pedogenic carbonate pisoliths and intraformational mudclasts (Smith and Ward 2001) may be concentrated as lags at the base of each erosional contact marking landscape degradation. Hiller and Stravrakis (1984) interpreted the sandstone bodies as wide, shallow channels of a shifting braided pattern, while Retallack et al. (2003) ascribe several climate attributes to a variety of paleosols across the landscape. Recently, Gastaldo and Rolerson (2008) were unable to apply the criteria proposed by Retallack et al. (2003) to paleosols exhibiting bioturbation and pedogenic nodules, resulting in their recognition of more wetland paleosols in this part of the stratigraphy than previously interpreted.

Plant-fossil assemblages in this part of the sequence are poorly preserved and, in many instances, represent impressions of the most resistant parts including sphenopsid and unidentifiable axial remains. Gastaldo et al. (2005) report new collections from bedform drapes, scour-and-fill structures, and troughs within channel barforms, and from within overbank deposits in aggradational sequences above paleosols. Collections from Bethulie and Carlton Heights, to date, have been preserved in fine- to very fine sandstone matrices.

The Katberg Sandstone exposed along the N9 highway at Carlton Heights (Fig. 7.6) will serve to illustrate the relationships between plant-assemblage preservation and the role of landscape aggradation and degradation. A sequence of stacked wetland paleosols (Gastaldo and Rolerson 2008) overlie each principal fluvial sandstone body, where not removed through subsequent erosion. Each paleosol is comprised of silt-sized clasts. Thin, very fine-grained and rippled sheet sandstone bodies are interbedded with the siltstone, along with isolated lenses of pedogenic carbonate nodule lags. It is within one of the sandstone bodies that scattered, poorly preserved plant remains occur in a megaripple trough fill. There is no apparent order to the assemblage, with most fragmentary debris on the order of 5 cm in maximum dimension. Impressions of sphenopsid axes, dispersed fern pinnules (identified due to the vascular tissue having been accentuated by iron staining), and reproductive structures are dispersed in the matrix. The poor state of preservation, with resistant plant parts remaining solely as impressions, indicates that all soft tissues were degraded.

The timing when these plant parts underwent degradation is important to interpret the preservational attributes of the assemblage. The plant parts are allochthonous, being emplaced within the bedform trough via suspension-load settling. There is no evidence for physical abrasion of these parts that would be encountered during bedload transport (Gastaldo 1994). If the regional watertable had dropped following deposition, the most delicate plants (i.e., fern pinnules) would have

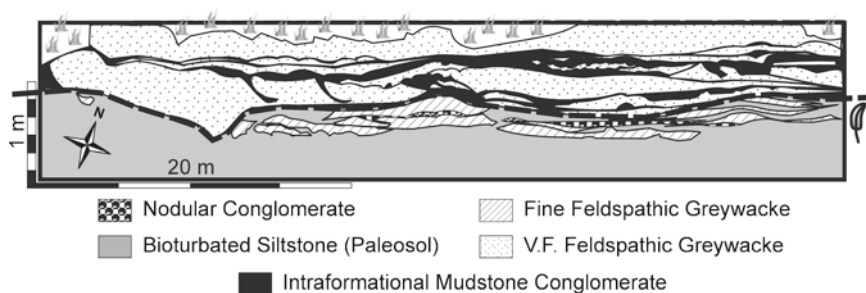


Fig. 7.6 An Early Triassic example of the effects of aggradational and degradational landscapes on plant-fossil assemblages in the Karoo Basin, South Africa. Only poorly preserved impressions of conducting tissues of fern pinnules, sphenopsid axes, and isolated reproductive structures remain in overbank interfluvial deposits of fine-grained sandstone that were not removed by subsequent landscape degradation (*dashed black line*). The plant debris remained beneath the watertable established following landscape degradation, but was subjected to oxygenated pore waters resulting in loss of the most labile tissues

decayed and been removed from the record without any evidence of having been there. The development of impressions of isolated plant parts in the sediment is not possible immediately following burial, due to the maintenance of pore-water pressures in the matrix. Some degree of compaction and dewatering must occur before the outlines of plant parts are imparted to the entombing matrix. Hence, plant-tissue decay occurred subsequent to continued burial and early compaction.

The Lower Triassic Katberg Formation is characterized by marked degradational events that removed significant portions of the stratigraphic record (Pace et al. 2009). Landscape degradation in the Katberg exposure at Carlton Heights resulted in the removal of the paleosol and overbank sheet sandstones in the north of the outcrop (Fig. 7.6). A multi-storied sand body exhibiting a braided configuration was emplaced above the erosional contact. Stratigraphically, the preserved plant-fossil assemblage occurs just above the lateral equivalent of this channel base. Landscape degradation and the development of a fluctuating watertable in adjacent sediments promoted decay of any previously buried organic matter (paleosol TOC values are very low, $\leq 0.5\%$; Gastaldo and Rolerson 2008). But, because of prior early compaction of the floodplain deposits, impressions of the most resilient plant tissues remained in the fine-grained sand when volatile tissues were lost. It is probable that subsequently introduced pore-water chemistries promoted diagenetic staining of the remaining structural tissues, accentuating their morphologies. The near absence of plant-fossil assemblages in the Early Triassic of South Africa can be explained as the result of changes in regional watertable in response to landscape-degradation events in the basin. Hence, the paucity of a plant-fossil record in this case is a taphonomic artifact of the basin (Gastaldo et al. 2005) rather than the absence of a terrestrial flora in response to terrestrial extinction (Ward et al. 2000).

5.2 *Plant Assemblages in Aggradational Landscapes*

5.2.1 **Eocene Willwood Formation, Western United States**

The Bighorn Basin in Wyoming, USA, is a large intermontane basin encompassing approximately 10,000 mi² of dominantly Cretaceous rock exposures in addition to localized outcrops of Early Mesozoic and Tertiary successions. Part of the Lower Eocene stratigraphy is assigned to the Willwood Formation, characterized by alluvial deposits consisting of vertically stacked floodplain mudrock (moderate to strongly developed paleosols), thick, multi-storied sheet sandstone bodies (meandering trunk channels), and heterolithic deposits consisting of ribbon sandstone surrounded by pedogenically altered mudrock (avulsion deposits associated with weakly developed paleosols; Kraus 1996, 2001; Fig. 7.7). Poorly developed paleosols within their heterolithic facies indicate high sedimentation rates controlled by avulsion (Kraus and Gwinn 1997). In contrast, better developed and more mature soils are indicative of low sedimentation rates on floodplains controlled by overbank deposition (Kraus 2002). A gleyed, drab-colored hydromorphic paleosol

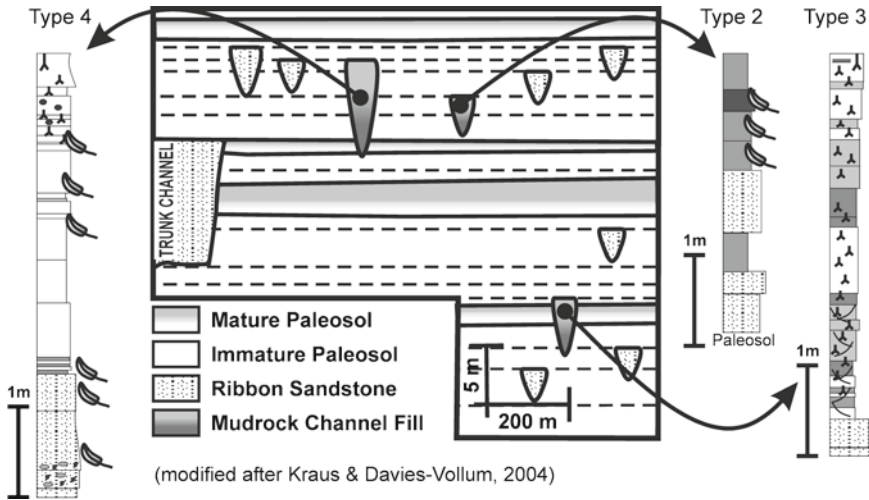


Fig. 7.7 An Eocene example of aggradational landscapes in the Willwood Formation, Wyoming, USA. Stacked poorly developed (immature) paleosols occur throughout the sequence, with interspersed better developed (more mature) paleosols. Fossil plant assemblages are most common within siltstone channels interpreted as abandoned avulsion channels within the floodplain. Kraus and Davies-Vollum (2004) recognize four different channel-fill types. Channel-fill Types 2 and 4 may contain well-preserved leaf mats and show the least effects of pedogenic activity. Channel Types 1 (not illustrated) and 3 rarely preserve plant debris due to pedogenic alteration of sediments

underlies the carbonaceous shale which, in turn, is overlain by the development of a yellow-brown paleosol in the former sequence (Davies-Vollum and Kraus 2001). The better developed soil types range from vertisols to Fersiallitic paleosols (no direct equivalent in the Soil Survey Staff 2006; Kraus 2002). Spatial variation in the more developed paleosols are suspected to be indicative of variations in regional drainage which are linked to grain size and, hence, permeability.

Plant-fossil assemblages are common in the carbonaceous shale facies, and have been used to reconstruct not only the paleoecological mosaic but also the paleoclimate of the Early Eocene (Davies-Vollum and Wing 1998). Carbonaceous shale bodies exhibit both tabular and lenticular geometries (Wing 1984), and the plant accumulations within display the entire spectrum from exceptionally well preserved (Davies-Vollum and Wing 1998) to very poorly preserved assemblages (Kraus and Davies-Vollum 2004). The best preserved material occurs in two facies. These are mudrock that are relatively low in total organic carbon and unaltered by soil-forming processes, and carbonaceous shales that also are the least pedogenically altered. Both facies are interpreted as overbank-wetland deposits (Davies-Vollum and Wing 1998; Davies-Vollum and Kraus 2001) and, more recently, lenticular bodies have been recognized as channel-fill sequences associated with avulsion events. Where plant debris is preserved in carbonaceous shale of a tabular nature, Davies-Vollum and Kraus (2001) interpreted the overlying paleosols as

having formed on the floodplain during avulsions. The underlying carbonaceous shale, then, represents forest-floor (in situ) accumulations within the distal floodplain. They note that conditions promoting organic-matter accumulation in this setting require a high water table, at or above the soil surface. Such a state results in waterlogged, anoxic conditions that may prohibit bioturbation and inhibit decay of organic material. Hence, the maintenance of a high local (or regional) water table during the initiation of each aggradational phase, associated with its rise in the section during the emplacement of stacked soil horizons, resulted in preservation of backswamp (autochthonous) debris.

Parautochthonous assemblages are found within mudrock facies interpreted as abandoned crevasse-splay feeder channels (Kraus and Davies-Vollum 2004). These often are overprinted by pedogenic activity in the later stages of, or following, channel fill, as plant colonization occurs and the most recent deposits were amalgamated into the floodplain (Wing 1984). Based on what can be determined from the literature, there does not seem to be evidence for the preservation of coeval litter (O) horizons at the soil-air interface equivalent with channel-fills. Four different categories of channel-fill successions are recognized by Kraus and Davies-Vollum (2004) in which plant part preservation varies. Poorly preserved plant assemblages are found in Types 1 and 3 fills (Fig. 7.7). In the former, decay of organic matter is attributed to the presence of circulating oxygenated water and low sedimentation rates, whereas pedogenesis overprints Type 3 fills and eliminates any buried organic matter. Type 3 fills seemingly are associated more commonly with more mature paleosol horizons (Kraus and Hasiotis 2006), resulting in the greater depth of pedogenic alteration and soil modification. The depth of root penetration and soil modification ultimately are controlled by variations in local watertable. Type 2 fills (Fig. 7.7) contain lenticular carbonaceous shale that preserves leaf mats as well as dispersed plant debris. The presence of organic-rich mudrock reflects slower sedimentation rates and a change in prevailing Eh/pH conditions promoting preservation (Gastaldo and Huc 1992). Lastly, Type 4 fills are characterized by conglomerate or sandstone basal deposits with allochthonous plant debris, and intervals of interbedded mudstone and siltstone in which moderate-to-well preserved parautochthonous fossils occur. Kraus and Davies-Vollum (2004) attribute the variety of fill characteristics to the differences in local drainage conditions that followed channel abandonment, as well as how rapidly the channel was filled. Their assemblages display similar taphonomic attributes to those described by Gastaldo et al. (1998) in Germany (see above). Hence, within the aggradational landscape of the Eocene Willwood Formation, fossil-plant assemblages have the highest preservation potential when they are maintained below the prevailing watertable, either in the short term (aggradational phase characterized by poorly developed paleosols) or longer term (the phase during which more mature soils developed). Once the local/regional watertable fell below the level of buried plant detritus in Types 1 and 3 channel fills, pedogenic activity removed most, if not all traces of the remains.

The type and nature of rhizoliths preserved within these aggradational paleosols recently have been used to identify differences in the paleodrainage of the landscape. Kraus and Hasiotis (2006) provide criteria to separate those soil (more

mature) types that formed under relatively well-drained from those formed under poorly drained (immature) conditions. Iron (hematite) and manganese staining adjacent to rhizoliths, as well as the presence of calcareous rhizoconcretions are considered to be indicative of better drained soils (albeit without stable isotope geochemical constraint), whereas rhizoliths in poorly drained paleosols commonly are surrounded by goethite rims and may be preserved by jarosite (pyrite oxidation product). Fluxes in the geochemical conditions of the soil, in response to climate, result in the variety of preservational states of rooting structures as well as previously buried litters.

5.2.2 Upper Jurassic Morrison Formation, Western United States

The Morrison Formation was deposited in a broad, continental, retroarc basin in western North America during the initial stages of collision- and subduction-related tectonic mountain building processes that eventually culminated in the Sevier orogeny (DeCelles 2004). Demko et al. (2004) interpret the Morrison to represent two aggradational sequences with an intervening degradational episode. The upper and lower boundaries, and the medial bounding surface separating the two aggradational successions, were interpreted as basinwide sequence-bounding unconformities. These were generated by changes in dynamic uplift and subsidence patterns due to hinterland tectonics and deep crust-mantle reorganization. The initial aggradational episode in the Colorado Plateau region is represented by the restricted-marine to lacustrine facies of the Tidwell Member and the overlying thick package of low-sinuosity fluvial sandstones and associated floodplain deposits of the Salt Wash Member. Overlying a surface marked by a widespread paleosol and minor incision, the second aggradational succession is represented by the mudstone-prone fluvial and lacustrine facies of the Brushy Basin Member. Paleosols, trace fossils, and other sedimentary paleoclimatic proxies indicate that the Morrison was deposited under arid to tropical wet-dry conditions (Demko and Parrish 1998; Demko et al. 2004; Hasiotis 2004; Turner and Peterson 2004).

Even though outcrop of the Morrison Formation covers a very large area of the Western Interior of the United States and is, in most cases, very well exposed in the modern semi-arid climate, plant fossils are typically very rare, with the exception of the very uppermost beds in the northern reaches of the outcrop belt in a coal and organic-rich mudstone interval (Parrish et al. 2004). However, traces of roots, moderately well-developed paleosols, a diverse and abundant soil ichnofauna associated with socialized insects, and the abundant and diverse remains of large herbivorous saurpoid dinosaurs for which the formation is famous, indicate that the landscape supported extensive vegetation (Demko et al. 2004; Hasiotis 2004; Engelmann et al. 2004). The plant fossils that do occur include coniferophytes, ginkophytes, cycadophytes, ferns, and other plants (Tidwell 1990; Parrish et al. 2004).

In a taphonomic study of fossil plants in the Morrison, Parrish et al. (2004) concluded that within the bulk of the formation, surface and near-surface conditions during deposition were such that even small fragments of carbonaceous debris were

not preserved. They attribute this relative paucity of identifiable plant remains to oxidation in well-drained soil conditions and the relative scarcity of durable woody vegetation on the landscape in general. The few well-preserved foliar assemblages described by Parrish et al. (2004) were associated with wetland and marginal-lacustrine depositional environments where watertable would have remained seasonally high in an otherwise well-drained landscape. However, both wetland and shallow or marginal-lacustrine settings that may have existed in overall net-evaporative climatic settings like the Morrison basin, must have been subject to regular low lake or watertable conditions within the realm of typical water budget fluctuations over tens to thousands of years, reducing the preservation potential of large volumes of organic material. Also, these types of depositional environments can only exist in the absence of high siliciclastic sedimentation rates, also reducing the potential for quick burial before decay or detritivory. Parrish et al. (2004) note that evidence suggests that the uppermost Morrison strata in the northern part of the depositional basin indicate of a less restrictive water budget, reflected not only in the type and amount of plant material preserved, but also in the types of paleosols and other paleoclimatic indicators (Fig. 7.8). They suggest that this may have been due to a lower overall temperature regime in that area at that time, rather than any relative increase in yearly precipitation as compared to the seemingly drier areas to the south.

The plant-fossil record of the Morrison Formation illustrates that even in overall aggradational settings, preservation of identifiable remains is still intimately linked to a balance between the A/S ratio and the maintenance of a perennially high watertable in the area where plants either are growing or introduced into the depositional environment. Paleosol types, ichnofossils, stable-isotope geochemistry of pedogenic and lacustrine/palustrine carbonates, and other sedimentary paleoclimatic indicators in the Morrison indicate that even though the landscape was dominantly aggrading through time, watertables were low or fluctuating (Demko et al. 2004; Dunagan and Turner 2004). Significant assemblages of identifiable plant material are rare and only present in deposits that accumulated in isolated lows on the landscape, near lake margins and floodplain ponds.

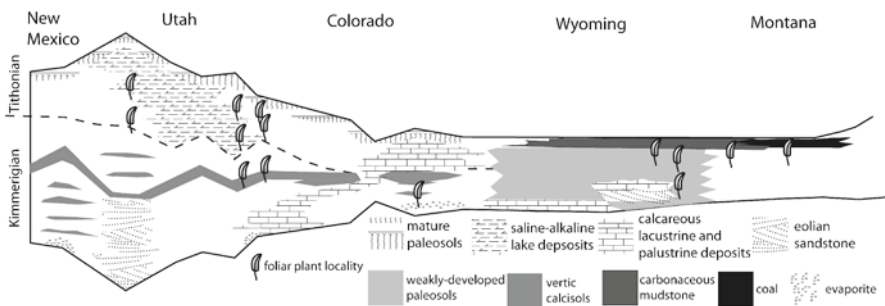


Fig. 7.8 Schematic north-south cross-section of the Morrison Formation in the western USA showing distribution of sedimentary paleoclimatic indicators and foliar fossil-plant localities (After Parrish et al. 2004)

6 Conclusions

There is no doubt that the general processes that promote preservation of a megafossil record (or any fossil record) are restricted in space and time. Based on actuopaleontological studies, it continues to be presumed that once plant detritus is introduced into an appropriate depositional setting, the chances for its preservation and retention in the stratigraphic record are high (which is not necessarily the case). This assumption is particularly true if the debris is buried rapidly, entombing it in a sedimentary package. Hence, when these environments of deposition are encountered in a basin without evidence for terrestrial plants, catastrophic perturbation often has been invoked to explain their absence. In other words, the lack of a plant-fossil record in a particular stratigraphic interval is considered to be more a function of ecosystem reorganization, extirpation, or extinction than the mere fact that prevailing conditions at the time of sediment accumulation were outside of the preservational window. In reality, without an understanding of the plant taphonomic attributes within the sedimentological and regional/basinal context, the absence of a plant-fossil record may be either a function of (1) sediment supply, attendant geochemistry, and climate preventing preservation at the onset; or (2) the interaction of landscape and climate, over the short (millennial) and long (lakh to millions of years) term, rather than extirpation or outright extinction.

Sediment deposition within continental landscapes, regardless of their geographic position, is a function of disequilibrium within the fluvial (graded) profile across the drainage system(s). Aggradational landscapes that form adjacent to river systems in disequilibrium are characterized by wetland (histosols, entisols, gleyed soils) and moderately drained soil types. Plant-fossil assemblages have the highest preservation potential in settings that are maintained below the maximum draw down of the regional watertable (e.g., channel barforms, abandoned avulsion channels, oxbow lakes, other limnic regimes, etc.). Under normal circumstances, plant debris that accumulates at the air-soil interface is recycled even when buried by overbank flood deposits due to the re-establishment of the regional watertable and exposure to oxygenating interstitial conditions. Preservation potential only increases when this O-horizon is displaced to a stratigraphic position some distance below the prevailing water table, and is maintained under geochemical conditions that retard or eliminate autocatalytic, bacterial, or detritivore activity (Fig. 7.2). Hence, the highest fidelity plant-fossil assemblages are preserved within aggradational landscapes above each disconformity (marked by a soil profile).

When fluvial systems are in equilibrium, the landscape is static providing sufficient time over which pedogenic features indicative of mature soil types form (e.g., thick O-horizons in histosols, well-developed Bk or K horizons in calcisols, etc.). Similar to aggradational landscapes, the highest preservation potential for plants exists in subaqueous environments (permanent standing bodies of water) when climate variables result in evaporation exceeding precipitation, and within peat mires/bogs when precipitation exceeds evaporation (and a perched watertable prevents or retards drainages; Gastaldo, 2010). In most instances where the landscape

is controlled by high evaporation rates under a dry or seasonally dry climate, there is an improbable chance for plant-part preservation anywhere in that landscape (although vertebrate assemblages often are preserved under these conditions). Deep penetration of pedogenic processes and drying by atmospheric gases results in organic matter disintegration. Preservation potential exists only for those materials existing below the level of deepest pedogenic penetration. Similarly, and although it may seem counterintuitive, in most instances where the landscape is controlled by high precipitation rates under a humid or everwet climate, there is a low chance for plant-part preservation in the landscape. This is because sediment that would have been available for transport out of the landscape is sequestered in soils by the rootstocks of dense vegetation. The perception of muddy rivers in Recent tropical regimes is a function of human agricultural and silvicultural activities (e.g., Staub et al. 2000).

Sediment removal within continental landscapes, again regardless of their geographic position, is a function of disequilibrium within the fluvial profile across the drainage system(s). Degradational landscapes are characterized by fluvial incision, sediment excision and re-entrainment of interfluvial deposits, and transport of these materials, along with any previously entombed organics, to a depositional site below the new base level. Fluvial incision resets the relative position of the water table in the “new” landscape in a stratigraphic position below its previous height. As such, watertable fluctuations in response to whatever prevailing climate now operating in the region will result in the re-exposure of once buried organic matter to renewed biogeochemical processes which, in effect, will remove all traces of organic residuum that may have been present originally.

The long-term fate of buried organic matter and the size of the sediment clasts in which it was entombed controls the ultimate preservational mode (Schopf 1975) of fossil-plant material collected in the field. Compressions, where a coalified residuum exists on the bedding-plane surface, are the result of plant debris buried within aggradational landscapes wherein the subfossil assemblage was maintained below the prevailing and subsequent water table(s). Devolatilization and depolymerization leave a residuum of the most refractory chemicals surrounded by matrix. The external impression (mold) of those plant parts is best developed in fine-grained matrices (clay and silt-sized clasts), whereas the fidelity of the impression decreases with an increasing clast size. When both the compression and impression of the original plant remains exist, this preservational mode has been termed an adpression (Shute and Cleal 1987). Impressions, where there is no evidence of an organic residual on the bedding plane, are the result of organic decay following early diagenetic compaction by burial. There would be no transfer of plant morphology to the sediment without a decrease in pore space within the entombing matrix. Hence, for such a transfer of morphology to exist, pore space must have been reduced through burial at some depth less than that required for lithification. But, the removal of all organic residuum from the matrix requires that active decay processes (or thermal heating in response to tectonic activity) were reestablished prior to cementation. Such processes occur when landscapes are static (equilibrium) or when the landscape has been degraded. The larger the size of the entombing clasts, the higher the probability

that once buried plant debris, now exposed to oxygenating waters under a reset watertable, will be removed from the stratigraphic record. Other preservational states of plant material, including permineralization, pyritization, replacement, etc. can be explained as a function of changing geochemical groundwater-solute concentrations and conditions, and have the potential to occur under all landscape phases.

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