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Facies Analysis and Plio-Pleistocene Paleoecology

Craig S. Feibel

Sedimentary rocks are the result of chains of complex genetic processes, including particle formation, transport, and accumulation, along with subsequent histories of in situ modification following deposition. Some processes may produce a variety of sedimentary products with only slight variation in the available components or environmental conditions. And some sedimentary products may result from one of a variety of processes or conditions. Typically, however, the suite of characteristics recognizable in sedimentary rocks provides detailed and often unambiguous clues to the processes and environmental conditions during their formation. These suites of characteristics are what enable us to recognize sedimentary facies. Analysis of associations of facies, along with their vertical and lateral relationships, allows the interpretation of sedimentary environments and the reconstruction of ancient landscapes. The ability to recreate ancient landscapes and the sequence of changes in environmental character through time is a valuable tool in ecological reconstruction from ancient records. Placing fossil or archaeological assemblages into this framework can greatly enhance our understanding of the conditions of site formation, the processes active before, during, and after assemblage, and the broader environmental setting of a particular locality.

A *facies* refers to "the sum of lithologic and paleontologic characteristics of a sedimentary rock from which its origin and the environment of its formation may be inferred" (Teichert 1958). The concept is an invaluable tool for the description and interpretation of sedimentary rocks, as it allows the easy integration of lithological characteristics, primary and secondary sedimentary structures, and pedogenic (soil) overprints. A facies is a distinctive rock type, recognizable on the basis of one or more defining characteristics, and it carries with it an interpretation of the processes known to result in that particular assemblage of characteristics. From a practical point of view we can consider facies strictly in terms of rock characters, or we can interpret them in terms of the related environmental conditions. In application, it is often more convenient to discuss the environmental interpretations that result from facies recognition, but it is essential to bear in mind that the interpretation rests upon recognition of characteristics, and that in some cases multiple processes can result in similar or identical characters. The basic units characterized as facies have fundamental significance in terms of environmental dynamics, and when considered sequentially within sedimentary packages, reflect discrete components of ancient landscapes. Using Walther's Law, that sequential facies packages reflect lateral relationships upon the landscape, the broader character of ancient landscapes can be reconstructed in considerable detail (Reading 1978).

SEDIMENTARY ROCK CHARACTERISTICS

The formation of sedimentary rocks results from the assemblage of components in a sedimentary basin or trap, or on an available surface. The components of sedimentary rocks are typically distinguished in three groups: particles (clasts or grains and their coatings), voids (empty space), and plasma (fluids filling void space or coating particles). The particles themselves are of primary interest, but in certain cases the voids (determining porosity and permeability) and plasma (often mediating cementation and coherence) can be significant. A common sedimentary analysis involves *granulometry*, the investigation of particle size (*texture*), shape (*sphericity*), and population distribution (*sorting*). Perhaps more significant, however, is the rock *fabric*, the arrangement or organization of those particles. The sorting, packing, and arrangement of particles, along with the accumulation of secondary cements, strongly controls rock properties and is highly reflective of processes involved in the history of that rock. It is also important to consider that some sedimentary particles are what may be termed *superparticles*, formed by the amalgamation of individual particulate components. This is a common process in soil development and in the behavior of very small particles (clays) or biogenic materials.

The genesis of individual sedimentary particles may proceed along a variety of pathways, and is often as complex as the rocks they assemble to form. The major groups of sedimentary particles are *detrital*, comprising weathered or fragmented precursor rocks; *chemical*, precipitated directly from a solution; *biogenic*, produced by biological activity; and *volcaniclastic*, resulting from the explosive fragmentation in volcanic settings. There are some special cases and gray areas in sedimentary particle characterization as well. Most detrital sediment is the product of weathering of silicate rocks, and is commonly termed *siliciclastic*. But carbonate rocks may also be subjected to weathering and erosion, producing detritus that is not siliciclastic in nature. The boundary between biological and chemical processes is not always distinct, and biochemical particles result. Biological processes may also produce components of naturally occurring rocks that are not mineral based, such as coals and peats, which are grouped as organic sediments. The significance of characterization at this level is that the individual components of a sedimentary rock often present major clues to the processes involved in rock formation. These primary lithologic characterizations will also be a central key in recognizing and naming sedimentary facies.

In Earth surface environments, sedimentary particles are available for removal (*erosion*), transport, and deposition. These processes may be dominated by simple influence of gravity (e.g., a rockfall), but most commonly they are mediated by fluid flow. In most sedimentary environments, water is the predominant fluid, but air can also be significant. Ice may also be important in certain settings. Two aspects of fluid flow are central to the processes by which sediments are eroded, transported, and deposited. These are the intensity of flow and its duration. At the deposition stage, these may impart to the sediment a suite of *primary depositional characteristics*, usually based on the way in which fluid dynamics interacts with granulometry to produce sediment fabric. Flow intensity is directly related to the size of particles that may be eroded and entrained, thus there is a good relationship between particle size and flow strength. Duration of flow determines how long the process has to interact with particles, and thus the degree to which it can affect certain properties. Sorting is the prime example, as a short duration flow has limited potential to separate particles based on size, shape, and density, and usually results in a poorly sorted deposit. A long-duration flow or repetitive-flow processes (e.g., oscillating waves) have greater capability to sort sediment into more uniform populations of particles. Without significant sorting, heterogeneous deposits may result with a *massive* or featureless fabric. Sorting may result in a wide variety of primary depositional structures such as graded bedding or cross-bedding. These more complex arrangements of particles are commonly informative as to flow characteristics and *bedforms*, the three-dimensional components of a depositional landscape.

Although most of the characteristics of a sedimentary rock are determined through the processes of particle genesis and deposition, postdepositional processes may impart significant additional characteristics to a sediment body, or overprint primary features. These processes are broadly classed as *diagenetic* or transformational processes. For sedimentary rocks deposited in terrestrial environments, an important stage in diagenesis is the group of early postdepositional processes, most significantly *pedogenesis* or soil formation. Postdepositional transformative processes may impart new characters to a sediment body, may transform components to different phases, or may alter the fabric of the sediment. Again, many of the processes involved at this stage have definitive characteristics through which process may be inferred.

The net result of these formative processes, both genetic and diagenetic, is a rock body (a *lithosome*) with a suite of characteristics. Some dominant characters are very common, such as being composed of sand-sized particles, and are thus not particularly definitive to process and environment of formation. A combination of characteristics, however, often limits the range of processes and environmental settings, and these are what generally define a facies. Thus a trough cross-bedded sand represents deposition by a migrating series of dunes. That characterization alone may not be sufficient to distinguish the environmental setting between an aeolian dune field, a river channel, or a channelized flow on the seafloor. The association of facies, however, is nearly always sufficient to place the individual facies within its broader sedimentary context.

Two additional points in approaching facies analysis must be emphasized here: the significance of scale and the importance of surfaces. A central aspect of facies characterization and understanding relates to the scale of observation and analysis. Some features of sedimentary rocks can be observed at a fine scale, and indeed some facies analysis is done at the microscopic level. The level of observation, however, must be scaled to the magnitude of the features and the processes involved. Large-scale trough cross-beds, produced by migrating dunes, are commonly tens of centimeters in thickness and meters across. Thus they are features observable primarily at the outcrop scale and could not be recognized through a microscope. Most facies studies, and particularly the approaches highlighted here, depend heavily upon outcrop-scale analysis. A second consideration is the impact of surfaces, the breaks between sedimentary units. Since facies analysis revolves around the tangible components of the sedimentary record, the rocks themselves, it is easy to miss the importance of the surfaces that bound individual sediment bodies. In a temporal sense, these surfaces typically reflect more time than the actual accumulation of the rock record. In terms of process, the recognition of characteristic surfaces can provide crucial additional data for analysis. Erosional surfaces, easily recognized by scalloped scours and downcutting, reflect transitions from an accumulational regime to an erosional phase. Soil surfaces reflect often long periods of time when accumulation is minimal, but pedogenesis of subjacent materials may record important environmental variables. Every surface in the sedimentary record reflects a shift or transition in ambient process, and thus a major portion of the sedimentary history of a sequence is written in these intangible boundaries.

SEDIMENTARY FACIES ANALYSIS

The characterization of sedimentary rocks by facies has a long history, dating back to introduction of the term by Gressley in 1838 (Teichert 1958). Methodologies and particularly the terminology used to undertake this type of analysis have changed considerably through time. A minor revolution in facies analysis took place in the late 1970s and early 1980s, and thus two superficially different approaches may be encountered in literature from the not-too-distant past.

Early analyses grouped "facies" as genetically related packages of sediment, with numerous characteristic components as dominant or minor ingredients in the mix. Thus a package composed of lenticular sediment bodies, in an upward-fining sequence of gravel and sand, with a progression of primary sedimentary structures from trough cross-bedded sands, through planar-bedded sands, to ripple-marked sands would be classed as a facies, and interpreted to represent a meandering river channel or point-bar sequence. In this approach, the individual facies were often named in a descriptive fashion, such as the "lenticular conglomerate and sandstone facies." This approach successfully recognized packages of sediment, with characteristic components in a common but not invariable progression, which reflected a suite of processes in a particular depositional environment, such as a meandering river channel. Continued development of this approach, however, focused attention on the range of variability in the ways in which such packages could be compiled, and led to the recognition that the best fundamental units of analysis were the individual components of these packages. The "revolution" of the 1980s thus moved the appellation *facies* to the fundamental rock unit (in better keeping with its original intent) and viewed the larger packages as *facies associations*. This is the approach adopted by most sedimentologists today, and it is used in the examples set forth here.

A typical facies classification scheme groups sedimentary rocks by primary lithological characteristics, usually particle genesis and/or grain size, and further distinguishes individual facies with primary or secondary features. Thus a study may involve a series of facies comprised of gravels, and distinguished by bedding characteristics, or a series of biogenic deposits, further distinguished by fossil content. Recurring packages of these facies are recognized as facies associations, typically characterized by the dominant facies and their common progression, along with minor facies that may or may not fill out the association. In this approach, the individual facies reflect discrete combinations of components and processes. The associations reflect how these facies come together within natural depositional systems (or in postdepositional landscapes). The degree of potential variability in component facies and their sequence within a facies association is a direct reflection of the real complexity of Earth surface systems.

A number of excellent volumes focus on methodology and examples of this facies approach (Walker 1979; Walker and James 1992; Posamentier and Walker 2006). It is also important to note that much of our understanding of sedimentary facies is based on an interplay between the rock record and modern landscapes. Thus many investigations link ancient facies characters directly to modern analogs (e.g., Miall 1978). To illustrate the facies classification approach, and to explore its utility in ecological reconstructions, an example from the Koobi Fora Formation in northern Kenya is presented here.

FACIES CLASSIFICATION IN THE KOOBI FORA FORMATION

Geologists investigating sedimentary strata of the Koobi Fora region in northern Kenya developed a facies scheme to characterize the deposits and interpret depositional environments beginning in the 1970s (Bowen 1974; Burggraf and Vondra 1982). The initial facies characterization (using the early descriptive approach) recognized four major depositional environments in this nonmarine setting: alluvial fan, fluvial, deltaic, and lacustrine. Continued study and an increased focus on details in localized portions of the sequence resulted in a further elaboration and subdivision of the scheme, recognizing variants on the fluvial and lacustrine facies (Burggraf et al. 1981; White et al. 1981; summarized in Feibel 1983). This development reflected both the natural evolution of thought with more focused study as well as trends in sedimentological philosophy at the time. A comprehensive synthesis of this facies scheme as a model for rift-valley sedimentation was presented by Burggraf and Vondra (1982).

The mid-1980s brought an extensive reevaluation of Koobi Fora stratigraphy (Brown and Feibel 1986) as well as the application of refined facies approaches to the Koobi Fora deposits. Much of the following analysis was first set out in an early report (Feibel and Brown 1986) and subsequently refined and expanded

(Feibel 1988; Feibel et al. 1991). Gathogo (2003; Gathogo and Brown 2006) adopted this scheme in his study of the Ileret region.

The Koobi Fora Formation (Figure 2.1) attains a composite thickness of some 560 meters (Brown and Feibel 1986), exposed discontinuously over about 1200 square kilometers to the northeast of Lake Turkana. These strata reflect deposition in a variety of fluvial, lacustrine and deltaic environments within a series of half-grabens over the interval from circa 4.2 to 0.6 mya (million years ago) (Feibel 1994). Within this setting, the strata express considerable lateral variation reflecting a complex ancient landscape, as well as significant shifts through time. The latter reflect both the dynamics of an evolving landscape as well as extrinsic factors such as climate change, which is an important control on sedimentary dynamics. The Koobi Fora region provides an excellent laboratory in which to explore facies relationships and their significance, because the extensive area of outcrop coupled with superb temporal and stratigraphic control (Brown and Feibel 1991; McDougall and Brown 2006; Brown et al. 2006) allows for an investigation of both temporal and spatial variability over a dynamic landscape.

Facies at Koobi Fora

In the following treatment, individual facies are delineated following the pattern set out by Miall (1977). Dominant lithology along with primary and secondary structures, biogenic character, and bounding surfaces were the primary criteria in identifying the facies. The initial facies characterization developed in this way (Feibel and Brown 1986; Feibel 1988) has required only minor modifications to accommodate continued analysis of the deposits at Koobi Fora, as well as other strata of Miocene to Recent age in the Turkana Basin (e.g., Gathogo 2003; Gathogo and Brown 2006). The only significant extension of the scheme relates to the expanded recognition of characteristic fossil soils, or *pedotypes* (Wynn and Feibel 1995; Wynn 1998), that has further elaborated environments of postdepositional modification.

In this analysis, twenty-one sedimentary facies are described to characterize the Plio-Pleistocene deposits of Koobi Fora (Table 2.1). The major lithological components of significance there are gravels (G), sands (S), fine clastics (muds, M), bioclastic sediments (B), and pedogenically modified sediments (P). Because the relationship between process and product is central both to recognizing sedimentary facies and to their interpretation, a brief discussion of the individual facies and their characteristics and interpretation is presented here.

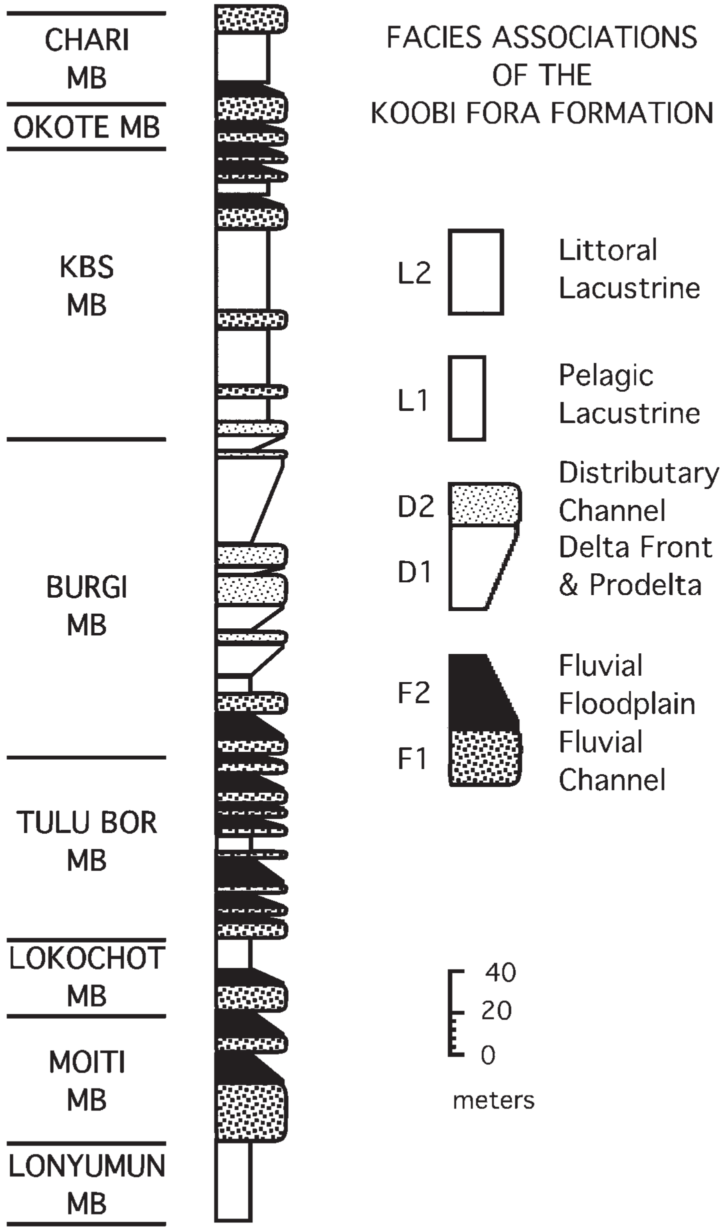

FIGURE 2.1. *Composite stratigraphic column of the Koobi Fora Formation (after Brown and Feibel, 1986). Dominant facies associations and stratigraphic subdivisions are indicated. Note that occurrences of the crevasse association are too small to represent at this scale.*

Table 2.1. Facies classifications applied to the Koobi Fora deposits.

Code	*Facies*	*Features*	*Interpretation*
Gms	gravel, matrix-supported, massive	massive	debris flow
Gm	gravel, massive or weakly-bedded	horizontal bedding, clast imbrication	longitudinal bar, lag deposits
Gp	gravel, bedded	planar cross-beds	linguoid bars and bar accretion
Sm	sand, massive	massive	bioturbated?
Se	sand, bedded	epsilon cross-strata	point-bar surfaces
St	sand, bedded	trough cross-beds	dunes
Sp	sand, bedded	planar cross-beds	linguoid bars, transverse bars, sand waves
Sr	sand, laminated	ripple marks	ripples
Sh	sand, laminated	horizontal lamination	planar bed flow
Fm	sand, silt, clay; massive	massive, bioturbation	suspension deposits (overbank or lacustrine), biologically reworked
Fl	sand, silt, clay; laminated	planar laminated (to thinly bedded)	suspension deposits
Fb	sand, silt, clay; bedded	thin to medium bedded	flow reduction deposits; levee
Ba	cryptalgal carbonate	mats, oncolites and stromatolites	biogenic deposit, cryptalgal
Bd	diatomite	massive to weakly laminated	biogenic deposit, diatomaceous
Bo	ostracod carbonate	massive	biogenic deposit, ostracods
Bm	molluscan carbonate	massive or weakly bedded	biogenic deposit, mollucs
Be	*Etheria* carbonate	agglutinated	oyster bioherm, reef
Pa	mud with sand veins	anastomosing veins	desiccation-cracked vertic epipedon with sand fill
Pv	clay	slickensided dish fractures	compressional vertic subsoil horizon
Pp	clay	prismatic structure	structural subsoil horizon
Pk	pedogenic carbonate	massive or laminated, rhizoliths	Bk horizon

Gravel (G) Facies. The massive gravel (Gm) and planar cross-bedded gravel (Gp) facies are typical of gravel bars in a variety of fluvial systems. The large particle size and shape characteristics of cobbles and pebbles often render stratification weak or absent. In this case, then, the massive nature of the deposit is a primary characteristic. Thin horizontal beds and planar cross-beds typically represent accumulation on the upper surfaces and fronts of bars respectively. Discontinuous thin occurrences of the Gm facies are common at the base of many fluvial cycles, representing the gravel lag accumulated along a channel bottom. At Koobi Fora, these gravels are typically composed of granule- to pebble-sized quartzofeldspathic material. In exposures close to the basin margin, or in the upper half of the formation, these lags may be dominated by well-rounded volcanic clasts. The active reworking of floodplain muds by cutbank erosion leads in some cases to the accumulation of soil carbonate nodules in a Gm facies. These are locally significant, and have been observed to reach up to 0.5 meters in thickness. In rare cases, such as Locality 261–1 (Coffing et al. 1994), the Gm lag may be extensive, and dominated by fossil bone and teeth to the point of forming a "bone bed."

The matrix-supported gravel (Gms) facies is a rare occurrence, in which very poorly sorted material is chaotically mixed, and as a result the gravel component is matrix supported, that is the individual large clasts are often not in contact, but appear to "float" in the matrix of finer sediment. This is a characteristic feature of high-viscosity flows such as debris flows. In the Koobi Fora examples these likely result from short-term fluidization of channel debris, and travel distances are likely short.

Sand (S) Facies. Sands are a prominent component of the sedimentary section of the Koobi Fora Formation (Feibel 1988). Diagenetic cementation is fairly limited in these deposits, and it is likely that many examples of the massive sand (Sm) facies represent disruption of primary fabrics in the near-surface environment. Two levels of organization of primary sedimentary structures are apparent in the sand facies: the structures produced by migration of individual bedforms and the structures produced by channel-form evolution (Miall 1985). Migrating dunes of sand produce trough cross-stratification (St), ranging in size from small-scale (centimeters) to giant (meters) depending on the magnitude of the bedform. Planar cross-bedding in sands (Sp) results from the migration of sand waves. Horizontally laminated sands (Sh) and ripple-laminated sands (Sr) are produced by the movement of thin layers of sand as sheets and ripples. Within a channel, all of these bedforms may be organized along the gently inclined surface of a point bar. This compound-bar form may leave in the sedimentary record a larger-scale structure, referred to as epsilon cross-stratification (Se; Allen 1963a), the signature product of a laterally accreting point bar.

Mud (M) Facies. Fine-grained accumulations ranging from pure clays through silty clays, silts, and sandy clay/silts are all considered collectively here, in part as they form a closely related continuum and in part because they are subject to very similar accumulation and modification processes. Muds typically accumulate through the slow settling of suspended particles in a low-energy environment. Most commonly this reflects lacustrine or floodplain water bodies, but it is likely that aeolian contributions are pervasive but difficult to recognize.

Massive muds (Fm) may result from settling processes in which accumulational discontinuities are too subtle to be reflected in lamination or bedding. Massive muds may also result from pervasive disruption of a primary fabric through burrowing of organisms or rooting, without leaving diagnostic evidence of the biological effects.

Pulsed accumulation of fines, such as that which results from annual floodwater influx may result in laminated or thinly bedded muds (Fl). Larger-scale influx or source-proximal settings (e.g., a levee) may be reflected in bedded muds (Fb).

Mudstones with obvious soil overprint are considered separately in the pedogenic (P) category, but it should be noted that those facies exist in a continuum with the unmodified muds considered here.

Bioclastic (B) Facies. A wide range of organisms produce skeletal material that may dominate a sedimentary facies. Biomineralization in the form of carbonate minerals (calcite and aragonite) are most common, but siliceous forms are also well represented. In most cases body fossils supply ample evidence for classification, but certain biological groups (e.g., algae and bacteria) produce biosedimentary structures in which body fossils may be rare or absent. The algal biofacies (Ba) is most prominently reflected in various stromatolite forms but also in mat forms and more problematic biosedimentary crusts. Diatomites (Bd) are typically massive, low-density siliceous accumulations of diatom frustules, but may also be preserved as more dense recrystallized rocks when diagenetically altered.

Ostracods (calcareous bivalved crustaceans) are nearly ubiquitous in aquatic habitats, but less commonly accumulate to produce discrete biogenic deposits (ostracodites, Bo). Molluscs may be diverse and abundant in aquatic settings, and may be concentrated in both life and death accumulations as distinctive facies (Bm). A distinctive variant of molluscan biofacies is the bioherm (reef) produced by the Nile oyster *Etheria* (Be).

Pedogenic (P) Facies. Early diagenesis of typically fine-grained lithofacies in the near-surface environment results in a broad suite of distinctive features that characterize fossil soils (e.g., Retallack 1990).

The Pa facies was first recognized by Feibel (1983) along the Koobi Fora Ridge. The marginal lacustrine strata there have mudstones that are characterized by sand-filled veins in an anastomosing network. In rare examples where a plan-view of the facies can be seen, the veins have a polygonal pattern. That the veins represented sand-filled crack networks was obvious, but their genesis was not understood until recently. While it was apparent that the drying of smectitic (shrink-swell) clays would open the crack networks, it was unclear how the infill occurred. An aeolian source was possible, but it seemed that introduction of sand by water would cause the clays to swell and close the cracks before they could fill. Recent observations on the Koobi Fora landscape have shown, however, that the crack networks may open up *beneath* an overlying sand cover, and that the fill is more likely a passive process than an active one. In any case, these sand-filled crack networks are a common feature of vertisols worldwide (Wilding and Puentes 1988). In the Koobi Fora Ridge examples, the Pa facies often occurs alone, as an isolated soil horizon. Elsewhere, however, the Pa facies often occurs stratigraphically above the Pv facies, which is a genetically related subsoil.

The same shrink-swell clays responsible for the Pa facies also dominate the Pv facies. This, however, is a subsoil horizon. It is usually observed beneath a Pa facies unit, or alone in an erosionally truncated profile. The process that dominates the Pv is not the contraction (shrink) phase of clay dynamics, but the expansion (swell) phase. Because the Pv is a subsoil horizon, the expansion of swelling clays must displace surrounding material. Significant stress is built up in the process, and the clay fails along arcuate ("dish-shaped") fractures. Movement along these fractures generates the characteristic slickensides in the surfaces of the clay peds.

The Pk facies, commonly referred to as the calcic or "Bk" horizon in soil parlance, is perhaps the most commonly recognized characteristic facies of ancient soils. The utility of pedogenic carbonate in isotopic analysis for paleovegetation reconstructions (e.g., Quade and Levin, Chapter 3, this volume) and paleoprecipitation estimates (Retallack 1994) have been particularly important.

Facies associations

The packaging of individual facies into characteristic associations establishes the critical transition from discrete sedimentary elements to genetically sig-

Table 2.2. Facies associations applied to the Koobi Fora deposits.

Association Code	*Environment*	*Dominant Facies*	*Minor Facies*
F1a	Meandering Fluvial Channel	Se, St, Sp, Sh, Sr	Gm
F1b	Braided Fluvial Channel	Sp	Fm
F2	Fluvial Floodplain	Fm, Pk, Pss, Pa	Pp
L1	Pelagic Lacustrine	Fm, Fl, Bd	Bo
L2	Marginal Lacustrine	Fm, Fl, Bm, Ba, Bo, Pa	St, Sr
D1	Prodeltaic and Delta Front	Fl, Fm	St, Sr
D2	Distributary Channel	Ss, St, Sp, Sr	
C1	Crevasse Channel	St, Sr	Sm
C2	Crevasse Splay	Sm, Fb, Pv	Sr

nificant sequences that reflect components of ancient landscapes. The facies associations described here reflect the specific case of the Koobi Fora Plio-Pleistocene record, but amply demonstrate the methodology of linking facies to establish a context for ancient communities. Much of the description of Koobi Fora facies associations below was first presented by Feibel (1988) and has been expanded in light of subsequent research (Table 2.2).

Fluvial Facies Associations. Three fluvial facies associations are expressed in the deposits of the Koobi Fora Formation (Figure 2.2). These reflect two distinct types of fluvial channel form (F1a and F1b), and a highly variable development of associated floodplain environments (F2).

The Fla sequence typically begins with a basal erosional scour having 10–40 centimeters of local relief, and several meters of relief where channel margins can be seen. The surface of the erosional scour is marked by thin lenticular accumulations of a gravel lag, consisting of quartzo-feldspathic granules and pebbles, along with clay pebbles. These gravel lags represent the Gm facies. This is overlain by a fining-upward progression of sands displaying trough cross-bedding (St), horizontal lamination (Sh), and ripple lamination (Sr), and capped by the fine-grained floodplain deposits. Commonly, however, this ideal sequence has been truncated by a succeeding channel, or primary stratification features have been destroyed by bioturbation. In the Fla sequences of the Lokochot Member and the Tulu Bor Member, channel accretion surfaces (Se facies) are preserved by cementation differences. This type of sequence has been well documented from both modern and ancient examples, and represents deposits of a meandering river channel (Allen 1965, 1971). One extremely well-preserved Fla sequence of the Tulu Bor Tuff (Feibel 1988) allows a calculation of paleochannel width at that point in time. Using the relationship determined by Leeder (1973), a bankfull depth of 9.4 meters measured in one section suggests a bankfull width of about 214 meters. The actual magnitude may have been slightly larger, as the section is truncated at the top. The F1b channel deposits have a less well-developed basal erosional surface, and commonly lack the lenticular Gm accumulations. They are dominated by planar crossbedded sands (Sp) in cosets of 10 to 30 centimeters, with interbeds of horizontally laminated sands (Sh). Capping floodplain deposits are either thin (5–10 cm) or lacking entirely. F1b channels are restricted to the lower Lokochot Member and the lower KBS Member.

The F1b sequence matches modern and ancient examples of braided-stream channels (Smith 1972; Miall 1977). In particular, the features of this association are very similar to the Platte type braided-river profile, a sand-dominated system where sedimentation occurs primarily along the transverse and linguoid bars of a shallow, perennial braided stream. Fluvial floodplain deposits, termed the F2 association (Feibel and Brown 1986), occur superposed on an F1 sequence. They are characterized by their fine-grained nature (dominantly silts and clays), lack of aquatic invertebrate fossils, abundant pedogenic alteration (facies Pv, Pp, Pk; see section on pedogenesis for details), and bioturbation.

Deposits of the F2 association are common throughout the formation. They are generally unremarkable in weathered section except for their darker colors and locally abundant surface lags of white pedogenic carbonate concretions (a Pk lag). Where modern streams have excavated natural exposures, however, the F2 deposits exhibit a wide array of distinctive structures, primarily related to pedogenesis.

Another aspect of significance within the fluvial facies association is the character of the transition from the F1, or channel component, to the F2, or floodplain interval. This transition ranges in character from very gradual, through stepwise in character, to a very abrupt transition. The gradual transition reflects the classical upward-fining sequence seen in the point-bar environment, where a gradual migration of the channel leads to a lowering of energy conditions, and is seen in fining of sedimentary particles and a shift in primary sedimentary structures (Allen 1963b). The F1–F2 transition in this case is marked by a continuous gradation from fine sands, through silts, to clays. It is not uncommon, however, for the transition to be recorded in a more stepwise transition, with sharp stratigraphic breaks and steps in particle-size change. In some such cases, reversals occur where the general upward fining trend is interrupted by a shift to coarser particles. In some cases this reflects minor complexity on a point-bar

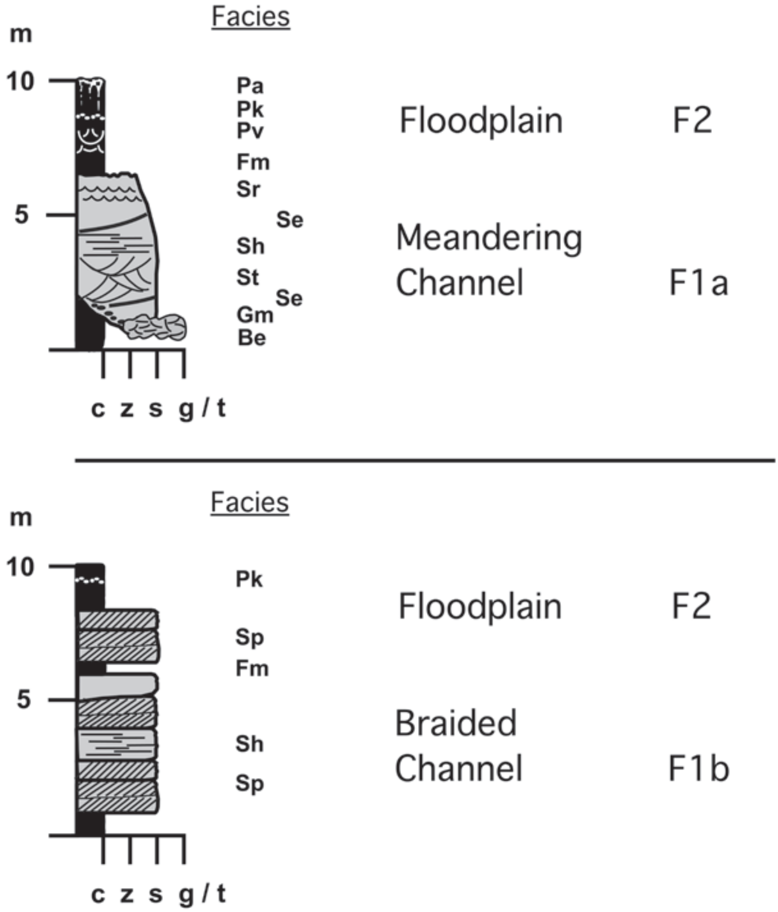

Figure 2.2. *Characteristic features and sequences of the fluvial facies associations. Idealized stratigraphic columns are depicted, showing thickness and dominant lithology (c ,clay; z, silt; s, sand; g, gravel; t, tuff). Facies codes are described in Table 2.1.*

surface, such as chute cutoffs. In other examples the abrupt shifts reflect episodic accumulation. An excellent example of this is seen in a complete point-bar sequence preserved in the Tulu Bor Tuff (Feibel 1988; 1999). In this case, the upward-fining progression in the tuff is broken by three abrupt shifts to much finer-grade material. These are interpreted to record three flood–slack couplets during the accumulation history of the tuff, and may actually record three annual events of flood-season and waning-stage flow. The extreme example of abrupt F1–F2 transition is documented in fluvial associations in which a fluvial couplet is seen. In this case, a relatively coarse sand F1 association is overlain abruptly by a clay F2, with no transitional particle-size gradation. These fluvial couplets are interpreted to reflect avulsion events (Smith et al. 1989), in which the position of the active channel has abruptly shifted on the floodplain, and a channel F1 association is left to be overlain by distal floodplain vertical accretion.

Lacustrine Facies Associations. Deposits of lacustrine facies comprise about one-third of the Koobi Fora sediments. Characteristic sequences of the lacustrine facies allow them to be grouped into two major associations (Figure 2.3), one representing an open-water or pelagic lacustrine environment (Ll) and one the marginal lacustrine setting (L2).

The features that characterize the Ll facies association are sedimentation from suspension, with a low clastic input. Claystones predominate, either in massive (Fm) or finely laminated (Fl) facies. Biogenic accumulations may be associated, including diatomites (Bd) and scattered or abundant ostracods (Bo). Molluscs also occur scattered throughout these facies in some localities. This association is not well represented in the Koobi Fora sequence, suggesting that either dominantly marginal conditions or proximity to a source of clastic detritus (resulting from deltaic encroachment) were more typical of the lacustrine phases. Pelagic lacustrine deposits are known from the lower Lonyumun Member, Lokochot Member, mid–Tulu Bor Member, and upper Burgi Member.

The littoral lacustrine (L2) association is marked by a heterogeneous mixture of facies, often closely and repetitively interbedded. The term *littoral* is used here in its lacustrine sense, meaning “the shore region where the water is shallow enough for continuous mixing and for photosynthesis to the bottom” (Beadle 1981). In practice this implies that most of the littoral zone is affected to some degree by wave action. The facies common in this environment include molluscan carbonates (Bm) that are dominantly life assemblages but that can be seen to grade laterally into more arenaceous, reworked shoreline deposits representing death assemblages. Another common facies in this assemblage is the cryptalgal biolithite (Ba), which includes carbonates in the form of oncolites, stromatolites, and cryptalgal mats. Some ostracodites (Bo) occur in this association, although more commonly the ostracods are relatively dispersed. The association is dominated volumetrically by fine-grained clastics, commonly massive (Fm) but locally with well-developed pedogenic features (Pv). Thin (1–2 m) beds of laminated siltstones and fine-grained sandstones (Fl) also occur in this association. These are actually a very thin variant of the prodeltaic association discussed below. The characteristic facies and sequences of the littoral lacustrine association are best developed in the uppermost upper Burgi and the

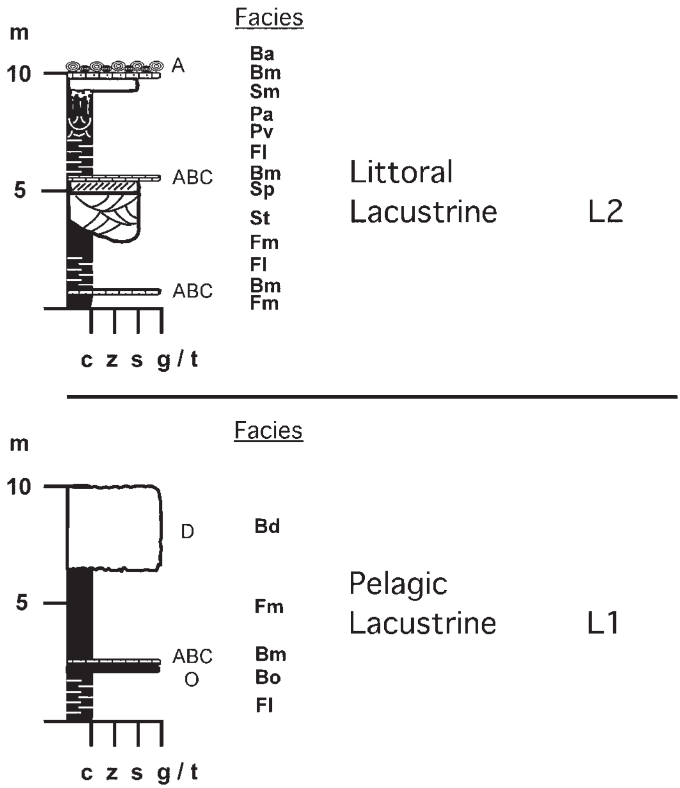

Figure 2.3. *Characteristic features and sequences of the lacustrine facies associations. Idealized stratigraphic columns are depicted, showing thickness and dominant lithology (c, clay; z, silt; s, sand; g, gravel; t, tuff). Facies codes are described in Table 2.1.*

KBS Members. The association is also well represented in the upper Okote and Chari Members at Ileret.

A major distinction between the two lacustrine associations is the frequency of subaqueous/subaerial cyclicity and the degree of development of littoral biotic accumulations. The facies of the pelagic lacustrine association imply conditions minimally affected by nearshore processes (wave activity, high clastic influx). The facies of the littoral lacustrine association are indicative of a closer proximity to the shoreline and a relatively frequent lateral migration across the shore transition. There is no indication that the pelagic lacustrine association developed in a deepwater (profundal) environment. There are rare examples of desiccation and minor soils development in the pelagic clays of the Lonyumun Member. Analysis of the deltaic facies associations (following section) suggests that maximum lake depths were on the order of 40 meters. One of the most distinctive features of the littoral lacustrine environment at Koobi Fora is the well-developed molluscan carbonate (Bm) facies. Feibel (1983) suggested that one of the best available analogs for this facies was modern Lake Chad. There, relatively shallow (2–3 m) waters host extremely abundant mollusc populations that are compositionally almost identical to those of the Koobi Fora Formation (Lévèque 1972). Recent history has also shown that this sort of environment is quite unstable, and changes in lake level result in large fluctuations in lake area.

Deltaic Facies Associations. Facies associations attributable to deltaic environments comprise some 20 percent of the Koobi Fora deposits. Temporally, however, they are overrepresented, as the characteristic high-sedimentation rates of the deltaic environment produces an extremely thick section over a relatively short period of time. The deltaic environment can be characterized by two facies associations, one representing the prodelta and delta-front environments (D1) and the other the product of distributary channels (D2). The characteristic features of the deltaic facies associations are illustrated in Figure 2.4.

The prodelta and delta-front association (D1) is dominated by laminated fine-grained clastics (Fl), including clays, silts, and fine sands, which occur in thick (10–20 m) packages. The laminated sediments are commonly capped by a massive (commonly bioturbated) fine-grained sequence (Fm) or by delta-front sands (Sm). As mentioned earlier, the distributary channel association (D2) is very similar to the F1 associations, and is most easily distinguished by its sequential position atop a D1–D2 couplet. The medium- to coarse-grained sands of this association occur as narrow lenticular bodies within the D1 deposits, or as broader lenticular deposits overlying them. Trough (St) and planar (Sp) cross-bedding and ripple lamination (Sr) is generally weakly developed, or has been destroyed by bioturbation. The D1 and D2 associations are well developed in the upper Lokochot Member and in the upper Burgi Member. They are not represented in association with the lakes of the Lonyumun Member or Tulu Bor Member. A very thin variant of these associations is represented in the KBS through Chari Members. The D1 interval is only 1–2 meters thick, and the D2 association is characterized by a deeply erosional base. These thin intervals of D1–D2 sedimentation are characteristically associated with the littoral lacustrine deposits of the upper Koobi Fora Formation.

The well-developed D1–D2 sequences of the Lokochot and upper Burgi Members allow for determination of water depths (deVries Klein 1974). For the Lokochot Member lake, the deltaic infilling suggests water depths of 10–15 meters. The upper Burgi Member lake records several intervals of deltaic pro-

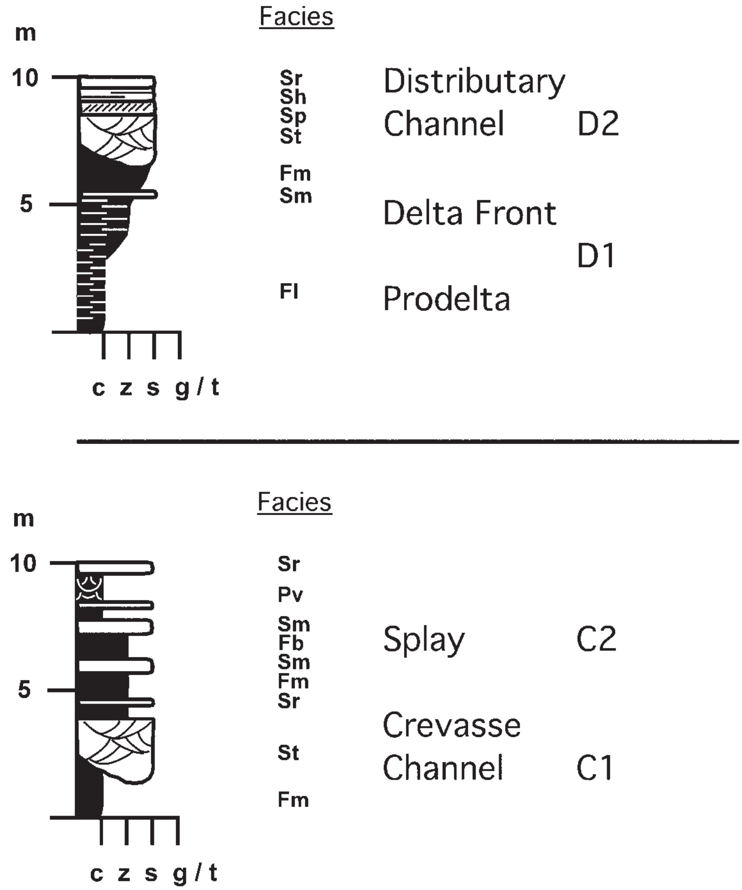

FIGURE 2.4. *Characteristic features and sequences of the deltaic (*top*) and crevasse (*bottom*) facies associations. Idealized stratigraphic columns are depicted, showing thickness and dominant lithology (c, clay; z, silt; s, sand; g, gravel; t, tuff). Facies codes are described in Table 2.1.*

gradation, with water depths of different phases ranging from 20 to 40 meters (Feibel and Brown 1986). It is also possible to use this method to characterize the water bodies present at Koobi Fora in the KBS, Okote, and Chari Members. These intervals of upward-coarsening cycles are only 1–2 meters in thickness, and suggest that the water bodies present had depths of that order. Whether this represents the total depth of a lake or only a marginal zone cannot be determined from the data available.

The extremely well-developed lamination of silts and very fine-grained sands in the D1 association has important implications. The lamination generally occurs on a 0.5-centimeter scale and is interpreted to represent a strongly seasonal influx of detrital clastic material. A similar sort of lamination is seen in modern Lake Turkana, where it results from the seasonal nature of the annual Omo River floods (Yuretich 1979). A similar origin is inferred for the Fl deposits of the Dl association. This implies that the monsoonal seasonality of rainfall characteristic of the Ethiopian Highlands today was a feature of the Plio-Pleistocene as well.

Crevasse Facies Association. A small-scale variant sharing some characteristics of the deltaic facies association is the crevasse facies association (Figure 2.4). These packages occur throughout the Koobi Fora Formation, but are particularly well developed in intervals of high sediment accumulation, such as the Okote Tuff Complex (Brown and Feibel 1985) and comparable strata at Ileret (Quinn and Lepre 2005). Crevasse systems develop as breakouts of major channels, with the crevasse itself cutting through channel bank or levee deposits, and a splay forming as a minidelta on the adjacent floodplain or interdistributary bay (Elliott 1974; Bristow et al. 1999).

The crevasse channel (C1) component is typically dominated by trough cross-bedded sands (St) and climbing ripple-laminated sands (Sr) within an erosionally scoured channel-form body. In some examples, the basal portion of this channel is lined with coarse tuff, which may be pumiceous. The C1 component merges laterally with the splay complex (C2), which is dominated by thin massive (Sm) or rippled (Sr) sands, bedded silts (Fb), and clays with pedogenic overprint (Pv). Crevasse splays result in rapid localized accumulation on the floodplain or distributary bay, and may be significant in the burial and preservation of fossil or artifact accumulations.

FROM FACIES ASSOCIATIONS TO LANDSCAPES AND PALEOECOLOGY

The sedimentary records that accompany fossil and archaeological assemblages are a rich archive of environmental indicators upon which paleoecological reconstructions can be based. Facies analysis provides a useful methodology for recognizing fundamental rock units, genetically significant associations, and for their use in reconstructing both depositional and postdepositional sedimentary environments. The direct connection between fossils or artifacts and their sedimentary context provides an immediate link between the sedimentary proxies and the associated records, but in addition the facies provide a means of extending the analysis in two dimensions: across spatial landscapes and through time.

A facies analysis is a first step in reconstructing the dynamic landscape that supported ancient communities, and for understanding the interplay of processes that preserved fossil and artifact assemblages. It provides some information on crucial variations of the hydrologic system in the region, central to ecological relationships, as well as a basis for reconstructing patterns of vegetation structure based on soil and moisture relationships (Tinley 1982). Although the geological substrate represents a nonbiological actor in ecological relationships, it is nonetheless a crucial element in paleoecological reconstructions. Upon this reconstructed landscape of sediments, landforms, and associated moisture constraints, it is possible to array the biological and cultural evidence, the sum of the paleontological and archaeological records, and investigate their spatial and temporal relationships and interactions, which is the central goal of paleoecology.

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