GEOLOGY OF THE OMO-TURKANA BASIN OF EASTERN AFRICA: TECTONICS AND CLIMATE IN THE EVOLUTION OF A PLIOCENE AND PLEISTOCENE SEQUENCE

by

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STATEMENT OF DISSERTATION APPROVAL

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ABSTRACT

Many studies present climate change as the principal cause of the most prominent shifts in depositional environments in the Omo-Turkana Basin (eastern Africa) during the Pliocene and Pleistocene Epochs. For example, some studies relate the development of major lake sequences to presumed high precipitation during Milankovich's eccentricity maxima. The hypothesis of this dissertation is that major shifts in depositional facies in the basin took place only when tectonic activities allowed, including the development of at least four of the basin's most prominent lake sequences at about 4.0 Ma, 3.5 Ma, 3.2 Ma, and 2.0 Ma. Corroborating field evidence includes pronounced changes in the basin's depositional facies in association with tectonic-related features such as prominent syndepositional faulting and local to basin-wide basaltic volcanism about 4.0 Ma, 3.2–3.3 Ma, and 2.02–2.18 Ma. A distinction between climatic and tectonic influences on ancient habitats in the Omo-Turkana Basin is critical in paleontological and paleoanthropological studies because the sedimentary record of this basin and associated fossils and archeological artifacts are widely cited as among the best examples of the role of climate in the evolution of modern African mammals and humans at large. The hypothesis presented in this dissertation illustrates geographic location and tectonic-related physiography as significant factors in the Omo-Turkana Basin in addition to regional and global climate.

This dissertation submits that substantial deposition of Pliocene and Pleistocene strata in the Omo-Turkana Basin was initiated and sustained primarily by rift tectonics. The Omo River, with headwaters in the Ethiopian volcanic highlands, is presented as the main supplier of both sediments and water in the basin since the Early Pliocene. The geologic record shows that the river has remained perennial, regularly draining outside the basin, even reaching the Indian Ocean, and at times flooding the basin to form expansive tectonic lakes. Precession- to eccentricity-scale astronomical forcing of climate seems to have played a subordinate role in modulating the sediment supply and the formation of extensive lakes except for second- to third-order fluctuations in lake levels—a rise by a few tens of meters possibly for a few thousand years.

Dedicated to Christina W. Gathiga, Francis H. Brown, and Meave G. Leakey.

…[T]he past history of our globe must be explained by what can be seen to be happening now… No powers are to be employed that are not natural to the globe, no action to be admitted except those of which we know the principle…(Hutton, 1788)

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My thesis advisory committee, comprised of F. H. Brown, T. E. Cerling, R. L. Bruhn, J. M. Harris, and M. A. Chan, helped me greatly in compiling the results of this study. They provided me with constant support, with organizations of ideas, and with guidance to references about the subject of the study. It was a privilege to have F. H. Brown and R. L. Bruhn join me during field work in the Loiyangalani area, where they helped decipher some problematic relations between geological units, and also utilized satellite imagery in helping with interpretations of the geology. I express my great appreciation to F. H. Brown for allowing me to use his immense database on the composition of tuffs of the Turkana Basin, and for sharing his ideas about the geologic evolution of the basin.

My enthusiasm for, and familiarity with, the field geology of the area comes mostly from M. G. Leakey, who enabled me to visit and observe contemporaneous deposits in many parts of the basin since 1997. I am fully aware that R. E. Leakey has been of major assistance to all work in the basin, though he remained in the background during the time of my study due to his other responsibilities. I thank him for his foresight in easing logistical challenges in this remote region. Others who helped me very significantly in the field were my local assistants Hillary Sale, Marco Barini, Justus Edung, Gabriel Ekalalei, Cyprian Nyete, and Simon Ilar, who never complained even though on many days we walked more than 30 km, sometimes through thick thornbush, to map isolated outcrops.

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CHAPTER 1

INTRODUCTION

Detailed studies of the geology of the Omo-Turkana Basin in eastern Africa (Fig. 1) have been motivated primarily by the quest to understand the context of a large variety of fossils from the basin and elsewhere. Fossils represented in the basin range from dinosaurs (Arambourg and Wolff, 1969; Sertich et al., 2005) to the earliest occurrence of an anatomically modern human (McDougall et al., 2004). The basin has attracted attention in the research on prehistory because of a combination of unique attributes that include extensive paleontological sites with highly fossiliferous deposits along with associated archeological sites. Moreover, age control of the basin's Pliocene and Pleistocene sequence is excellent with over 40 different dated volcanic units (McDougall and Brown, 2006, 2008; Brown et al., 2006; Brown and McDougall, 2011; McDougall et al., 2012). Many hominid species have been retrieved from the basin, including the type specimens of *Homo ergaster* (Groves and Mazak, 1975), *H. rudolfensis* (Alexeev, 1986), *Australopithecus anamensis* (Leakey et al., 1995), and *Kenyanthropus platyops* (Leakey et al., 2001). Many parts of the basin are recognized as World Heritage Sites (http://whc.unesco.org/en/news/143) by the United Nations because of some combination of the aforementioned attributes.

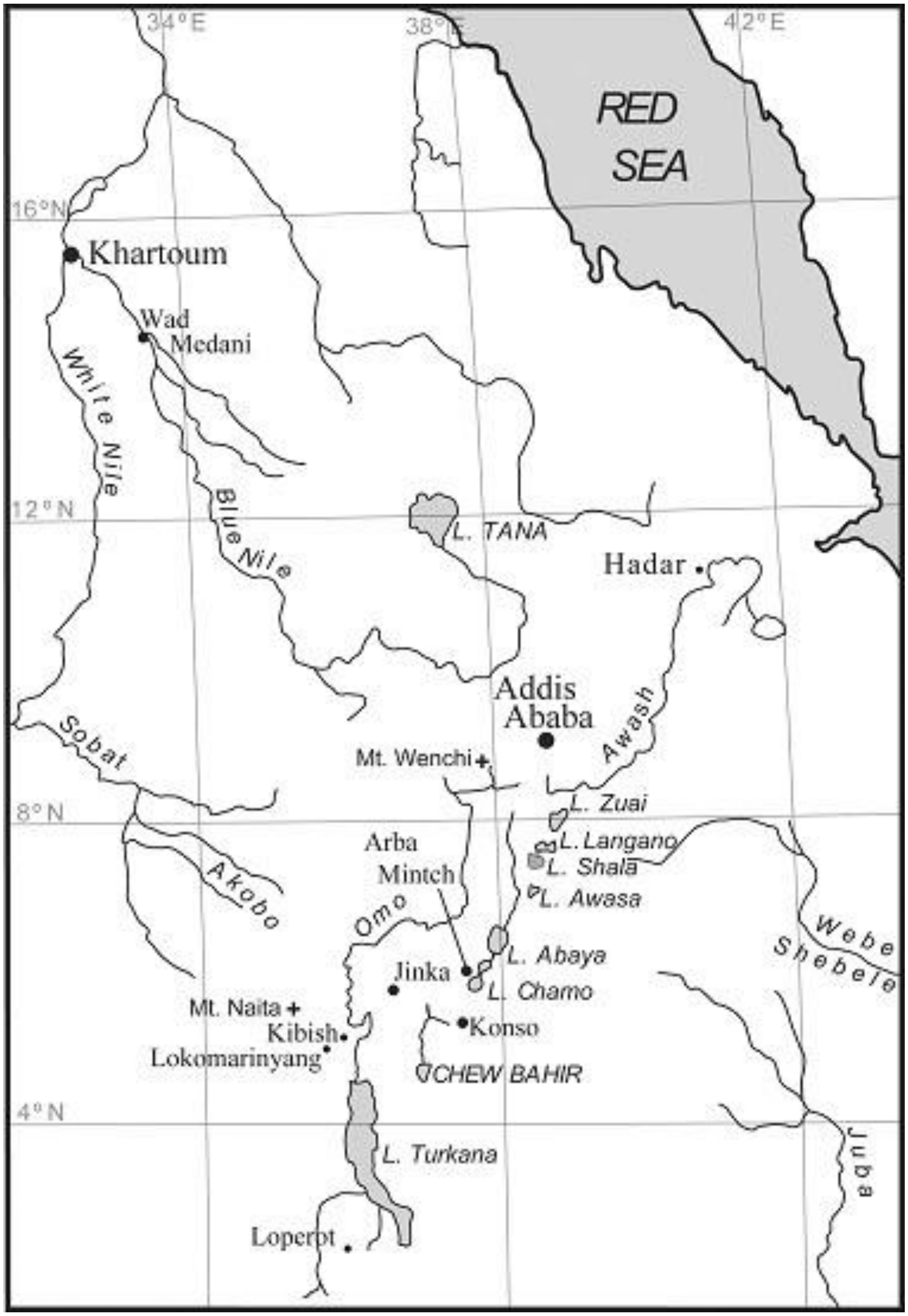

Figure 1. Map showing the location of the Lake Turkana drainage basin, which defines the Omo-Turkana Basin in eastern Africa.

A composite section made up of exposed strata in the Omo-Turkana Basin represents an exceptionally continuous geologic record—spanning the Miocene to the present—that contains abundant fossils (de Heinzelin, 1983; Brown and Feibel, 1986; Owen and Renaut, 1986; Harris et al., 1988a, b; Harris and Leakey, 2003; Leakey and Harris, 2003). Fossil collections from this sequence of strata provide key reference points for local and continental biostratigraphic schemes in which the succession of mammalian taxa in families such as Suidae, Elephantidae, and Bovidae has been combined with isotopic ages (Cooke and Maglio, 1972; Coppens, 1972; Maglio, 1973; Beden, 1976; White and Harris, 1977; Cooke, 1978; Coppens et al., 1978; Beden, 1979a, b; Harris and White, 1979).

The basin's geologic record and fossil data also have been extensively analyzed for possible effects of climate on the evolution of macromammals, particularly early humans (Vrba, 1996; Behrensmeyer et al., 1997; Bobe et al., 2002; deMenocal, 2004; Hernández-Fernández and Vrba, 2006; Anton, 2007). Some studies report astronomical climate forcing as the main cause for pronounced changes in depositional environments within the basin during the Pliocene and Pleistocene Epochs. The development of lacustrine facies, marking times when the basin was filled with water, is one commonly cited line of evidence used to support this view (Trauth et al., 2005, 2007, 2009; Behrensmeyer, 2006; van Bocxlaer et al., 2008; Maslin et al., 2014). This dissertation provides new field-based evidence that shows that the main lake sequences in this basin were tectonically related. It also describes how, in addition to global climate, the geographic location and tectonically related physiography of the Omo-Turkana Basin may influenced the local distribution of mammals (including hominids) in the basin and may even have contributed to the basin being a refuge for water-dependent mammals in the face of progressive aridification of the region.

New field-based evidence presented in this dissertation suggests that rift-related tectonic activities in the Omo-Turkana Basin were key to the development of major depositional environments in this basin throughout the Pliocene and Pleistocene Epochs. For example, structural movements and local basalt flows coincide with the development of expansive lacustrine intervals and alluvial fan deposits during the Pliocene. Tectonic activity, which can be confused with climatic effects in many rift basins, is known to control the development of major depositional settings including alluvial fans and lacustrine facies (Frostick and Reid, 1989; Carroll and Bohacs, 1999). No comprehensive study that attempts to distinguish between climatic and tectonic signals in the Omo-Turkana Basin has been reported prior to this dissertation.

Study Area: Physical and Human Geography

This dissertation uses the term Omo-Turkana Basin in reference to the outcrop area of the Omo and Turkana stratigraphic sequences (de Heinzelin, 1983; Brown and Feibel, 1986; Harris et al., 1988a, b; Leakey and Harris, 2003). The basin is part of the drainage area of Lake Turkana and extends from southern Ethiopia to northern Kenya (see Fig. 1). The lake's catchment area is about 143,000 km², including 7,000 km² occupied by the present lake surface. Elevation in the drainage basin ranges from about 365 m (above mean sea level) at lake level to nearly 3,560 m in parts of the Ethiopian highlands, and to nearly 4,000 m on Mt. Elgon on the border between Kenya and Uganda. The northern part of the basin, including the Ethiopian highlands, falls within the tropical climatic zone and is characterized by summer rains and deciduous woodland, whereas the central part in Kenya corresponds to the transitional semiarid climatic zone with savanna vegetation (Walter, 1970; Hernández Fernández and Vrba, 2005). The very arid southern part of the basin in Kenya is considered as a Class VI ecologic zone, with semidesert vegetation (Pratt et al., 1966; Woodhead, 1970; Pratt and Gwynne, 1977). Three wildlife protection areas are situated in the northern and central parts of the basin: i) Omo National Park in the Lower Omo Valley, Ethiopia; ii) Mago National Park in the Usno Valley, Ethiopia; and iii) Sibiloi National Park along the eastern side of Lake Turkana, Kenya.

The indigenous inhabitants of the Omo-Turkana Basin are ethnically diverse, and their languages are commonly reflected in the formal names of geologic units in the basin. They are primarily pastoralists but they also practice fishing. Native speakers of eastern Sudanic languages including the Turkana and Samburu are dominant along the western and southern parts of Lake Turkana, respectively. The area west of the Omo River is occupied by the Nyang'atom and Mursi who also speak eastern Sudanic languages. Native speakers of east Cushitic languages including Rendille, Gabbra, and Dhaasanac dominate in that order from south to north along the eastern part of the lake. The Dhaasanac people extend northward from Kenya along the lower reaches and delta of the Omo River in southern Ethiopia. Observations made in the course of this study indicate that inhabitants of the regions along the Omo River and around Lake Turkana have retained their material culture and traditions better than most other native east Africans.

Previous Studies: Review of Geology and Paleontology

The history of research work in the Omo-Turkana Basin can be divided into three phases: i) pioneering work done during the first four decades of the 1900s; ii) detailed, multidisciplinary investigation that focused on parts of the basin during the 1960s and 1970s; and iii) refinement of the stratigraphic framework and regional correlation in the 1980s and 1990s largely based on previous studies. A brief history of research and major findings in the basin is given by Harris et al. (2006).

Some early documentation of expeditions in the Lake Turkana and Lower Omo Valley regions comes from the accounts of S. Teleki von Szék and L. R. von Höhnel on their first expedition in 1888 (von Höhnel et al., 1891). The two explorers formally named the lake "Rudolf" in honor of the Crown Prince Archduke Rudolf of Austria. In 1975, the government of Kenya renamed the lake "Turkana" after the most populous ethnic group in the region. The first paleontological collection from the basin, which included Pliocene and Pleistocene mammals, resulted from an expedition led by R. du Bourg de Bozas to the Lower Omo Valley in 1902-1903 (Bourg et Bozas, 1903, 1906), followed three decades later by extensive geologic and paleontologic studies in the Lower Omo Valley and West Turkana by a team that included C. Arambourg (Arambourg, 1933, 1935, 1943, 1947, 1948). Various workers conducted parallel work in other parts of the basin (e.g., Fuchs, 1934, 1935; Champion, 1935, 1937), focusing particularly on geologic mapping. Brown (1989) recounts some of the experiences of early European explorers in the Omo-Turkana Basin before the 1960s and during the first phase of studies.

The present understanding of the stratigraphy of the Omo-Turkana Basin began to take shape during the second phase of studies in the mid-1960s. At that time, F. H. Brown was assigned (Howell and Coppens, 1983) to carry out geological studies in the Lower Omo Valley in preparation for a subsequent collaborative team, the International Omo Research Expedition, led by C. Arambourg, F. C. Howell, L. S. B. Leakey, and R. E. F. Leakey (de Heinzelin, 1983). The team conducted detailed paleontological and geological studies; F. H. Brown, J. de Heinzelin, P. Haesaerts, and J. Chavaillon are among the geologists who described the Omo Group (de Heinzelin, 1983), including the Usno Formation (up to 182 m; de Heinzelin and Haesaerts, 1983a), the Shungura Formation (~766 m; de Heinzelin and Haesaerts, 1983b), and the Mursi Formation (at least 110 m; de Heinzelin and Haesaerts, 1983c). K. W. Butzer named the Nkalabong Formation (~ 90 m) during this time as well (Butzer, 1973). These workers divided the Shungura Formation into 13 members: at the time, the Basal Member and Members A through G were regarded as being Pliocene, whereas and H and J to L were regarded as being mostly Pleistocene. Recent changes in the definition of the boundary between the Pliocene and Pleistocene Epochs (Ogg and Pillans, 2008) now place Basal through Member C in the Pliocene and higher members in the Pleistocene. The Usno Formation is made up of 20 Pliocene units (U-1 to U-20). The Mursi Formation includes eight informal units, M-T to M-L, all of which are Pliocene. Subdivision of the Shungura Formation into members is primarily based on the occurrence of thick tuffs, while further subdivision into submembers commonly distinguishes upward fining beds separated by low-level erosional surfaces. Figure 2 is a map of the general outcrop geology of Pliocene and Pleistocene rocks in the basin.

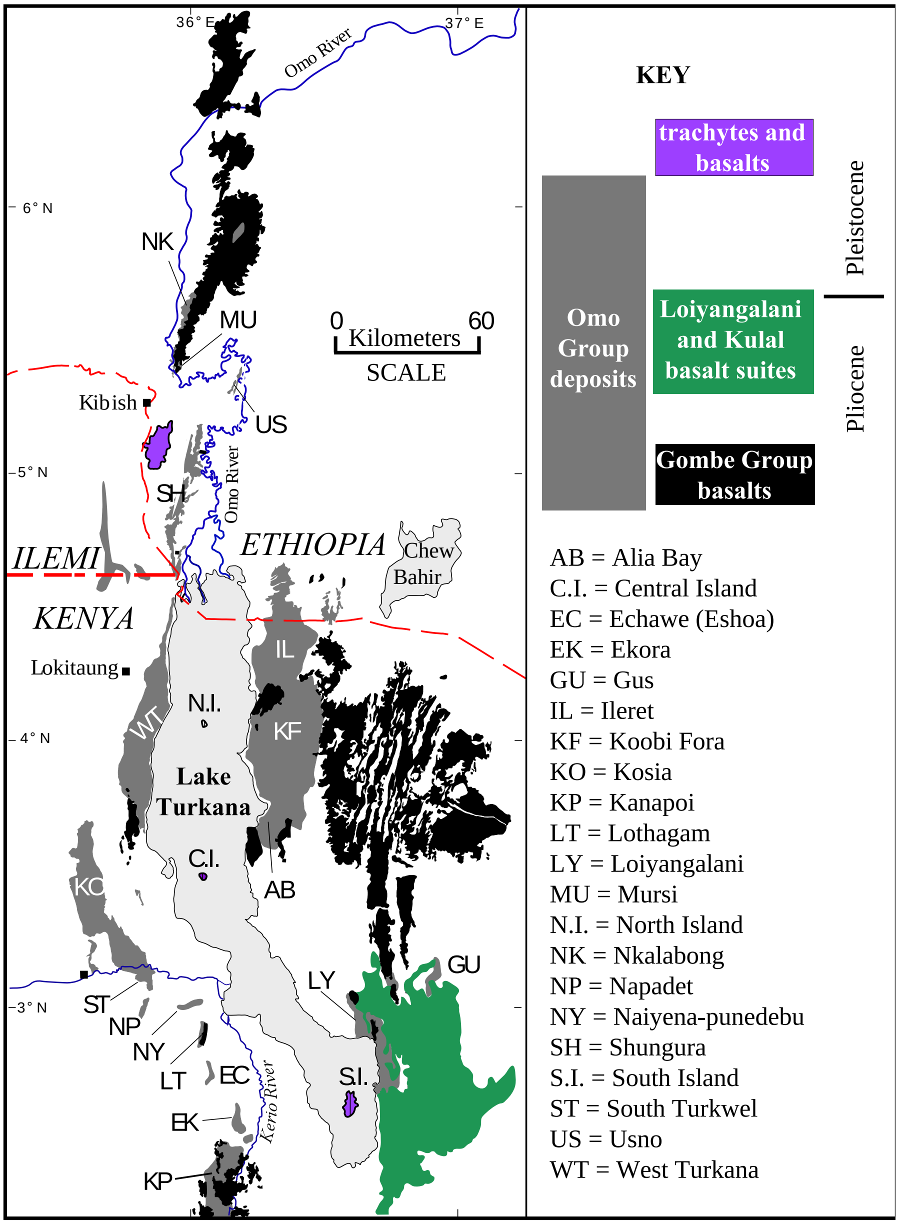

Figure 2. Map showing principal outcrops of Pliocene and Pleistocene sedimentary deposits along with volcanic rock units of similar age in the Omo-Turkana Basin.

Geological and paleontological studies in the Kenyan parts of the Omo-Turkana Basin during the 1960s and 1970s included detailed studies in the Kanapoi and Lothagam areas by a team led by B. Patterson (Patterson and Howells, 1967; Patterson et al., 1970). Powers (1980) carried out a detailed investigation of the geology in the southwestern part of the basin in the late 1970s. In 1968, R. E. Leakey led a team that included personnel from the National Museums of Kenya and international researchers to investigate newly discovered fossiliferous outcrops east of Lake Turkana (Behrensmeyer, 1970; Patterson et al., 1970).

The stratigraphy of this eastern part of the basin was investigated by many workers (Fitch and Miller, 1970; Vondra et al., 1971; Fitch et al., 1974; Cerling, 1976, 1977; Findlater, 1976a, b; Leakey and Leakey, 1978) during the 1970s, and it received far more attention than other units of the Omo Group in the basin, particularly considering the number of master's theses and doctoral dissertations that were written on the area during the decade (Bowen, 1974; Acuff, 1976; Bainbridge, 1976; Burggraf, 1976; Findlater, 1976a; White, 1976; Cerling, 1977). In contrast, the western part of the basin received little attention during the same period, and the Pliocene and early Pleistocene geology of the area was virtually unknown until 1981 when a reconnaissance study was conducted by F. H. Brown and J. M. Harris accompanied by a team including K. Kimeu and P. Nzube from the National Museums of Kenya (Harris et al., 1988a, b).

The third phase of geologic studies in the basin, which started in the early 1980s, resulted in major revisions of the local and regional stratigraphic framework primarily based on geochemical identification and dating of tuffaceous beds (McDougall et al., 1980; Brown and Cerling, 1982; Cerling and Brown, 1982; McDougall, 1985). Findings from these studies facilitated and necessitated revision of the stratigraphic framework in the Omo Group sequences, particularly the Koobi Fora Formation (565 m thick; Brown and Feibel, 1986) east of Lake Turkana. Brown and Feibel (1986, p. 299) define the Koobi Fora Formation as

> sedimentary strata of the Koobi Fora region of Pliocene and Pleistocene age that lie disconformably or unconformably on, or are in fault contact with, Miocene and Pliocene volcanic rocks and/or associated sediments, and are disconformably overlain by the late Pleistocene and Holocene Galana Boi Beds of Bowen and Vondra (1973).

The formation is divided into eight members based on chemically distinctive volcanic ash layers. From bottom to top, the members are: Lonyumun, Moiti, Lokochot, Tulu Bor, Burgi, KBS, Okote, and Chari. Using the definitions given in Ogg and Pillans (2008), the lower four members and the Burgi Member below the Lokalalei Tuff are Pliocene and the remaining three are Pleistocene, placing the boundary between these Epochs at 2.6 Ma.

Revisions of the stratigraphy in the eastern part of the basin allowed for refined correlation of the Omo Group sequence across the basin including the Shungura Formation of Lower Omo Valley and the then newly defined Nachukui Formation (~730 m) of the western part of the basin. Harris et al. (1988a, p. 3) describe the Nachukui Formation as

> poorly consolidated sandstones, siltstones, claystones, and conglomerates that rest disconformably on, or are in fault contact with, Miocene volcanic rocks and intercalated sediments and/or on Precambrian gneisses.

They divide the formation into eight members on the basis of volcanic ash layers of distinctive chemical composition, most of which are correlated with units in the Shungura and Koobi Fora Formations. The Pliocene members are named Lonyumun, Kataboi, and

Lomekwi in order from the base, and the Pleistocene members are named Lokalalei, Kalochoro, Kaitio, and Nariokotome, again with the oldest first in the sequence.

Stratigraphic marker beds in the Omo Group include volcanic ash beds, which have been the main basis for regional correlation within the Omo-Turkana Basin (Figure 3) and for correlation with sequences of similar age in the surrounding basins (Brown, 1982; Haileab and Brown, 1992, 1994). Tephrostratigraphic correlation of the Omo-Group sequence with deep-sea strata in the Arabian Sea and Gulf of Aden also has been achieved (Sarna-Wojcicki et al., 1985, Brown et al., 1992; deMenocal and Brown, 1999; Feakins et al., 2007). Feldspars from volcanic eruptions that are incorporated in pumices in tuffaceous beds and also intercalated basalt flows provide precise age control for many levels in the Omo Group sequence (McDougall and Brown, 2006, 2008; McDougall et al., 2012) that are summarized in Table 1. Identification of magnetic polarity boundaries (magnetostratigraphy) in the basin (Brock and Isaac, 1974, 1976; Hillhouse et al., 1977; Powers, 1980; Brown and de Heinzelin, 1983; Hillhouse et al., 1986; McDougall et al., 1992; Kamau, 1994; Kidane et al., 2007, 2014; Joordens et al., 2011, 2013; Kidney, 2012) has provided additional control on the age of various parts of the sequence, and also an independent check on age estimates in other parts of the sequence (Table 2).

Table 3 shows the taxonomic representation of macromammals recovered from the Pliocene and Pleistocene strata of the Omo-Turkana Basin and includes genera of primates, carnivores, proboscideans, perissodactyls, and artiodactyls. The list incorporates data from various publications (Leakey and Leakey, 1978; Harris, 1983; Harris et al., 1988b; Wood, 1991; Leakey et al., 1995; Isaac and Isaac, 1997; Werdelin and Lewis, 2000; Bobe and Eck, 2001; Leakey et al., 2001; Bobe et al., 2002;

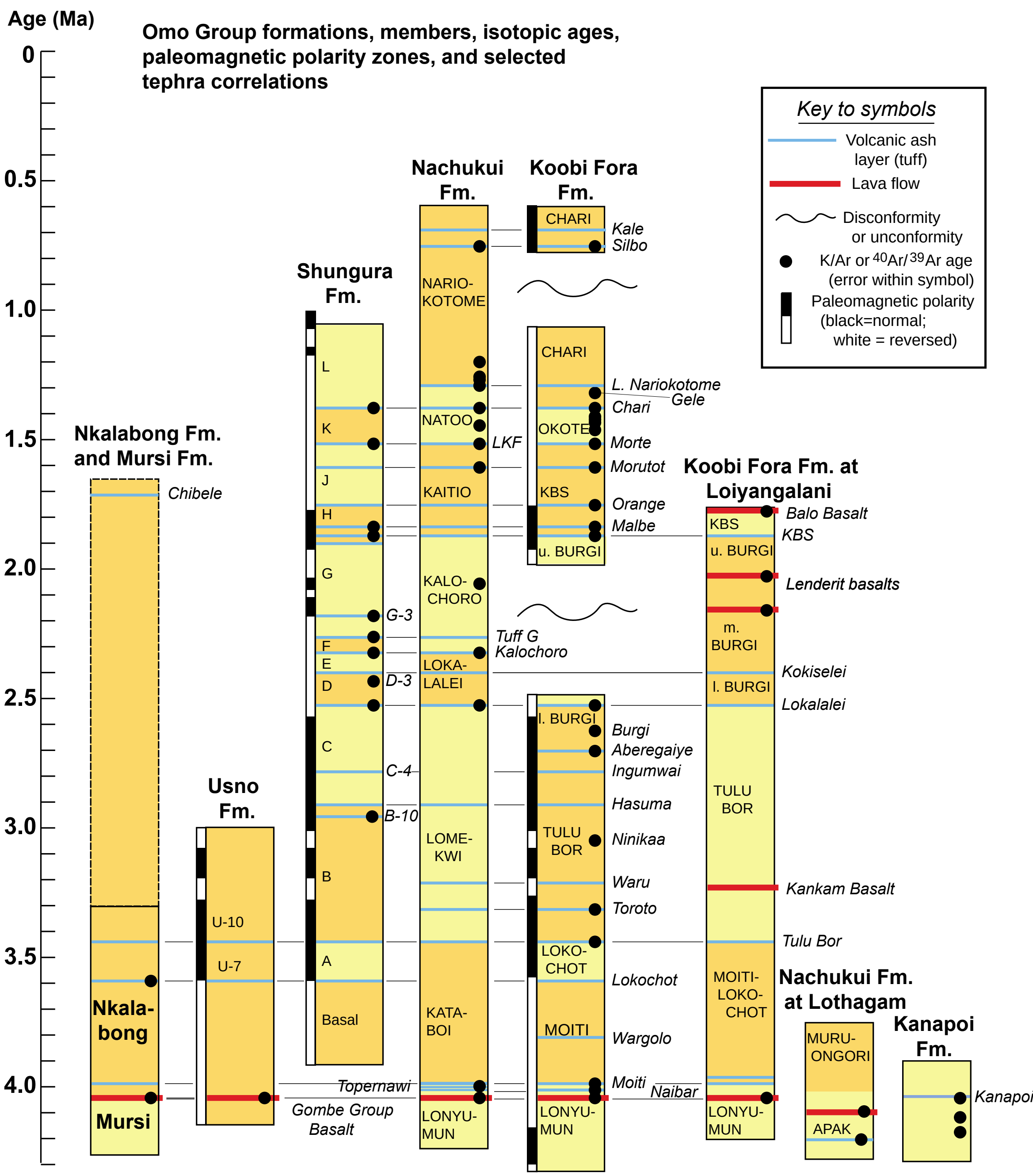

Figure 3. Correlated units of the Pliocene to Pleistocene Omo and Turkana Group in selected regions of the Omo-Turkana Basin. Yellow and orange are used to highlight superjacent member-level stratigraphic units; red lines show intercalated basalt flows; blue lines represent correlated rhyolitic tuff beds; and filled black circles indicate levels of samples for which isotopic ages have been determined.

Table 1. $^{40}Ar/^{39}Ar$ ages of the Omo Group units in the Omo-Turkana Basin. Dating was conducted on anorthoclase. See text for further explanation.

Stratigraphic Unit	Age and Standard Deviation (Ma)*
Silbo Tuff	0.751 ± 0.022 [1]
U. Nariokotome Tuff	1.230 ± 0.020 [1]
M. Nariokotome Tuff	1.277 ± 0.032 [1]
L. Nariokotome Tuff	1.298 ± 0.025 [1]
Gele Tuff	1.326 ± 0.019 [1]
Chari Tuff	1.383 ± 0.028 [1]
Ebei Tuff	1.475 ± 0.029 [1]
Karari Blue Tuff	1.479 ± 0.016 [1]
Koobi Fora Tuff	1.485 ± 0.014 [1]
Lower Koobi Fora Tuff	1.476 ± 0.013 [1]
Morte Tuff	1.510 ± 0.016 [1]
Lower Ileret	1.527 ± 0.014 [1]
Tuff K	1.526 ± 0.015 [2]
Morutot Tuff	1.607 ± 0.019 [1]
Orange Tuff	1.760 ± 0.026 [2]
Malbe Tuff	1.843 ± 0.023 [1]
KBS Tuff	1.869 ± 0.021 [1]
Kang'aki Tuff	2.063 ± 0.032 [3]
Tuff G-3	2.188 ± 0.036 [3]
Tuff G	2.271 ± 0.041 [2]
Kalochoro Tuff	2.331 ± 0.015 [3]
Tuff F	2.324 ± 0.020 [3]
Tuff D3-2	2.443 ± 0.048 [3]
Lokalalei Tuff	2.526 ± 0.025 [3]
Burgi Tuff	2.630 ± 0.017 [2]
Aberegaiye Tuff	2.70 ± 0.02 [4]
Tuff B-10	2.965 ± 0.014 [3]
Ninikaa Tuff	3.066 ± 0.017 [3]
Toroto Tuff	3.308 ± 0.022 [3]
Tuff B-δ	3.419 ± 0.036 [2]
α-Tulu Bor Tuff	3.438 ± 0.023 [3]
Lokochot Tuff	3.596 ± 0.045 [3]
Moiti Tuff	3.970 ± 0.032 [3]
Topernawi Tuff	3.987 ± 0.025 [3]
Naibar Tuff	4.02 ± 0.04 [2]
Kisimei Tuff	4.06 ± 0.02 [4]
Kanapoi Tuff	4.108 ± 0.029 [5]

Table 1 continued.

Stratigraphic Unit	Age and Standard Deviation (Ma)*
Kanapoi ‘upper pumiceous bed’	4.147 ± 0.019 [5]
Kanapoi ‘lower pumiceous bed’	4.195 ± 0.033 [5]
Apak Member ‘pumice clasts’	4.244 ± 0.042 [6]

Ages calculated on anorthoclase where uncertainties in arithmetic mean ages are the standard deviation of the population relative to a reference age of 28.10 Ma for the Fish Canyon Tuff sanidine fluence monitor.

1) McDougall, I. et al. (2006)
2) McDougall, I. et al. (2012)
3) McDougall, I. and Brown, F.H. (2008)
4) Kidney, C. (2012)
5) Leakey, M. G., et al. (1998)
6) McDougall, I., and Feibel, C.S., (1999, 2003)

Table 2. A list of Pliocene and Pleistocene magnetostratigraphic units along with age estimates of the boundaries from Horng et al. (2002), Kidane et al. (2007), and Gradstein et al. (2004).

Designation	Alternate Name	Age (Ma)
C1n	Brunhes Chron	0.000–0.781
C1r	Matuyama Chron	0.781–2.581
C1r.1n	Jaramillo Normal Subchron	0.988–1.072
C1r.2n	Cobb Mt. Normal Subchron	1.173–1.185
C2n	Olduvai Normal Subchron	1.778–1.945
C2r.1n	Reunion II Normal Subchron	2.060–2.080
C2r.2n	Reunion I Normal Subchron	2.150–2.200
C2An.1n and C2An.3n	Gauss Chron	2.581–3.032
C2An.1r	Kaena Reversed Subchron	3.032–3.116
C2An.2n		3.116–3.207
C2An.2r	Mammoth Reversed Subchron	3.207–3.330
C2An.3n		3.330–3.596
C3r	Gilbert Chron	3.596–6.073
C3n.1n	Cochiti Normal Subchron	4.187–4.300
C3n.2n	Nunivak Normal Subchron	4.493–4.631
C3n.3n	Sidufjall Normal Subchron	4.799–4.896
C3n.4n	Thvera Normal Subchron	4.997–5.235

Table 3. A list of published mammalian genera (excluding orders Lagomorpha, and Rodentia) from Pliocene and Pleistocene strata in the Omo-Turkana Basin.

Order	Family	Subfamily	Tribe	*Genus*
Primates				
	Cercopithecidae			
		Cercopithecinae	Cercopithecini	*Cercopithecoides**
				Cercopithecus
			Papionini	*Parapapio**
				Theropithecus
		Colobinae	Colobini	*Paracolobus**
				*Rhinocolobus**
	Galagidae			*Galago*
	Hominidae			
		Homininae	Hominini	*Australopithecus**
				Homo
				*Kenyanthropus**
				*Paranthropus**
Carnivora				
	Canidae			
		Caninae	Canini	*Canis*
				Lycaon
	Felidae			
		Felinae		*Felis*
				(?) *Leptailurus*
				(?) *Caracal*
		Machairodontinae	Dinofelini*	*Dinofelis**
			Homotherini*	*Homotherium*
		Pantherinae		*Panthera*
	Herpestidae			*Helogale*
				Mungos
	Hyaenidae			
		Hyaeninae	Hyaenini	*Crocuta*
				Hyaena
				cf. *Hyaenictus**
				*Ikelohyaena**
				*Pachycrocuta**
			Ictitheriini	*Ictitherium**
	Viverridae			
		Viverrinae		*Civettictis*
				*Pseudocivetta**
				Viverra
	Mustelidae			
		Lutrinae		Gen. indet.

Table 3 continued

Order	Family	Subfamily	Tribe	*Genus*
Proboscidea				
	Deinotheriidae*			
		Deinotheriinae*		*Deinotherium**
	Elephantidae			
		Elephantinae	Elephantini	*Elephas*
				Loxodonta
		Stegotetrabelodontinae		*Stegotetrabelodon**
	Gomphotheriidae*			*Anancus**
Perissodactyla				
	Chalicotheriidae*			*Ancylotherium**
	Equidae			*Equus*
				*Eurygnathohippus**
				*Hipparion**
	Rhinocerotidae			
		Rhinocerotinae	Ceratotheriini	*Ceratotherium*
			Dicerotini	*Diceros*
Artiodactyla				
	Bovidae			
		Aepycerotinae	Aepycerotini	*Aepyceros*
		Alcelaphinae	Alcelaphini	*Alcelaphus*
				Beatragus
				Connochaetes
				*Damalacra**
				Damaliscus
				*Megalotragus**
				*Parmularius**
		Antilopinae	Antilopini	*Antidorcas*
				Antilope
				Gazella
			Neotragini	*Madoqua*
				Raphicerus
		Bovinae	Bovini	*Pelorovis**
				*Simatherium**
				Syncerus
				*Ugandax**
			Boselaphini	*Tragoportax*
			Tragelaphini	*Tragelaphus*
		Caprinae	Caprini	*Capra*
		Cephalophinae		*Cephalophus*
		Hippotraginae	Hippotragini	*Hippotragus*
				Oryx

Table 3 continued

Order	Family	Subfamily	Tribe	*Genus*
Artiodactyla				
	Bovidae			
			Reduncini	*Menelikia**
				Redunca
	Camelidae			*Camelus*
	Giraffidae			*Giraffa*
				*Sivatherium**
	Hippopotamidae			
		Hippopotaminae		*Hexaprotodon**
				Hippopotamus
	Suidae			
		Suinae	Potamochoerini	*Kolpochoerus**
				Potamochoerus
			Phacochoerini	*Metridiochoerus**
				Phacochoerus
		Tetraconodontinae*	Nyanzachoerini*	*Notochoerus**
				*Nyanzachoerus**

* Extinct taxonomic group

Alemseged, 2003; Harris and Leakey, 2003; Leakey and Harris, 2003; deMenocal, 2004; Jablonski and Leakey, 2008).

Objectives

Climate and tectonics are among the main factors believed to have significantly changed habitats in eastern Africa where many mammals including humans evolved at least during the last five million years, but the contributions of each factor are poorly understood (Maslin and Christensen, 2007). The primary goal of this dissertation is to isolate climatic signals from those of tectonics based on the sedimentary record. For this purpose, the Omo Group (de Heinzelin, 1983; Owen and Renaut, 1986; Harris et al., 1988a, b; Ward et al., 1999; Harris and Leakey, 2003; Leakey and Harris, 2003; Gathogo et al., 2008) is used as an example because of its exceptional mammalian fossil record, including early humans (Leakey and Leakey, 1978; Feibel et al., 1989; Wood, 1991) and well-dated strata (Brown and McDougall, 2011), that record both climatic (Brown, 1995; de Menocal, 1995; Lepre et al., 2007; Brown and Fuller, 2008) and tectonic signals.

Three objectives were set for this dissertation: i) to evaluate the effects of tectonic movements on depositional environments by accounting for faulting, subsidence, uplift, and basaltic volcanism; ii) to investigate the relative magnitude of climatic signals as indicated by changes in depositional environments in relation to climatic patterns recorded in the deep sea, particularly the Mediterranean, in response to orbital forcing or Milankovitch cycles (Hays et al., 1976; Kroon et al., 1998; deMenocal, 2004); and iii) to develop a conceptual geologic model that incorporates both rift tectonics and climate in explaining the development of the Omo Group sequence and its fossil record.

Methodology

Field data used in this dissertation come from work done in the Omo-Turkana Basin between 1997 and 2012 including findings from studies conducted for my master's thesis (Gathogo, 2003). Field methods and laboratory procedures used in this study were consistent with those utilized in recent studies in the Omo-Turkana Basin as described by Gathogo (2003). Stratigraphic sections were described and geological mapping was conducted using standard procedures (Compton, 1962). The stratigraphic nomenclature of Brown and Feibel (1986) is utilized with the addition of a middle Burgi Member from Gathogo and Brown (2007) in order to be consistent with other recent studies in the basin (Gathogo and Brown, 2006a, b; Gathogo et al., 2008). Geologic and stratigraphic terminology used in this dissertation follows the North American Stratigraphic Code (2005).

Motivation for this study came from work that began in the Kanapoi area south of Lake Turkana (Fig. 2) following an invitation from M. G. Leakey who led a paleontological research team to that site from the National Museums of Kenya. The visit to the Kanapoi area resulted in further detailed studies in other parts of the Omo-Turkana Basin where stratigraphic descriptions, interpretation of geologic settings, and mapping have since been published, including areas to the west (Leakey et al., 2001; Brown and Gathogo, 2002), southwest (Brown and Gathogo, 2005), southeast (Gathogo et al., 2008), and east (Gathogo and Brown, 2006a, b; Leakey et al., 2008) of Lake Turkana. The Lower Omo Valley in the northern part of the lake has also been visited, and features of its local geology are included in recent basin-scale studies (Haileab et al., 2004).

Fieldwork specific to this dissertation was designed to allow the investigation of strata from similar time intervals across the basin, particularly units that show pronounced lateral variation in depositional settings and thickness. The tasks at each location included: i) a detailed study of the sedimentary sequence for local and regional correlation by identifying marker units; ii) the interpretation of depositional settings, including local ephemeral streams and alluvial systems from the basin margin, perennial streams of the axial fluvial system, and lake sequences; iii) an investigation of the roles of local geomorphology and possible tectonic activities, with the latter being recognized by the presence of deformational features such as faults and angular disconformities or association with basalt flows or intrusions; and iv) an analysis of climate-related events in the sequence by looking for consistency in the placement of strata of possible climatic significance into the well-known cycles of varying length caused by orbital variations in the absence of tectonic events.

The geology of selected areas is emphasized for specific time intervals to best illustrate characteristic sedimentary features related to climatic or tectonic events. Principal depositional environments represented by the Pliocene and Pleistocene sequence in the Omo-Turkana Basin are assigned in the following sections primarily based on field observations and supplemented by reference to published data by other workers including reports of recent analogs in the basin (Butzer, 1971; Cerling, 1977; Butzer, 1980) and interpretations of the geologic record in the Omo Group deposits (Bowen, 1974; Powers, 1980; Harris et al., 1988a, b; Brown and Feibel, 1991; Harris and Leakey, 2003; Leakey and Harris, 2003). A synopsis of the major depositional facies recognized in the basin by studies cited above is presented below.

Fluvial facies are characterized by fining-upward sequences that grade from channel-form sandstones to mudstones and commonly end with silty claystones. Common pedogenic features include carbonate concretions, rhizoliths, and slickenside fractures. Fluvial units associated with the axial system (Omo River) commonly show features of an extensive perennial system, including the occurrence of the African freshwater oyster *Etheria elliptica* (de Heinzelin, 1983; Brown and Feibel, 1986; Harris et al., 1988a, b; Gathogo et al., 2008). This type of oyster lives in perennial rivers (Yonge, 1962; Adam, 1986) including the Omo and Nile as well as in stable freshwater lakes (Abell et al., 1995) such as Lake Victoria. Fluvial and lacustrine varieties of oysters in the Omo-Turkana Basin can be distinguished based on shell morphology (Gathogo, 2003). Some of these perennial fluvial systems contain geochemically characterized tuffaceous deposits and can be traced for hundreds of kilometers within the basin where they can be (and have been) dated if they also contain feldspar-bearing pumices. Local ephemeral fluvial systems from the basin margins are typically braided, poor in clay, and associated with prominent vertical incision. These ephemeral channel deposits commonly grade to conglomerates of alluvial fans proximal to the basin margin. Further subdivisions of the fluvial facies in the basin include channel, levee, floodplain, and swamp deposits (Butzer, 1971, de Heinzelin, 1983; Feibel and Brown, 1986; Gathogo, 2003).

Deltaic deposits are better sorted than the fluvial facies described above and mostly comprise thinly bedded silts and sands. These rocks are identified in many parts of the basin where they commonly exhibit low-angle crossbedding and coarsening-upward sequences. Interdistributary channels (Butzer, 1971; de Heinzelin, 1983) and distributary channels (Feibel and Brown, 1986) are high-energy lithofacies of deltaic settings in the Omo-Turkana sequence. Low-energy delta fringe deposits in mudflats and lagoons have also been described, and these are principally composed of reduced clays with isolated slickenside fractures (Haesaerts et al., 1983). Fine-grained deposits in the prodelta or on the delta front are characterized by planar laminations and thin bedding that represent detritus that settled from suspension, and in some instances preserve evidence of bioturbation.

Lacustrine deposits in the Omo-Turkana sequence are primarily identified based on the presence of aquatic invertebrates, including gastropods, bivalves, ostracods, algal stromatolites, and diatoms. Other common features include laterally consistent strata such as planar clayey to silty laminations and thin beds. Pelagic lacustrine settings (Gathogo, 2003) are characterized by low-energy features such as clay-rich silts that normally reflect reduced environments (grayish olive to green), and are locally associated with diatomaceous silts (Butzer, 1971; Findlater, 1976). The lacustrine variety of *Etheria elliptica* is used as an indicator of a stable freshwater lake (Gathogo, 2003). Shallow water and marginal lacustrine deposits are characterized by sandy to pebbly deposits that are crossbedded or planar bedded and commonly contain abundant mollusks, ostracods, or algal stromatolites. Other diagnostic features of shallow lake environments in the basin include trace fossils left by the activities of mollusks (e.g., *Lockeia*) and fish (e.g., *Piscichnus*) (Feibel, 1987; Harris et al., 1988a, b; Ekdale et al., 1989; Bromley, 1996; Lamond and Ekdale, 2001). Gathogo (2003) discusses some of the paleoenvironmental implications of invertebrates and trace fossils in marginal lacustrine settings of the Omo Group deposits. Studies of modern stromatolites from Lake Tanganyika along the western branch of the East Africa Rift System have morphological features that can be used to estimate depth ranges of lacustrine sequences (Cohen et al., 1997).

Modern Lake Turkana generally represents pelagic lacustrine settings such as those described above (e.g., Butzer, 1971; Findlater, 1976) because it accumulates fine layers of olive clays and diatomite at least in its southern part (Yuretich, 1979). Strata characteristic of pelagic lake settings reflect dominantly vertical settling of fine sediments from the water column even at shallow water depths (e.g., 10 m) regardless of the water chemistry (Gathogo, 2013). Therefore, modern Lake Turkana and its Holocene precursor fit the definition of a pelagic lake regardless of the sharp differences in water chemistry (from fresh to alkaline) and water depth (a drop of ~80 m from the Holocene highstand to the current level). Evidence for differences between the Holocene and current lake includes the presence of lacustrine *Etheria elliptica* that indicates a stable freshwater lake (see Gathogo, 2003) in shoreline deposits ~80 m above the current lake levels. These changes have taken place over a very short geological time interval (less than a precession cycle) when compared to the entire geological history of the lake (at least 4 Ma; discussed below). This is why many workers combine the modern lake and its Holocene precursor in geological studies of the basin (e.g., Fig. 3 of Feibel, 2011). However, other studies of Pliocene and Pleistocene lacustrine sequences in eastern Africa (including those in the Omo-Turkana Basin) consider water chemistry and depth as important parameters (Maslin et al., 2014). For example, Trauth et al. (2007) classify ancient lacustrine sequences on the basis of the ecological nature of diatoms and associated diatomite. The classification defines two types of paleolakes: deep freshwater lakes, and alkaline lakes. In deep freshwater paleolakes, diatoms are dominated by planktonic freshwater assemblages that form nearly pure, white diatomites that preserve laminations in many exposures. In contrast, alkaline lakes are shallow and diatom assemblages are composed of benthic–epiphytic alkaline water types that form diatomites with a significant clastic component.

The classification of lacustrine sequences described by Trauth et al. (2007) and adopted in other recent studies (e.g., Maslin et al., 2014) implies that modern Lake Turkana and its Holocene precursor fit in separate classes. Because the chemistry of lakes can change over short time intervals, it is impractical to use this classification in the study of earlier Pleistocene and Pliocene lacustrine strata. Primarily for this reason, Trauth's classification is not adopted in this dissertation. A secondary reason relates to additional distinctions that are made in classifying ancient lakes. For example, Maslin et al. (2014) propose that a deep freshwater lake should have a size of several 100 km^2 and water depths in excess of 150 m. Most modern lakes in the eastern branch of the East African Rift System do not meet these criteria because they are either too small (e.g., Lake Nakuru, 56 km^2; Lake Bogoria, 38 km^2) or too shallow (5–50 m water depth; e.g., Lake Baringo, Lake Naivasha, Lake Elmenteita), including the freshwater ones, or have more or less saline characteristics (Tiercelin and Lezzar, 2002; Lake Magadi, Lake Elmenteita, Lake Bogoria, Lake Turkana). Many modern lakes along the western branch of the East African Rift System are deeper and freshwater (Tiercelin and Lezzar, 2002), and these do meet the criteria of deep freshwater lakes as described by Maslin et al. (2014). Their criteria are also difficult to apply in practice in geological studies of the Pliocene and Pleistocene lacustrine sequences of the Omo-Turkana Basin. Perhaps this why Maslin et al. (2014; their Fig. 3) and associated studies misclassify the largest freshwater paleolake (Lonyumun Lake) in the Omo Group sequence as a shallow alkaline lake but identify subsequent smaller paleolakes (mostly Pleistocene) as large deep lakes.

CHAPTER 2

FIELD OBSERVATIONS, DATA ANALYSIS, AND INTERPRETATIONS

This chapter focuses on the first two objectives outlined in Chapter 1: to report the evidence for and depositional implications of tectonic activity such as faulting, subsidence, relative uplift, and basalt volcanism, and to assess climatic signals, particularly with respect to astronomical forcing. For this purpose, the Omo Group sequence is described in eight parts, representing time intervals defined by members of the Koobi Fora Formation (Brown and Feibel, 1986). Each part reviews current knowledge about depositional settings within the temporal interval and emphasizes new findings about syndepositional tectonic features. Examples of such features include faults and disconformity surfaces that laterally transition into angular unconformities or coincide with basalt intrusions in various parts of the basin. References are made to climate proxies records in the Mediterranean region that are influenced by mainly climatic factors, including the monsoon rains, shared with the study area at present and even since at least the early Pliocene (Rossignol-Strick, 1985; deMenocal, 1995; Lourens et al., 1996; Emeis et al., 1998, 2000).

Results summarized in Table 4 reveal that deposition of the Omo Group has been primarily controlled by tectonic-related events since the early Pliocene. For example, the most prominent lake sequences, those of Lakes Lonyumun (~4.0 Ma), Lokochot (~3.5 Ma), Waru (~3.2 Ma), Lorenyang (~2.0 Ma), and Nachukui (~1.2 Ma), coincide temporally with tectonic events in the basin. These lacustrine intervals do not occur at 400 kyr eccentricity maxima in climatic cycles as proposed by Trauth et al. (2005, 2007, 2008). Eccentricity maxima and the age of pelagic lake sequences in the basin are shown in Figure 4. The precessional cycle (about 20 kyr) is likely the most prominent climate-related signal in the Omo Group, and may correspond to minor fluctuations in lake level, and may also be related to deposition of cyclic units in fluvial parts of the sequence that range in thickness from 5 to 10 m.

Observations presented here indicate that onset of deposition of the Omo Group sequence coincided with structural movements that included tectonic inversion and reactivation of major structures (Bruhn et al., 2010). Some of these structures are the Balo Fault near Loiyangalani (southeast), the Kosia Fault (part of the Lokichar Fault; southwest), and the Lomekwi Fault (west). Depositional patterns in the Omo Group continued to be largely controlled by subsequent tectonic events. Each episode began with events that included faulting, the development of alluvial fans and prominent ephemeral channels, thick lake sequences, and basaltic volcanism. The volcanism includes the Gombe Group basalts (about 4.0 Ma), the Kankam Basalt (3.2-3.3 Ma), the Lenderit basalts (2.02 ± 0.02 and 2.18 ± 0.02 Ma), and the Balo Basalt (1.79 ± 0.02 Ma) (Gathogo et al., 2008). There are no known volcanic eruptions associated with tectonic episodes at ~3.5 Ma, ~1.7 Ma, ~1.2 Ma, or 0.3 Ma.

Table 4. A summary of characteristic geological features of the Omo Group sequence in the Omo-Turkana Basin.

Member-level Geologic Intervals	Principal Depositional Environments	Structural Settings and Basalt Volcanism
Lonyumun Mb. and correlative units; ca. 4.4 Ma to 3.97 Ma	Begins with a shallow lake and local alluvial fans near high topography along the basin margins; transition to major rivers mostly from the northern and southern parts of the basin that deposit clay-rich mud; silt- to sand-rich deltaic conditions pave way to the first major basin-wide lake (Lonyumun Lake)	East-west basin extension, tectonic subsidence, and faulting initiate local and basinwide sedimentation; ends with Gombe Group basalts (ca. 4.0 Ma)
Moiti Mb. and correlative units; 3.97 Ma to 3.60 Ma	Begins with deposition of an airfall volcanic ash (Moiti Tuff) in the Lonyumun Lake; the first major axial and perennial fluvial system that represents the proto-Omo River is established shortly thereafter; the river flows from northern parts of the basin as meander channels that rework the Moiti Tuff and pumices, and deposit clay-rich sediments in local expansive floodplains in many parts of the basin	No significant structural events, with subsidence rates being nearly matched by sedimentation rates; results in a balanced-fill fluvial basin (see text)
Lokochot Mb. and correlative units; 3.60 Ma to 3.44 Ma	Begins with an airfall volcanic ash (Lokochot Tuff) deposited on expansive flood plains of the axial and perennial fluvial system (proto-Omo River); the ash is reworked by a perennial proto-Omo River in the same manner as the Moiti Tuff; significant changes in local depositional settings occur ca. 3.5 Ma in many parts of the basin where alluvial fans and ephemeral channels from the basin margin become common in some area, while a major lake (Lokochot Lake) occupies other areas; sedimentation is significantly reduced in many areas, erosion is common in some areas, and shallow lakes occur in other areas	Major tectonic readjustment of the basin occurs at ca. 3.5 Ma with reactivation of border faults along many parts of the basin margin; results in significant longitudinal subsidence to form a narrow depositional trough, with rapid flooding along the fault scarps developing into lakes

Table 4 (continued).

Member-level Geologic Intervals	Principal Depositional Environments	Structural Settings and Basalt Volcanism
Tulu Bor Mb. and correlative units; 3.44 Ma to 2.62 Ma	Begins with widespread deposition by the axial fluvial system; transitions to alluvial fans and channels from basin margins, followed by a pelagic lake (ca. 3.2 Ma) in some places, shallow lake in many locations; sedimentation reduced significantly in some places or much of local section is eroded, but the axial fluvial system of proto-Omo River resumes and persists to the end	Major tectonic changes 3.3-3.2 Ma with reactivation of major structures, including border faults along basin margins and eruption of the Kankam Basalt (ca. 3.2 Ma); results in significant axial subsidence and relative uplift in many regions of the basin, particularly in the east and south.
Burgi Mb. and correlative units; 2.62 Ma to 1.87 Ma	The axial fluvial system is replaced by alluvial fans and channels from basin margins, followed by a pelagic lake (ca. 2.2 Ma); sedimentation reduced significantly at many locations or much local eroion; deposition by the axial fluvial system resumes about 2.05 Ma ago at a few locations, including along the eastern margin of the basin	Major tectonic activity 2.2-2.0 Ma with reactivation of long faults mostly in the western region and near Loiyangalani; ends with intrusion of Lenderit Basalts (2.02 to 2.18 Ma) and relative uplift in much of the eastern region
KBS Mb. and correlative units; 1.87 Ma to ca. 1.60 Ma	Pelagic lake persists in some areas near the basin center and the western margin; shallow lakes alternate with the axial fluvial system at many locations in the eastern region; sedimentation is reduced significantly in many areas in the southern and eastern regions after 1.75 Ma, when ephemeral channels from the eastern basin margin become common and erode much local section	Moderate tectonic episode between 1.79 and 1.75 Ma begins with extrusion of Balo Basalt (1.79 ± 0.02 Ma); basin tilt towards the central axis results in most deposition near the western margin; relative uplift isolates many locations near the eastern and southern margins
Okote Mb. and correlative units; 1.60 Ma to 1.38 Ma	Begins with with deposition of tuffaceous silty units by the proto-Omo River, some eolian deposits; minor channels from basin margins, and minor playa lakes or submerged deltas near central parts of the basin; the axial fluvial system resumes over much of the eastern region slightly before 1.42 Ma, and hosts local floodplain lakes	No widespread tectonic episode, but minor localized faulting between 1.54 and 1.42 Ma; sediment accumulation shortly before 1.42 Ma limited more by supply from the axial fluvial system than by the availability of accommodation space in the basin

Table 4 (continued).

Member-level Geologic Intervals	Principal Depositional Environments	Structural Settings and Basalt Volcanism
Chari Mb. and correlative units; 1.38 Ma to Middle Pleistocene	Begins with the axial fluvial system, followed by large channels possibly from alluvial fans entering from the basin margins beginning shortly before 1.3 Ma, when shallow lakes begin to form in many parts of the basin; sedimentation continues until 0.75 Ma in the Nachukui Fm., but only until ~1 Ma in the Shungura Fm.; sedimentation is reduced and/or erosion removes section in many locations and a pelagic lake occupies parts of the basin at least before 0.75 Ma; deltaic and fluvial settings of the axial fluvial system resume at some locations shortly before 0.75 Ma, with lacustrine deposition in the Nachukui Fm. and Koobi Fora Fm.; interruption by shallow lakes marks the final depositional stage of the Omo Group before the beginning of the Late Pleistocene	Record poorly understood due to outcrop limitation, but involves substantial tectonic episodes: one between 1.3 and 1.2 Ma; another before 0.7, and a third at 0.3 Ma before deposition of the Kibish Formation (Turkana Group) began in the Ethiopian part of the basin

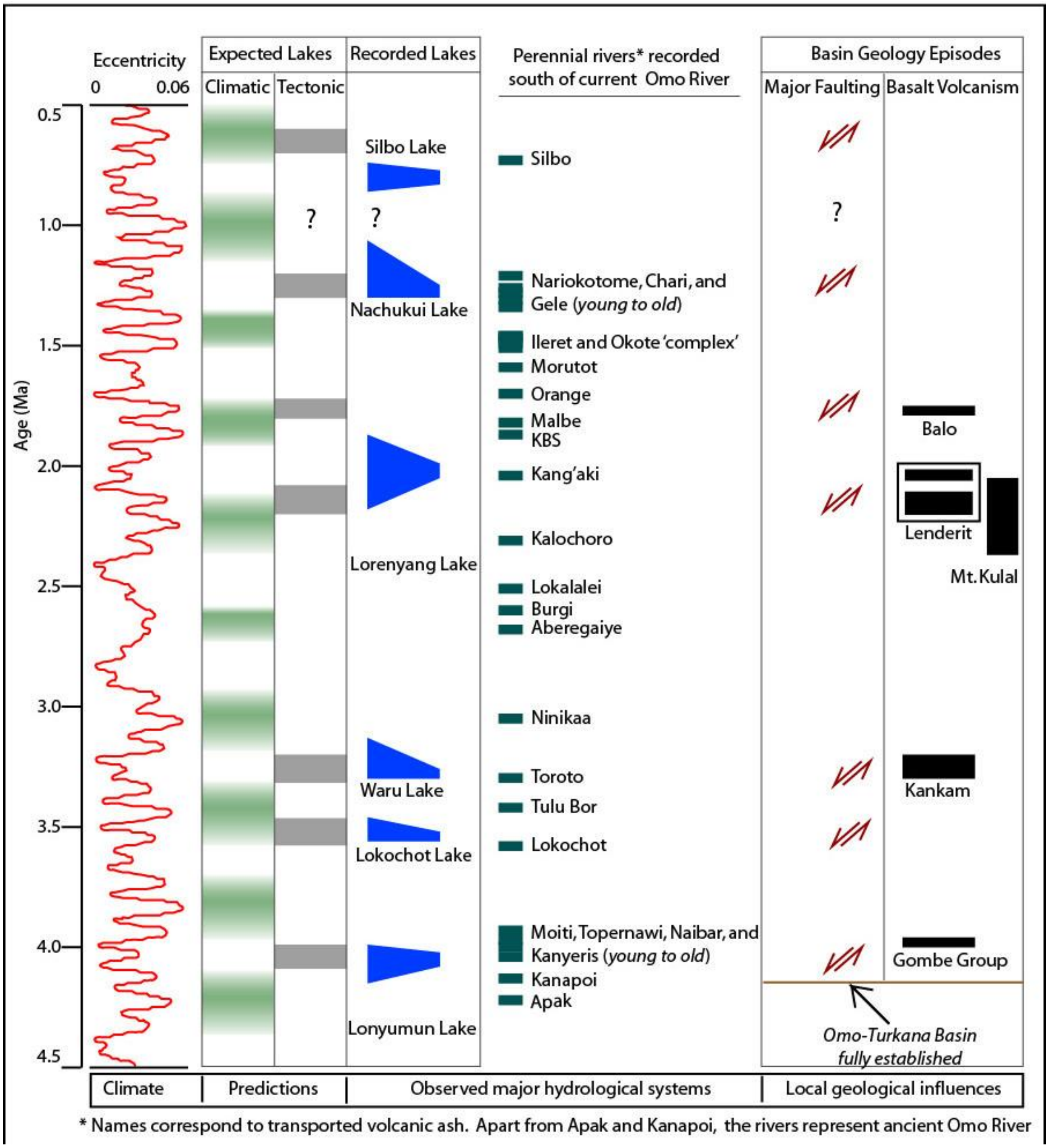

Figure 4. Summary diagram showing the timing of global eccentricity cycles (eccentricity) as well as pelagic lakes and tectonic episodes in the Omo-Turkana Basin. Also shown are predictions about when climatic and tectonic lakes could have occurred in the basin. Lacustrine sequences that are developed in only one part of the basin and associated with expansive floodplain lakes in other parts of the basin are not included in this diagram. An example of a local lake is the Kokiselei Lake of Feibel (2011). Evidence for the perennial nature of the rivers includes the occurrence of the riverine oyster *Etheria elliptica*.

Field observations and interpretations of the depositional environments in the Omo-Group sequence as presented here are in agreement with those of most previous studies (Bowen, 1974; Powers, 1980; de Heinzelin, 1983; Brown and Feibel, 1986, 1991; Harris et al., 1988a, b; Harris and Leakey, 2003; Leakey and Harris, 2003). New information on the geology of this basin comes from field observations that relate syndepositional tectonic features such as faulting and associated basalt intrusions and extrusions with depositional effects, including the development of lacustrine and alluvial fan settings. These new findings have significant implications for the evolution of climatic influences on regional depositional settings where many mammals and hominins evolved during the Pliocene and Pleistocene. Interpretations presented in this dissertation linking climatic cycles and development of major lacustrine sequences in the basin contrast with recent studies (Trauth et al., 2005, 2007, 2008, and related citations; Behrensmeyer, 2006; van Bocxlaer, 2008), with tectonics viewed here as being the overriding factor. This dissertation also reinforces the geologic significance of some structural observations made in previous studies (Patterson et al., 1970; Brown and de Heinzelin, 1983; Dunkleman et al., 1988; Harris et al., 1988a, b), examples of which include major disconformities and a change in dip between major units. The following sections focus on the description and interpretation of new findings that relate tectonic activity in the basin to changes in depositional environments. The depositional settings are also put in context with climatic patterns from astronomical forcing as recorded in the Mediterranean Sea.

Proxy measures of climatic conditions in deep sea cores include terrestrial dust peaks (Brown, 1995; deMenocal, 1995), organic-rich layers (sapropels) (Lourens et al., 1996), oxygen isotope profiles in calcareous invertebrates (Lourens et al., 1996; Liesecki and Raymo, 2005), sea surface temperature estimates (Lourens et al., 1996), and astronomical solutions of two of the Earth's orbital time series: precession and obliquity (Laskar, 1990; Lourens et al., 1996; Paillard et al., 1996). The remote climatic record is used in this study to evaluate whether climatic changes coincided with fluctuations in deposition in the study area, or whether substantial control on deposition is affected by tectonic processes.

Lonyumun Member Time: ca. 4.3–3.97 Ma

New findings from this dissertation include field evidence for pronounced rift-related structural movements that coincide with significant changes in sedimentation rates, lithological patterns in depositional settings, and basaltic volcanism. Lacustrine deposits of the Lonyumun Lake (ca. 4.0 Ma) dominate the outcrop area (Harris et al., 1988a, b; Brown and Feibel, 1991; Leakey and Harris, 2003; Haileab et al., 2004; Bruhn et al., 2010; Feibel, 2011) of the Lonyumun Member interval (up to 110 m of measured section, at Mursi; de Heinzelin and Haesaerts, 1983) and possibly represent the largest lake that has existed in the basin since the Early Pliocene. These findings challenge the role of climate as the overriding factor in controlling major depositional settings in the basin including the occurrence of the main lacustrine sequence of the Lonyumun Lake (Trauth et al., 2005, 2007, 2009; Maslin et al., 2014).

Detailed geologic studies of deposits in the Lonyumun Member time interval in the Omo-Turkana Basin began earlier than those for intervals of the Omo Group sequence after the discovery of early hominins and other mammals in the Kenyan part of the Omo-

Turkana Basin (Patterson, 1966; Patterson et al., 1970). These deposits are well known for their mammalian fossil assemblage (Coppens, 1972; Howell et al., 1983; Harris et al., 1988a, b; Leakey et al., 1995; Leakey and Harris, 2003) with representation of pioneer genera that include one of the earliest bipedal hominins *Australopithecus anamensis* and other mammals that become common throughout the region in the course of Pliocene and Pleistocene time. Richly fossiliferous strata are exposed in isolated outcrops in the northern and southern margins of the basin where fluvial deposition occurred. Although strata exposed in these outcrops have received the most attention, they account for less than one third of the composite section in this interval.

Depositional settings recognized by various workers (de Heinzelin and Haesaerts, 1983; Brown and Feibel, 1986; Feibel and Brown, 1986; Harris et al., 1988a, b; Leakey and Harris, 2003; Gathogo and Brown, 2006; Gathogo et al., 2008) in strata of the Lonyumun Member allow division into three parts: i) a basal part characterized by high-energy deposits that represent localized depositional settings including alluvial fans and ephemeral channels from nearby high terrain where Miocene volcanic rocks and, in some places, metamorphic rocks defined the basin margin; ii) a middle part in which moderate- to low-energy deposits typical of deltaic distributary channels transition through marginal lacustrine to pelagic lacustrine facies corresponding to Lonyumun Lake; and iii) an upper part composed of moderate to moderately high-energy deposits interpreted as distributary channels of a prograding delta, graded to the main axial fluvial system of the perennial proto- Omo River. The composite section and outcrop area of the Lonyumun Member is dominated by the lower part.

Here it is proposed that accumulation of the Lonyumun Member time sequence was initiated by rift-related structural events that led to accentuated vertical motion along major longitudinal faults. Evidence of activity during such events comes from outcrop traces of major faults alongside high-energy deposits of the lower Lonyumun Member at locations such as Lomekwi and Kosia. The best example of tectonic activity during this time interval comes from the southern part of the basin in the Loiyangalani region where the timing of sediment accumulation during the Lonyumun Member interval coincides with structural inversion along the Balo Fault (Gathogo et al., 2008; Bruhn et al., 2010). Tectonic inversion in the Turkana region during the Early Pliocene has been described by various workers (Morley et al., 1999; LeGall et al., 2005), and reversal of motion along the Balo Fault is a local manifestation of basinwide tectonic events that marked the initial stages of the Omo-Turkana Basin as a coherent depositional area. Lack of older sedimentary or volcanic rocks between the Lonyumun Member and the metamorphic basement east of the Balo Fault indicates that the inversion took place after the Miocene since the western block includes Miocene sediments and basalts (Savage and Williamson, 1978; Brown et al., 2016).

Activity along the Balo Fault in the Loiyangalani region in the Early Pliocene coincides with local sedimentation in other parts of the basin, particularly along major structural features in the western part of the basin, for example, the Lomekwi-Kokiselei and Kosia-Lokichar Faults (Figs. 5 and 6). Depositional settings of the Lonyumun Member at Kosia (this dissertation) share many similar patterns with correlative deposits at Lothagam (Savage and Williamson, 1978; Leakey and Harris, 2003) and Kanapoi (Powers, 1980; Leakey et al., 1998; Harris and Leakey, 2003), and thus suggest common

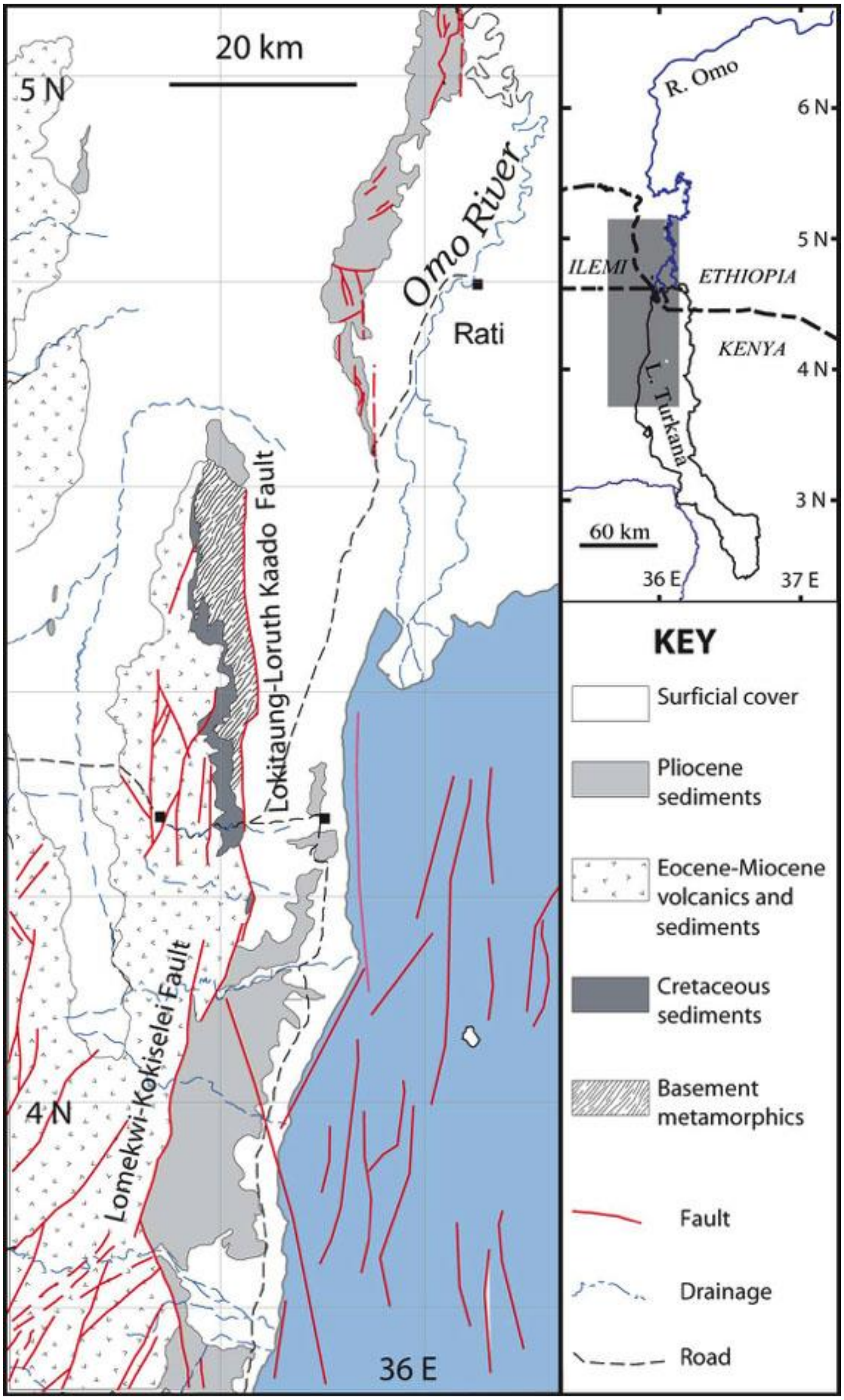

Figure 5. Map showing major faults and outcrop geology in the northern and western parts of the Omo-Turkana Basin. The Pliocene sediments are part of the Omo Group sequence. Some faults are located based on Morley et al. (1999). Location of the detailed map in relation to the larger basin is shown at top right.

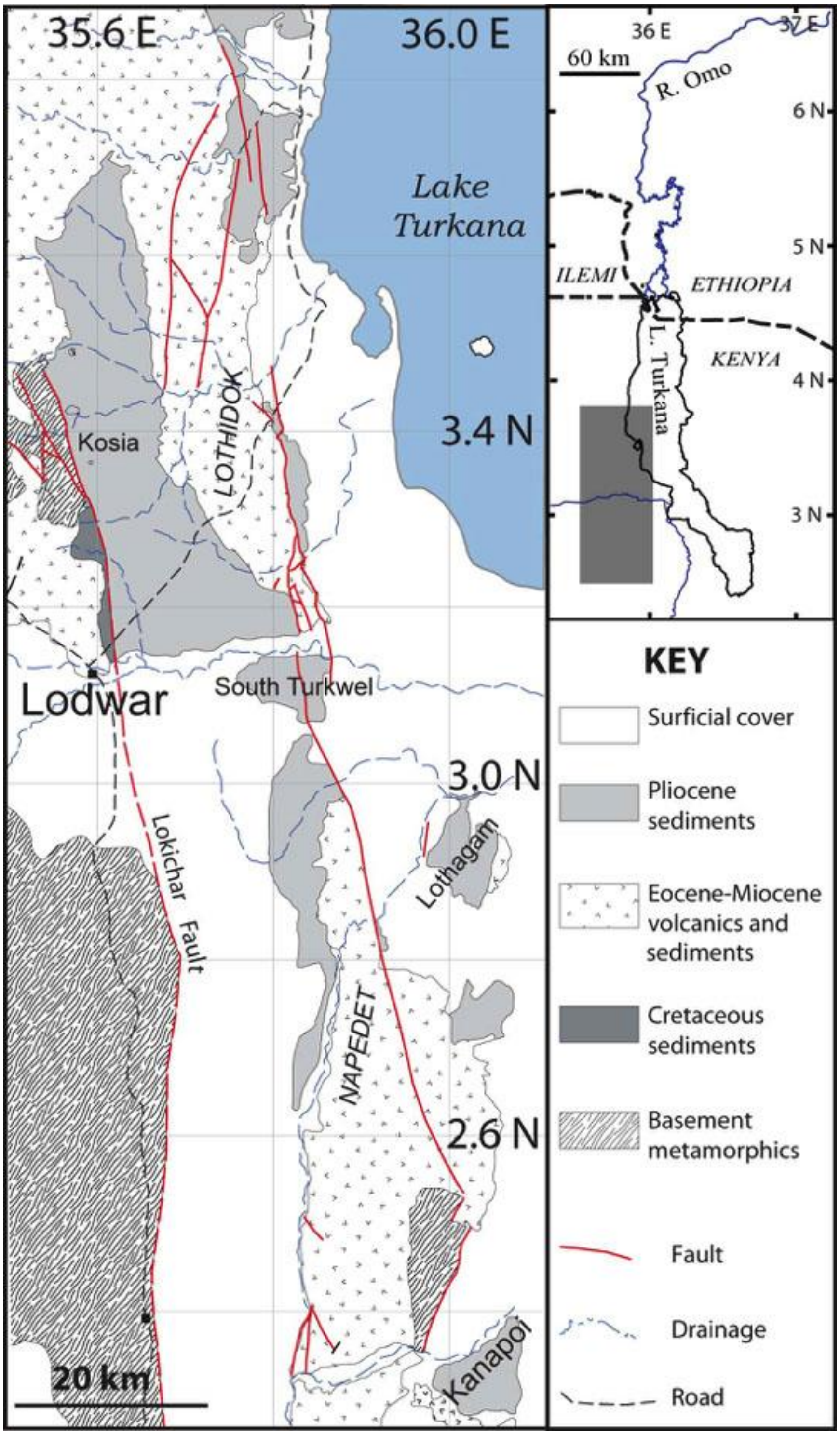

Figure 6. Map showing major faults and outcrop geology in the southwestern and southern parts of the Omo-Turkana Basin. The Pliocene sediments are part of the Omo Group. Some faults are located based on Morley et al. (1999). Location of the detailed map in relation to the larger basin is shown at top right.

controlling factors. Consequently, tectonic activity that triggered sedimentation along the Balo, Kosia-Lokichar, and Lomekwi-Kokiselei Faults may also have been the main cause of increased sedimentation in many areas (e.g., Kanapoi, Lothagam, and eastern Ileret) although outcrop examples of syndepositional fault activity have not been reported for these areas.

Tectonic activity in the Omo-Turkana Basin during the Lonyumun Member interval continued with subsidence along normal faults consistent with east-west extension and relative uplift along the eastern and western margins of the basin where alluvial fans accumulated. For example, in the western part of the Lomekwi drainage, the lower and middle portions of the Lonyumun Member are conglomerates adjacent to the Lomekwi-Kokiselei border fault where Miocene basalts form the landscape to the west. Basaltic pebble conglomerates and thin beds of mollusk-packed sandstones occur near the base and diatomaceous units are interbedded with pebbly sandstones near the top. These strata are interpreted as having been deposited in a rapidly subsiding part of a lake proximal to a steep fault escarpment, with a modern analog being the western margin of the Kokoi basaltic highland along the northeastern shores of Lake Turkana. Lithologic depositional features akin to those just described occur in correlative deposits of the Allia Bay region (Brown and Feibel, 1986). Subtle syndepositional faulting is identified in the lower subinterval deposits farther north in the Lower Omo Valley at Mursi where, for example, beds underlying the Gombe Group basalts are displaced vertically (up to 1 m) by a longitudinal fault (de Heinzelin and Haesaerts, 1983). Tectonic activity in the basin culminated with the formation of an extensive pelagic lake (Lonyumun Lake) at the same time as the Gombe Group basalts (4.0 Ma; Haileab et al., 2004) were intruded into the sedimentary strata and erupted as lava flows. The distribution of these basalts possibly reflects the location of local volcanic centers and related structures such as fracture sets and faults that are responsible for the orientation of basalt dykes at many locations (Gathogo and Brown, 2006; Gathogo et al., 2008). At least two episodes of basaltic eruption are noted because two flows are separated by a thin interval of deltaic to lacustrine deposits in many areas, including on the margin of the Suregei highland (east of Ileret) (Watkins, 1983, 1986), Kokoi (south of Ileret; Gathogo and Brown, 2006), Loiyangalani (Gathogo et al., 2008), and Kataboi (Haileab et al., 2004). These two basalt flows are chemically indistinguishable and were erupted over a brief time interval (Haileab et al., 2004).

The end of tectonic activity after these basalt eruptions records the first evidence for the presence of the perennial fluvial system of the Omo River in many parts of the basin. Deposits of this system began in transitional deltaic settings grading to fluvial settings, particularly at locations where the Gombe Group basalts are absent (Brown and Feibel, 1986; Harris et al., 1988a, b; Gathogo and Brown, 2006). For example, the earliest known occurrence of the freshwater oyster *E. elliptica* (exclusive to perennial rivers) in the Omo Group occurs in western Lomekwi (pers. obs.) in an interval with rhyolitic tuffs from the Omo system that includes the Topernawi Tuff (with pumices) (Harris et al., 1988a, b) and Moiti Tuff channels.

Examination of climatic patterns associated with astronomical forcing (Fig. 7) reveals that the timing of Lonyumun Lake does not coincide with the most probable period for intensified monsoonal precipitation at an eccentricity maximum from 4.25 to 4.21 Ma (Fig. 4) when sapropel layers in the Mediterranean Sea were common and thick.

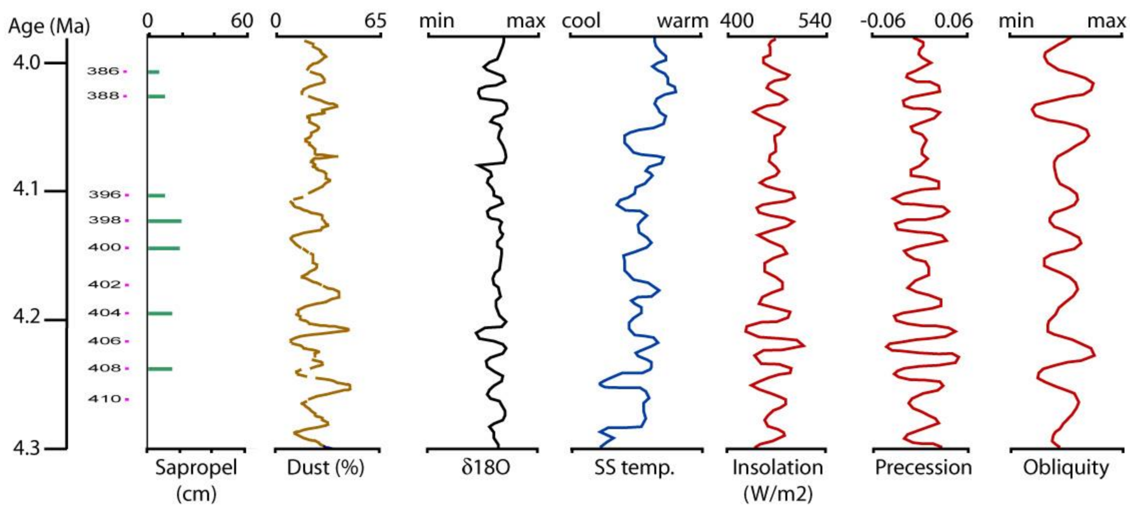

Figure 7. Climatic proxy records for the Lonyumun Member time interval (ca. 4.3–3.97 Ma) based on marine deposits from the Mediterranean Sea. Associated eccentricity cycles and predicted lake occurrence times are shown in Figure 4. See text for data sources and explanation. Numbers shown on the left (to the right of the ages) represent insolation cycles.

In the Omo-Turkana Basin, this period of postulated wet climate coincides with fluvial settings that have been dated about 4.2 Ma (Leakey et al., 1995; McDougall and Feibel, 1999) at Lothagam and Kanapoi, and also in the northeastern part of the Koobi Fora region in Areas 13 and 41 (Kidney, 2012). These observations confirm that depositional settings in the Omo-Turkana Basin were more likely controlled by creation of accommodation space through tectonic movements than by climatic conditions during Lonyumun Member time.

Timing, magnitude, and duration of depositional facies at submember scale, including fluvial cycles, may correspond to precession-scale climatic cycles, but this has yet to be proven. Obliquity forcing of climatic events was minimal in the eastern and northern parts of Africa (Lourens et al., 1996) during the Lonyumun Member interval.

The timing of lacustrine deposits in the Omo-Turkana Basin during Lonyumun Member time does not support correlation of this thick lacustrine sequence (Trauth et al., 2005, 2007, 2009; Maslin et al., 2014) with maxima in eccentricity cycles. Moreover, these authors misclassify the Lonyumun Lake sequence as a shallow alkaline lake (e.g., Fig. 3 of Maslin et al., 2014). Instead, the lake was reasonably fresh (Findlater, 1976a; Cerling, 1976, 1977), is represented by thick laminated sequences of diatomite and olive claystone, and is the most extensive pelagic lacustrine sequence in the basin since the early Pliocene with a probably minimum area of ~21,000 km^2.

Moiti Member Time: 3.97–3.60 Ma

Strata representing the Moiti Member time sequence are up to 60 m thick in Allia Bay (Brown and Feibel, 1986). Mammalian fossils (Feibel et al., 1991) from this time interval are relatively few and come from only a few locations where they are associated with fluvial deposits of the axial perennial system. One find from such settings is a hominin, *A. anamensis* (Leakey et al., 1995). No evidence of major tectonic movements during this time interval has been found in compiling this dissertation. Tectonic subsidence rates in the basin appear to have been closely matched by depositional rates of the axial fluvial system resulting in what is termed here a balanced-fill fluvial basin, where vertical incision by channels is low due to a small gradient. It is likely that the basin was kept full of sediment during this interval, and that the proto-Omo River exited the basin so that no lake formed.

Depositional settings (described above) in the upper part of the underlying sequence and including Lonyumun Lake prevailed in many locations at the time of eruption of the

Moiti Tuff (3.97 Ma). Starting with the interval between airfall and fluvial deposits of the Moiti Tuff, three main depositional settings are recognized in the basin between 3.97 and 3.60 Ma: i) short-lived, low-energy pelagic lacustrine; ii) short-lived, moderately low-energy settings in delta plain and distributary channels; and iii) dominant, fluvial settings of the axial proto- Omo River.

The airfall Moiti Tuff captures pelagic lacustrine settings of Lonyumun Lake in many parts of the basin, including Gus (Gathogo et al., 2008), Loiyangalani (Gathogo et al., 2008), Allia Bay, and Kosia, which indicates that Lonyumun Lake covered the largest area of any Omo Group lake and possibly maintained that size through at least two precessional cycles. Local depositional settings also captured by the airfall ash include deltaic conditions elsewhere in the southern part of the Koobi Fora region, southern Kosia, and southern Loiyangalani. In the latter, the occurrence of symmetrical ripples formed by wave activity in shallow water and impressions of blade-like leaves typical of swamp grass indicate transitional environments associated with marshland or submerged delta plains. In the Ileret area, the Moiti Tuff is contained within quartz- and perthite-rich gravels that represent alluvium from the eastern basin margin, deposition of which must have begun in latest Lonyumun times.

Fluvial deposition of the Moiti Tuff is recognized in several parts of the basin, including West Turkana (southern Topernawi, Nasechebun, and Kataboi) (Harris et al., 1988a, b), northern Kosia, Ileret (Gathogo and Brown, 2006; Kidney, 2012), and Allia Bay (Brown and Feibel, 1986) where it was deposited in and by a meandering channel of the perennial axial system of proto-Omo River. A browser-dominated mammalian assemblage typically associated with floodplains and gallery forests (Feibel et al., 1991) comes from this lower part of the Moiti Member at one locality in Allia Bay. The remaining strata (up to 60 m) are fluvial deposits of the axial fluvial system consisting of clay-rich mudstones representing expansive floodplains (with paleosols). These mudstones are interrupted by sandy intervals that were deposited in meandering channels. These deposits are difficult to place stratigraphically without the confining marker beds (the Moiti and Lokochot Tuffs; Brown and Feibel, 1986) due to lack of lithologically distinct units or features such as biolithic beds, conglomerates, or erosional surfaces. These deposits are either missing in the southwestern part of the basin or compose the upper portion of the Muruongori Member (Leakey and Harris, 2003) at Lothagam, which is of reversed paleomagnetic polarity (Powers, 1980). A few mammalian fossils (Leakey and Harris, 2003) come from undifferentiated deposits at Lothagam in this interval.

Based on the records of astronomically forced climatic events (Fig. 8), the Moiti Member experienced about the same number of solar insolation cycles (precessional scale) as the underlying Lonyumun Member and a prominent eccentricity maximum (3.85–3.80 Ma; Fig. 4), but lacustrine deposits are not present during this maximum at the base. Intensified monsoon rains during the eccentricity maximum as indicated by a thick sapropel layer and high sea surfaces temperature in the Mediterranean Sea coincide with meandering fluvial settings of the perennial axial system. It is possible, and perhaps likely, that precession-scale climatic cycles modulated the deposition of fluvial units. These observations suggest that changes in tectonic settings favoring formation of a lake, but which are lacking during Moiti Member times, are more necessary than climatic changes for the formation of an extensive lake within the basin. Absence of a lake during Moiti Member time does not support the hypothesis (Trauth et al., 2005, 2007, 2009;

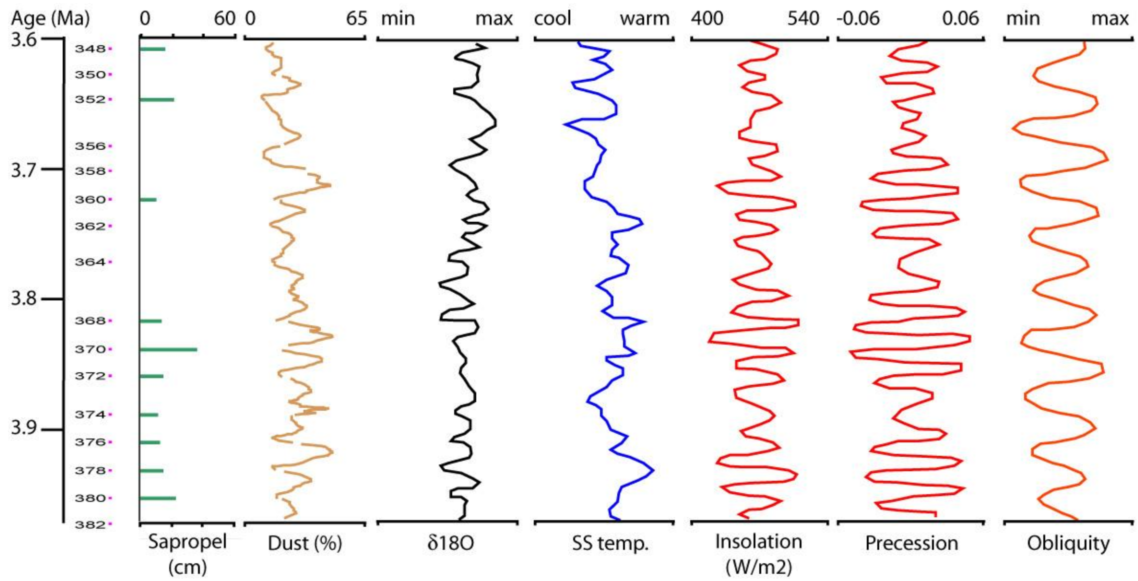

Figure 8. Climatic proxy records for the Moiti Member time interval (3.97–3.60 Ma) based on marine deposits from the Mediterranean Sea. Associated eccentricity cycles and predicted lake occurrence times are shown in Figure 4. See text for data sources and explanation.

Maslin et al., 2014) that eccentricity cycles determined the occurrence of major lake sequences in the basin with tectonics only playing significant role on very long time scales on the order 100,000 years.

Lokochot Member Time: 3.60–3.44 Ma

Deposits of the Lokochot Member constitute a composite section about 36 m thick (measured at Allia Bay; Brown and Feibel, 1986) and they account for a more abundant and widespread occurrence of mammalian fossils than underlying strata of the Omo Group. Perhaps the most publicized find from these deposits is *Kenyanthropus platyops* (Leakey et al., 2001), whose discovery changed the previous understanding about hominid evolution by adding a new genus alongside *A. afarensis* (Balter, 2001;

Lieberman, 2001). Deposits in this interval exhibit extreme variety in depositional settings ranging from alluvial fans through axial fluvial system to pelagic lake conditions. Outcrop geology in many locations provides abundant structural evidence for a tectonic episode that coincides with a very pronounced change in depositional settings—the development of large pelagic lake. By contrast, climatic influences on depositional settings appear to have been secondary and are generally masked by tectonic influences (Fig. 4).

Three main depositional settings are recognized in strata of the Lokochot Member: i) an axial fluvial system of the perennial proto-Omo River; ii) alluvial fans and associated ephemeral channels from the basin margins; iii) local, low- to moderately low-energy lacustrine settings; and (iii) lacustrine settings ranging from shallow to pelagic conditions. A prominent disconformity that locally passes into an angular unconformity is found within, below, or above the Lokochot Member. The member is less than 35 m thick in most outcrop areas (de Heinzelin and Haesaerts 1983a, b; Brown and Feibel, 1986; Harris et al., 1988a, b; Gathogo et al., 2008), and is completely eroded over short distances in some locations (Gathogo et al., 2008).

Deposition of the Lokochot Member sequence began when the Lokochot Tuff was laid down widely across the basin as an airfall ash where it captures floodplain settings typical of the axial fluvial system of the proto-Omo River at least at Kataboi (Harris et al., 1988a, b) in West Turkana, Ileret in East Turkana (Gathogo and Brown, 2006), and also at Echawe (F. H. Brown, pers. comm.) in South Turkana. The airfall Lokochot Tuff is recorded as far away as the Lake Albert basin in the western branch of the East African Rift System (Pickford et al., 1991), Kipcherere in the Baringo Basin (Namwamba, 1993), and the Gulf of Aden (Sarna-Wojcicki et al., 1985; Brown et al., 1992). Fluvial depositional settings continued past the time when the Lokochot Tuff was finally deposited by the axial fluvial system of the proto-Omo River in many locations in the study area, including Kibish, Usno and Shungura in the north, Kokiselei and Lomekwi in the west, and Allia Bay in the east. Strata of this fluvial interval attain variable thicknesses (up to 14 m, at Usno; de Heinzelin and Haesaerts, 1983) before a major break or major change in depositional settings. Minor hiatuses between upward fining sequences possibly represent insolation minima in precession-scale astronomical forcing on climate. Further, along the Lorienetom margin south of Liwan, the airfall Lokochot Tuff is preserved in a sequence of alluvial conglomerates derived from the basin margin, whereas at Loruth Kaado the same tuff is preserved in axial fluvial deposits.

A major event affected depositional settings across the study area shortly before 3.5 Ma that is recorded differently in disparate parts of the basin; here this event is interpreted as a manifestation of basinwide tectonic activity that lasted for only a brief period. The argument in support of this hypothesis is as follows. Significant structural movements associated this tectonic episode were greatest along preexisting longitudinal faults, particularly along border faults near the basin margins and in parts of the Loiyangalani region. High subsidence rates on the hanging block wall side of such faults resulted in the development of pelagic lakes where thick diatomite deposits occur, for example, at Nasechebun and Kataboi in West Turkana (Harris et al., 1988a, b) and Allia Bay in East Turkana (Brown and Feibel, 1986). Uplift along the footwall blocks of these major faults created high regions that became source areas for basaltic conglomerate and gravel incorporated in local alluvial fans, particularly where major ephemeral rivers from the basin margins already existed (e.g., at Lomekwi and Topernawi in West Turkana; Leakey et al., 2001), and metamorphic gravels (e.g., at South Turkwel) where basement metamorphic rocks formed the footwall in the drainage area of ephemeral streams from the basin margin (i.e., along the Kosia-Lokichar Fault; Fig. 6). Alluvial fan and ephemeral channel deposits at Lomekwi may have formed expansive spits in the Lokochot Lake. The Shungura and Usno Formations were minimally affected, particularly in relation to a change in depositional settings, possibly because the Lower Omo Valley is distal to the major border faults along the basin margins. Nevertheless, there is subtle evidence for tectonic activity in the northern regions as described next.

The fluvial sequence of lower Member A that conformably overlies Tuff A (= Lokochot Tuff) in the type area of the Shungura Formation is deformed by faulting that predates it (de Heinzelin and Haesaerts, 1983a). This faulting provides local evidence for the tectonic episode that is described above in other parts of the basin. Farther north, at Usno, correlative deposits record a unique change in depositional settings about 14 m above the base of Tuff H.C.N. (= Lokochot Tuff) where a sandy bed with mollusks is overlain by laminated clayey siltstones that show reducing conditions at the top (de Heinzelin and Haesaerts, 1983b). The mollusk-rich unit and the overlying fine-grained interval are presented here as local evidence of an intermittent fluvial lake coinciding with development of tectonically related lacustrine settings in many parts of the basin. This is also the case near Kibish, where the upper part of the Lokochot Member preserves diatomites directly below the Tulu Bor Tuff. As in the Shungura Formation, fluvial settings resumed at Usno and Kibish shortly thereafter.

The Lokochot Lake likely extended to the southern region of the basin, but absence of the Lokochot Tuff at many locations in southern parts of the basin makes it difficult to verify this assumption. For example, in the Kanapoi region, Feibel (2003) described a thin interval of ostracod-packed claystones and fissile green claystones interpreted as having been deposited in a lake or pond that may represents the Lokochot Lake in the upper part of the Kanapoi Formation. These claystones are overlain by the Kalokwanya Basalt, for which K/Ar dates of 3.11 ± 0.04 and 3.41 ± 0.04 Ma are available (Leakey et al., 1995). The older of these measurements is commonly accepted as a lower limit on the time of deposition at Kanapoi. This basalt is of reversed paleomagnetic polarity (Powers, 1980), so of the ages reported above, only the younger is acceptable because the other date (3.41 Ma) corresponds to the lower Gauss Normal Epoch (Table 2). It may be that neither age represents the time of eruption. Further, the β-Tulu Bor Tuff, dated between 3.41 and 3.43 Ma (McDougall et al., 2012), crops out at Namadang where it lies stratigraphically well above the Kanapoi basalt (F.H. Brown, personal communication).

Topographic changes associated with the tectonic movements described above resulted in pronounced changes in drainage patterns involving ephemeral rivers/channels and the ability of alluvial channels and rivers to erode local section. For example, deposition by a fluvial system dominated by a seasonal predecessor of the Turkwel River is invoked for the upper part of this sequence in the southern part of the basin (Ward et al., 1999; Leakey and Harris, 2003). Accordingly, many unconformity and disconformity surfaces in this time interval correspond to erosion in many parts of the basin prior to deposition during the remaining part of this time interval, but deposition occurred at fewer locations. For instance, in the Loiyangalani region, at least 45 m of the section between the airfall Moiti Tuff and fluvial Tulu Bor Tuff was eroded over a distance of a few tens of meters between the hanging-wall and foot-wall blocks of a large syndepositional fault (Gathogo et al., 2008) that is believed to have activated about 3.5 Ma. Angular unconformity is inferred at this location. It is because of this structural relationship and associated erosion that in many locations, including southern Ileret (Gathogo and Brown, 2006) and Il Naibar (Buchanan, 2008), outcrops of the Lokochot Tuff are localized, and coincide with faults, whereas overlying beds, including the Tulu Bor Tuff, form relatively extensive outcrops. Elsewhere, as in northern Area 40, there is a pronounced disconformity between the Moiti Tuff and the Tulu Bor Tuff, with no trace of the Lokochot Tuff preserved. Also, in the north central part of the Koobi Fora region, the Lokochot Tuff is absent in a well-exposed section below the Tulu Bor Tuff where it is underlain by the Lonyumun Member with neither the Lokochot nor the Moiti Tuff preserved.

Paleoclimatic data from the Mediterranean Sea (Fig. 9) show that during Lokochot Member time, variations from dry to wet conditions were more subdued than was the case earlier in the Pliocene. For example, terrigenous grains account for about 30% of core sediments in the Gulf of Aden (deMenocal, 1995) and sapropel layers are at most 14 cm thick (Lourens et al., 1996) during this interval compared to 47% and 34 cm in thickness, respectively, in older Pliocene intervals. From these observations, it does not appear that during this time there was a pronounced interval of high precipitation that might account for the formation of pelagic lakes even during the eccentricity maximum between 3.45 and 3.47 Ma (Fig. 4) shortly before deposition in the Lokochot Member time interval ended with deposition of the Tulu Bor Tuff. The Lokochot Lake facies

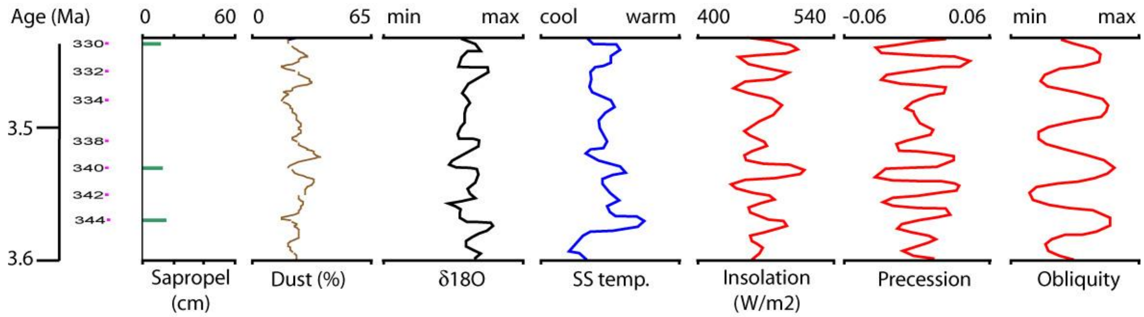

Figure 9. Climatic proxy records for the Lokochot Member time interval (3.60–3.44 Ma) based on marine deposits from the Mediterranean Sea. Associated eccentricity cycles and predicted lake occurrence times are shown in Figure 4. See text for data sources and explanation.

coincided with a period of very subdued climatic conditions, according to deep-sea records, with sapropels missing even during a subordinate (100 ka) eccentricity maximum at 3.49–3.50 Ma. The lack of correspondence between the climatic record and the Lokochot Lake suggests strongly that tectonic events were more important than climate in its development in the Omo Turkana Basin between ~3.52 and 3.44 Ma.

Based on the relations presented above, tectonics seem to have played significant role the formation of Lokochot Lake in the basin role on a shorter time scale than the 100,000 years suggested for tectonic effects by Trauth et al. (2005, 2007, 2009) and Maslin et al. (2014). These workers recognize the Lokochot Lake as a deep lake, but they placed it at a slightly older age (~3.6 Ma; their Fig. 3), which is incorrect because the underlying Lokochot Tuff (3.60 Ma) is separated from the lacustrine strata by ~10 m, whereas the overlying α-Tulu Bor Tuff (3.44 Ma) is in direct contact with those strata.

Tulu Bor Member Time: 3.44–2.62 Ma

Sedimentation during the time represented by the Tulu Bor Member (up to ~191 m thick), in the Lower Omo Valley (de Heinzelin and Haesaerts, 1983a) began with widespread deposition of fluvial deposits by the axial perennial system of the Omo River as indicated by the occurrence of the Tulu Bor Tuff filling large, deep channels. Outcrops of this tuff are described from Usno (de Heinzelin and Haesaerts, 1983b) to Shungura (de Heinzelin and Haesaerts, 1983a) in the northern part of the basin, Kokiselei to Kataboi (Harris et al., 1988a, b) in the western part, Ileret to Allia Bay (Brown and Feibel, 1986; Gathogo and Brown, 2006) in the eastern part, and Echawe and west of Kanapoi (F. H. Brown, pers. com.) to Loiyangalani in the southern in part (Gathogo et al., 2008). These deposits have produced more mammalian fossils than the underlying Pliocene sequence. Hominins include *A. boisei* (Walker et al., 1986; Feibel et al., 1989) in the uppermost part of this interval and other hominin specimens of as yet undetermined species, particularly from Lomekwi (Leakey et al., 2001). New evidence based on field observations and presented here indicates that deposition during this time interval was affected by large tectonic events about 3.2 Ma ago in much the same way as those just described above for Lokochot Member time. Moreover, a pelagic lacustrine sequence that locally coincides with this episode was intruded by basalt, just as was the case in development of the Lonyumun Lake. Climatic signals are generally evident as fluvial cycles and the timing of the pelagic lacustrine sequence did not coincide with climatic conditions favorable for the formation of major lake (Fig. 4).

Three categories of depositional settings are identified in the Tulu Bor Member time sequence: i) a major fluvial system with extensive floodplains and meandering channels of the axial perennial proto-Omo River dominating at least the lower interval; ii) high-energy settings of alluvial fan to ephemeral channels from the basin margins characterizing many parts of the basin beginning by 3.2 Ma; and iii) shallow lake to local pelagic settings coinciding with local basalt intrusions. These deposits can be lithologically subdivided into lower and upper sequences based on an extensive depositional hiatus or sharp change to lacustrine facies at the top of the basal fluvial sequence. These prominent changes represent the effects of a major structural event at about 3.2 Ma as mentioned above.

The Tulu Bor Tuff is commonly represented by fluvial deposition of two eruptions (-α and -β) (de Heinzelin and Haesaerts, 1983a, b) separated by a well-developed paleosol that contains mammalian fossils at some locations (e.g., Lomekwi) and a local basaltic tuff (the Lokhoya Tuff) at Loiyangalani (Gathogo et al., 2008). In some areas, particularly those outside the present boundaries of the basin, as at Kipcherere in the Baringo Basin, the Tulu Bor Tuff must have been deposited as an airfall ash, and the exposures at Echawe and west of Kanapoi may represent locally reworked airfall material. Thus some fluvial accumulations within the Omo-Turkana Basin may result from local reworking of an airfall ash layer into local topographic lows. Nonetheless, the distribution of channel-form deposits of the Tulu Bor Tuff (up to 17 m thick in the Allia Bay region) indicates stable tectonic settings that enabled the axial fluvial system to deposit the ash over a large part of the basin. Perennial flow in this river is shown by the occurrence of *E. elliptica* in many locations including Lomekwi and Ileret (Gathogo and Brown, 2006). The river is recorded across the basin covering about 280 km from Kibish (Brown and Fuller, 2008) in the north to Loiyangalani (Gathogo et al., 2008), where its tuffaceous deposits even extend slightly outside the catchment area of the present drainage basin. Alluvial settings associated with the basin margins were subdued during deposition of the earlier part of this interval. For example, floodplain deposits of the axial fluvial system including the Tulu Bor Tuff were laid down on conglomeratic deposits close to the basin margin (alluvial fan deposits of the proto-Topernawi) as far west as the underlying channel deposits of the Lokochot Tuff. Depositional settings of the axial fluvial system associated with the perennial proto-Omo River dominated most of the basin until shortly before eruption and intrusion of the Kankam Basalt (about 3.2 Ma, Gathogo et al., 2008), which coincides with structural movements during this interval.

Alluvial fan and ephemeral channel settings became common in many parts of the basin during and after a tectonic episode associated with the Kankam Basalt intrusions. Paleocurrent directions, geometry, location, and mineralogical composition indicate that these alluvial fans and associated alluvial channels came from basaltic ranges along the basin margins. They represent episodes of pronounced gradient increase along ephemeral rivers such as Topernawi (Harris et al., 1988a, b; Leakey et al., 2001) in the Lomekwi area west of Lake Turkana and Il Eriet (Gathogo, 2003; Gathogo et al., 2008) east of the lake. These alluvial deposits grade from muddy conglomerates with basaltic and metamorphic clasts to gravelly sandstones that contain clasts showing that they were derived from the basin margins. Supporting evidence for a tectonic episode at this time comes from the Loiyangalani region where there is an angular discordance of up to 4° dipping towards the basin center (west) between the lower and upper Tulu Bor Member deposits (Gathogo et al., 2008). This unconformity coincides with intrusion of the

Kankam Basalt, and includes at least two flows that confine thin deposits representing a transition from deltaic to pelagic lacustrine settings of the Waru Lake. Flows of the Kankam Basalt are believed to have occurred within a short time interval similar to that required by flows of the Gombe Group Basalt or the two eruptions of the Tulu Bor Tuff (-α and -β).

The tectonic episode during Tulu Bor time is believed to have reactivated and possibly extended major longitudinal faults (e.g., along western Lomekwi in West Turkana), thus causing significant structural movements and finally culminating with local basalt intrusions (e.g., in the region north of Loiyangalani; Gathogo et al., 2008). Such structural changes would have caused major shifts in drainage patterns with consequent significant changes in depositional settings. Thus increased accommodation space along rapidly subsiding hanging-wall blocks resulted in flooded subbasins that finally coalesced to form larger lakes. In the southern part of the basin at Loiyangalani the lacustrine settings are represented by diatomaceous claystones (pelagic environments) that laterally correlate with biolithic units containing gastropods and algal stromatolites (shallow lake margins). The development of this lacustrine sequence near Loiyangalani was intimately associated with intrusion of the Kankam Basalt (~3.2 Ma), with one flow forming the lake bottom and dikes and subsequent flows intruding pelagic deposits during the lake’s existence. Not only the basalt but also the pelagic lake deposits are thicker and more extensive in the eastern part of the Loiyangalani region (in the Kankam area) where fault-related subsidence rates are believed to have been higher than in the adjacent area to the west (Lenderit). Moreover, the general extension directions (east-west) of the basin and dip directions (west) of the main fault blocks are indicated by the orientation of dykes and pipe vesicles of the Kankam Basalt, respectively (Gathogo et al., 2008).

Evidence of an extensive lake at about 3.2 Ma occurs farther north in Allia Bay and includes a fine-grained interval composed of laminated siltstones and claystones associated with the Waru Tuff, and laterally with biolithic units that are rich in mollusks (Brown and Feibel, 1986). The latter represent shallow lake conditions and are locally observed at other locations, including Loiyangalani (Gathogo et al., 2008) and Lomekwi (Leakey et al., 2001), where they include the molluskan trace fossil *Lockeia*. Based on age, this pelagic lacustrine sequence is synchronous with the Waru Lake of Feibel (2011) and the Allia Lake of Trauth (2005, 2007, 2009) as well as Maslin (2014). Here the name "Waru Lake" is preferred because the Waru Tuff is clearly associated with this lacustrine sequence.

Both the Allia Tuff and the Waru Tuff lie between the Toroto Tuff (3.308 ± 0.022; Table 1) and the Ninikaa Tuff (3.066 ± 0.017; Table 1) in East Turkana at Allia Bay (Cerling and Brown, 1982; Brown and Feibel, 1986; McDougall and Brown, 2008). Deposits of the Waru Lake are described farther south of Lomekwi at Topernawi where an airfall ash, the Waru Tuff, was deposited in an extensive but short-lived lake (Harris et al., 1988a, b) whose deposits contain abundant sponge spicules and diatoms. Knowing of the record of this lake near Loiyangalani results in its minimum area being on the order of 8000 km^2. It is thus quantitatively different from the Kokiselei Lake (~2.5 Ma; see below) of Feibel (2011), although he considered both to be minor lacustrine sequences.

Lacustrine deposits are typically absent in parts of the basin where the upper part of the sequence during this time interval is dominated by coarse-grained deposits of alluvial fan and ephemeral channels, for example in western Loiyangalani, northern Koobi Fora, and Ileret. This absence can be explained by subsequent erosion by alluvial channels from the margins or/and critical reduction in depositional rates as might be expected on footwall blocks that were relatively uplifted during the tectonic episode.

As expected, because of the distal location of major faults along basin margins, there is no significant lithologic variation between the lower and upper Tulu Bor Member time sequences in the northern part of the basin at Usno and Shungura. The entire sequence is dominated by fluvial deposits of the Omo River (de Heinzelin and Haesaerts, 1983a, b). However, the structural disturbance described above in other parts of the basin is believed (this dissertation) to coincide with an unconformable surface with up to 4 m of erosional relief in the upper part of submember B-2 of the Shungura Formation (de Heinzelin and Haesaerts, 1983a) and possibly with the base of a gravelly unit with rolled bones in submember U-12 of the Usno Formation (de Heinzelin and Haesaerts, 1983b). These prominent depositional surfaces lie near the base of the Mammoth Subchron (de Heinzelin and Haesaerts, 1983b; Brown et al., 1978; Kidane et al., 2014), which has began 3.21 Ma ago (see Table 2).

The large number of mammalian fossils of the Tulu Bor Member time interval in the Usno Formation come mostly from below or within unit U-12 (‘Brown Sands’ and ‘White Sands’), particularly in the case of the latter (de Heinzelin and Haesaerts, 1983b). Mammalian fossils in the lower part of Member B of the Shungura Formation come mainly from submember B-2 (de Heinzelin and Haesaerts, 1983a). This temporal abundance in fossils is consistent with collections from other parts of the basin, including West Turkana at Lomekwi (Harris et al., 1988a, b) and East Turkana at Allia Bay (Feibel et al., 1991). It is predictable that the Tulu Bor Member time sequence should produce the highest number mammalian fossils in the Omo Group, given its temporal length (about 800 ka) and widespread exposures of strata deposited in settings favorable to mammalian habitats and subsequent preservation of the resulting death assemblages. A majority of published mammalian fossils come from among the thickest parts of the section of the Shungura Formation (about 191 m; de Heinzelin and Haesaerts, 1983a) in the Lower Omo Valley, and from the thickest part of the Nachukui Formation in the Lomekwi area (158.5 m; Harris et al. (1988a, b) in West Turkana. Aside from the collections at L2 in lower Member B of the Shungura Formation (de Heinzelin and Haesaerts, 1983a), the largest collection from that member comes from its upper part at L1, which is ~2.96–2.97 Ma in age, judging from the date on Tuff B-10 (2.965 ± 0.014 Ma).

The tectonic movements at ~3.2 Ma appear to have had far-reaching effects on deposition settings in many locations, except in the Lower Omo Valley. For example, distal alluvial fan and associated ephemeral channel settings from the basin margins become dominant at Lomekwi (Harris et al., 1988a, b), Ileret (Gathogo and Brown, 2006), most of Il Naibar (Brown and Feibel, 1986), and western Loiyangalani (Gathogo et al., 2008). Sedimentation rates were critically reduced and/or erosion enhanced in many locations, for example along the Karari ridge (Brown and Feibel, 1986) and in the eastern part of the Loiyangalani region (Gathogo et al., 2008). However, fluvial settings dominated by the axial perennial system of the proto-Omo River, similar to conditions described in the Lower Omo Valley, particularly Shungura (de Heinzelin and Haesaerts, 1983a), returned to many parts of the basin. These continued through the upper part of

Member B (B-3 to B-12) and Member C, and they pertained until the end of the Tulu Bor Member. This is when the Burgi Tuff was distributed in the basin by the proto-Omo River following its eruption at 2.622 ± 0.027 Ma (Table 1). The upper part of this interval represented by Member C (younger than 2.965 ± 0.014 Ma, age of B-10 Tuff; Table 1) of the Shungura Formation is either missing or poorly exposed west of Lake Turkana except at Lomekwi (Harris et al., 1988a, b) and it is likewise poorly represented in the northern and eastern parts of the Allia Bay area at Koobi Fora. Deposits indicative of a minor lacustrine episode occur at the top of this interval just below the Burgi Tuff at Allia Bay (Brown and Feibel, 1986) and possibly also farther north in the Shungura Formation by a thin interval between submembers C-7 and C-8 with shells of mostly *Caelatura* (bivalve) and *Cleopatra* (gastropod; de Heinzelin and Haesaerts, 1983a).

The timing, magnitude, and duration of astronomically modulated climatic cycles as recorded in the Mediterranean Sea (Fig. 10) seem to have generally resulted in precision-scale modulation of fluvial patterns in the Omo-Turkana Basin, but without sufficient strength for the development of lacustrine settings. The Waru Lake existed near the end of subdued climatic conditions during an eccentricity minimum when monsoonal precipitation is predicted to have been low, and which is documented by the scarcity of sapropel layers in the Mediterranean Sea. This is also a time when sea surface temperature was consistently low (Fig. 10). The Waru Lake coincides with a tectonic episode at about 3.2 Ma, but not with the eccentricity maximum at 3.07–3.04 Ma (Fig. 4). This maximum coincides with the Ninikaa Tuff (3.066 ± 0.017 Ma; McDougall et al. (1980); also Table 1), a unit that was deposited in East Turkana at its type locality by a channel of the axial fluvial system dominated by the perennial Omo River.

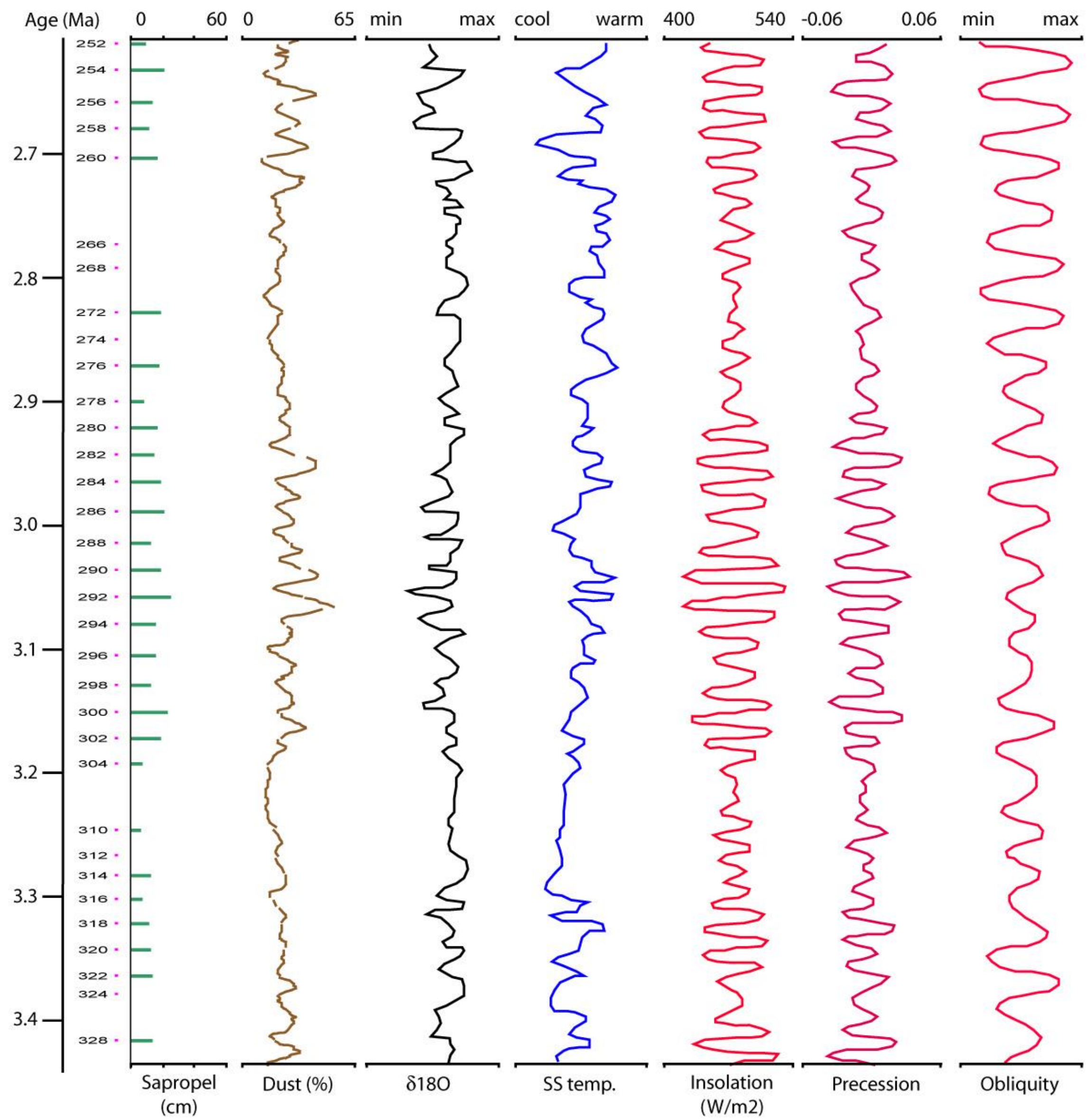

Figure 10. Climatic proxy records for the Tulu Bor Member time interval (3.44–2.62 Ma) based on marine deposits from the Mediterranean Sea. Associated eccentricity cycles and predicted lake occurrence times are shown in Figure 4. See text for data sources and explanation.

The beginning and end of Tulu Bor Member time captures parts of the extreme climatic conditions associated with an eccentricity maximum, but again, fluvial settings were dominant. However, the minor lake sequence described in the upper portion of Tulu Bor Member time possibly coincides with a subordinate eccentricity maximum, or perhaps coincides with intensified monsoonal precipitation following major changes in oceanic circulation. These changes in oceanic circulation are believed to result from the onset of Northern Hemisphere glaciations about 2.7 Ma (Raymo, 1994; Zhang et al., 2009). The latter notion supports observations by deMenocal (1995), that obliquity and eccentricity scale changes in climate did not significantly affect low latitude regions including eastern Africa until after the middle Pliocene.

In summary, Trauth et al. (2005, 2007, 2009) and Maslin et al. (2014) estimate that the Waru (their Allia) Lake began at ~3.1 Ma (their Fig. 3) some 100 ka later than is actually the case, and their placement is near the beginning of the main eccentricity maximum beginning at the same time (their Fig. 4). When placed properly in time, the main pelagic lacustrine sequence in the basin coincides better with tectonic events that change depositional conditions on a shorter time scale (Fig. 4) than is proposed by Maslin et al. (2014).

Burgi Member Time: 2.62–1.90 Ma

Deposits of the Burgi Member time interval represent the thickest portion of the Omo Group (up to 354 m) in the Shungura area (de Heinzelin and Haesaerts, 1983a). The upper half of these deposits is poor in mammalian fossils, because lacustrine sediments dominate the section (de Heinzelin and Haesaerts, 1983a; Brown and Feibel, 1986; Harris et al., 1988a, b). Some important published finds from the lower half of this interval include two hominin crania, the 'Black Skull' (*Australopithecus boisei*) and '1470' (*Homo habilis*) from the base of the sequence in West Turkana, and the top of the sequence in East Turkana, respectively (Feibel et al., 1989). A major tectonic episode is recorded in the geology of the basin at this time by erosion of a significant portion of the section in the eastern part of the basin, and subsequent development of a widespread lacustrine interval (Bruhn et al., 2010, and references therein).

Principal depositional settings recognized in the Burgi Member time interval are: i) fluvial settings in which the perennial axial system of the proto-Omo River laid down channel-form sandstones and clay-rich mudstones that represent expansive floodplains where soils developed, and ii) lacustrine settings that were dominated by pelagic conditions in many locations. These depositional settings were dominant in the basin in the order listed above. The lacustrine sequence has been linked to tectonic activity that produced structural discordance between strata in the eastern (Brown and Feibel, 1986; Patterson et al., 1970; Findlater, 1976) and the southern (Powers, 1980) parts of the basin. This structural disturbance coincides with a widespread erosional surface recognized throughout the eastern part of the basin that is informally known as the 'Burgi unconformity'.

Deposition during Burgi Member time began after eruption of the Burgi Tuff 2.62 Ma ago (Table 1), shortly before the end of the Gauss Normal Polarity Epoch (2.581–3.596 Ma; Table 2). The Burgi Tuff is recognized only in the Allia Bay region of the eastern part of the basin (Brown and Feibel, 1986; McDougall and Brown, 2006) where it lies within a 20 m thick section below the Lokalalei Tuff (2.526 ± 0.025; Table 1), and in a single outcrop north of the central part of the Koobi Fora Ridge. The western (West Turkana) and northern (Shungura) parts of the basin record much of this lower interval in sections about 10 m thick from the base of the Emekwi Tuff (Harris et al., 1988a, b) and Tuff C-9 of the Shungura Formation, respectively. These deposits record channels, levees, and floodplains of the perennial axial system of the Omo River that dominated much of the basin until shortly before deposition of the Kangaki Tuff of the Nachukui Formation (2.063 ± 0.032 Ma; Table 1). These fluvial deposits are preserved primarily in the northern (de Heinzelin and Haesaerts, 1983a) and western (Harris et al., 1988a, b) parts of the basin. Poorly drained floodplains that locally harbored small floodplain lakes (with gastropods) were common in many parts of the basin up to about 2.4 Ma near which time the Kokiselei Tuff was erupted, and possibly slightly later at Lokalalei. The locally pumiceous Burgi Tuff contains rhizoliths and is overlain by fining-upward units that begin with tuffaceous sandstones and end with pale olive brown to green claystones. A comparable lithology is observed below the Lokalalei Tuff in West Turkana and in the Shungura regions. de Heinzelin and Haesaerts (1983a) note that a reduced incipient soil in submember C-9-3 is widespread below the base of Tuff D in the Shungura Formation where it indicates very shallow water depth in a flat paleotopography.

It is based on the observations presented above that the Kokiselei Lake (~2.5 Ma) of Feibel (2011) is classified here as a local fluvial lake with marginal lacustrine features including mollusk-rich layers. Thus ranking the Kokiselei Lake as representing a situation similar to that of the Waru Lake (Fig. 3 of Feibel (2011)) would be misleading because the latter includes diatomaceous facies and is represented in many parts of the basin (see above, within the Tulu Bor Member time interval). Using terms such as

“Kokiselei Lake” to describe a local depositional phenomenon such as a floodplain lake is to be discouraged, because it detracts from the value of the pelagic lake deposits in elucidating the geological history of the basin. Similar floodplain lake deposits along with associated swamp deposits are described in detail by Butzer (1971), de Heinzelin (1983), and Gathogo (2003).

Wet conditions in the Omo-Turkana Basin that are described above in the basal portion of the Burgi Member time interval coincide with increased monsoonal precipitation in the eastern African region, as is indicated by thick sapropel layers in the Mediterranean Sea (Fig. 11). This period of increased precipitation continued from the eccentricity maximum at 2.58–2.54 Ma until shortly before the eccentricity minimum at 2.42–2.44 Ma. Despite evidence given above for reduced soils that may have resulted from increased precipitation, no pelagic lacustrine deposits have been recognized in the Omo-Turkana basin at this time.

Compressed sections characterized by undifferentiated mudstones and basaltic to metamorphic pebble conglomerates were deposited during lower Burgi Member time in the eastern part of the basin, particularly north of Allia Bay. Development of this lithology is interpreted as being the effect of syndepositional structural movements that were associated with a tectonic episode starting at about 2.2 Ma in which extensive faulting and local basalt intrusions occurred. The clearest geologic evidence for the timing of the tectonic episode is preserved in the Loiyangalani region where an angular unconformity is observed between the Kokiselei Tuff and the first Lenderit Basalt flow (2.18 ± 0.02 Ma; Gathogo et al., 2008). There the section between the tuff and the basalt ranges in thickness from 45 m to less than 5 m within a few kilometers depending on

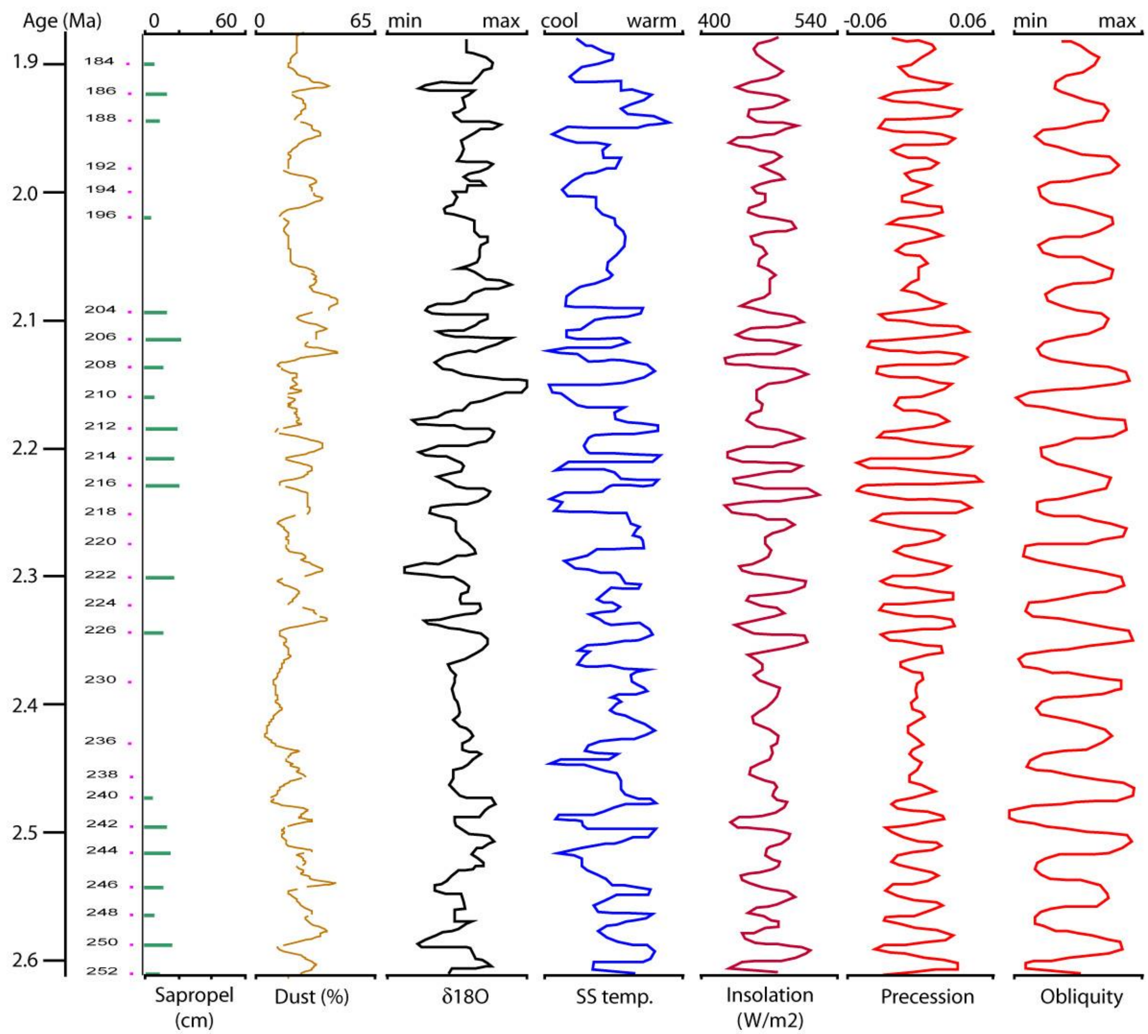

Figure 11. Climatic proxy records for the Burgi Member time interval (2.62–1.87 Ma) based on marine deposits from the Mediterranean Sea. Associated eccentricity cycles and predicted lake occurrence times are shown in Figure 4. See text for data sources and explanation.

the location of local longitudinal faults that developed or were reactivated during this interval. The basalt normally occurs as narrow flows (up to three) and dykes along large longitudinal faults.

The largest lake recorded in the basin has been termed the Lorenyang Lake, with a minimum area of ~14,000 km^2. Its deposits are well expressed along the northern side of the Koobi Fora Ridge at Koobi Fora, in the Nachukui Formation at Nanyangakipi, Kaitio, Kalochoro, Kokiselei, and Kangaki, and in both the type area and the Kalam area of the Shungura Formation in southern Ethiopia, and probably in the Lenderit drainage at Loiyangalani. These lacustrine strata comprise the upper Burgi Member of the Koobi Fora Formation (Brown and Feibel, 1986), part of the upper Kalochoro Member, and most of the Kaitio Member in the Nachukui Formation (Harris et al., 1988a, b), as well as upper Member G (units G-14 to G-26) in the Shungura Formation (de Heinzelin and Haesaerts, 1983a).

Interpretation of these strata is complicated by the following: i) difficulty determining the time of inception of lacustrine deposition in the Koobi Fora region because the strata lie above an unconformity within the Burgi Member, ii) deposition of lacustrine strata in the Koobi Fora region appears to have begun earlier in western sections than it did in eastern sections, and iii) the Morutot Tuff (1.60 Ma) may lie disconformably on the uppermost lacustrine strata deposited in Lorenyang Lake at Nanyangakipi, and neither the age of the top nor of the bottom of these strata is well-controlled in time.

With those caveats in mind, the age of the base of lacustrine strata deposited in the Lorenyang Lake is best controlled by the age on the Kang'aki Tuff (2.06 ± 0.03 Ma), which lies less than 12 m above the first lacustrine strata at Kang'aki. This is consistent with identification of the younger Reunion Event in upper submember G-13, which is estimated to have an age of 2.06-2.08 Ma (see Table 2). This paleomagnetic event is also believed to be recorded near the base of the section along the eastern part of the Koobi Fora Ridge at Koobi Fora (Joordens et al., 2013), but deposition of Lorenyang Lake strata evidently began somewhat later along the western part of Koobi For a Ridge, because the younger Reunion Event is not recorded in the section there (Joordens et al., 2011).

The age of the youngest deposits in the Lorenyang Lake is evidently diachronous because in the Shungura Formation, deposition ceases by the beginning of the Olduvai Event (1.945 Ma), but it continues until above the KBS Tuff (1.87 ± 0.02 Ma), along the western part of Koobi Fora Ridge (Joordens et al., 2011), and perhaps to the top of the Olduvai Event at 1.778 Ma (Lepre and Kent, 2010). Here the boundary of the pelagic lake deposits at Koobi Fora is somewhat arbitrarily placed at an algal marker layer (A2; Feibel, 1984) exposed along the central part of Koobi Fora Ridge, which has an estimated age of 1.78 Ma (Lepre and Kent, 2015). Along the eastern part of Koobi Fora Ridge, lacustrine deposition ceased before deposition of the KBS Tuff (1.87 Ma).

At Lenderit, two basalt flows dated at 2.18 ± 0.02 and 2.02 ± 0.02 Ma confine a pelagic lacustrine sequence (between 2.18 ± 0.02 and 2.02 Ma) that correlates with the older part of the Lorenyang Lake deposits in the northern part of the Omo-Turkana Basin. The relationship between structural movements and changes in depositional settings that include a widespread lake and local alluvial fans is poorly evidenced over much of the Omo-Turkana Basin. However, the geological relationships in the Loiyangalani region provide a reliable template for a possible relationship between tectonic activity, depositional settings, and stratal thickness in much of the basin. For example, maximum vertical displacement is expected to have occurred along preexisting longitudinal faults toward the basin axis (now offshore in Lake Turkana) and along border faults that define the western margins of the basin. As expected, the thickest portion of the Burgi Member time sequence represents less than one-quarter of the total depositional time, and the sections thicken towards the axis of the basin.

The pelagic lacustrine sequence described above corresponds to the Lorenyang Lake of Feibel (2011) and probably to a large deep lake that is discussed by Trauth (2005, 2007, 2009) and Maslin (2014), which they show as beginning about 1.9 Ma (see Fig. 3 of Maslin et al., 2014). Evidence presented above indicates that deposition in Lorenyang Lake had started by at least 2.05 Ma. Further, there is no field evidence for shallow alkaline lakes posited by Trauth (2005, 2007, 2009) and Maslin (2014; their Fig. 3) being prominent in the basin between 2.3 and 1.9 Ma.

Examination of the astronomically forced regional paleoclimatic record of the Mediterranean Sea (Fig. 11) reveals that the main lacustrine sequence of Burgi Member time occurred during an eccentricity minimum between 2.09 and 1.94 Ma ago (Fig. 4) when conditions were extremely dry. Monsoonal rainfall was minimal during this period as indicated by the scarcity of sapropels in the Mediterranean Sea, and this coincides with subdued amplitudes of solar insolation cycles. These observations confirm the tectonic origin of the main Late Pliocene lake sequence in the Omo-Turkana Basin (Fig. 4) that was originally suggested by early studies (Brown and de Heinzelin, 1983). Contrary to conclusions of recent studies (e.g., Maslin et al., 2014), global climate as expressed by eccentricity cycles was decidedly not the trigger for the origin of the Lorenyang Lake ~2.06 Ma ago.

Deltaic to fluvial depositional settings resumed at some places in the northern part of the basin and along its northeastern margins (e.g., Area 41; Gathogo, 2003) shortly before the beginning of the Olduvai Subchron at 1.945 Ma (e.g., Kidane et al., 2007). These deltaic and fluvial deposits are interpreted to represent diminishing accommodation space as sedimentation and tectonic subsidence rates approached a balance. Strata that record onshore depositional settings are described in the Shungura region (de Heinzelin, 1983), in eastern Ileret (Gathogo, 2003; Gathogo and Brown, 2006), at Karari (Brown and Feibel, 1986), and in the northern part of the Loiyangalani region (Gathogo et al., 2008). Mammalian fossils have been collected from these deposits except at Loiyangalani.

KBS Member Time: 1.90–1.60 Ma

Deposits of KBS Member time in the Omo-Turkana Basin are up to 169 m thick in West Turkana (Harris et al., 1988a). They are known for their exceptional richness in mammalian fossils, especially in the Koobi Fora and Ileret regions in the eastern part of the basin where the most extensive exposures occur. Geographic and temporal variability in depositional settings in various parts of the study area is more pronounced than is the case in most other member-level intervals of the Omo Group. This characteristic variability results from the combined effects of periodic astronomical climate forcing (Brown, 1995; Lepre et al., 2007) and tectonic activity (Gathogo and Brown, 2006). The latter includes a tectonic episode between 1.79 and 1.75 Ma that began with local extrusion of the Balo Basalt (1.79 ± 0.02 Ma; Gathogo et al., 2008) and culminated with vertical structural movements that affected many parts of the basin as described below.

Two main depositional settings are recognized during KBS Member time: i) lacustrine settings ranging from pelagic to shallow conditions, and ii) fluvial settings dominated by the perennial axial system of the Omo River, but locally interrupted by shallow lakes and ephemeral channels from the basin margins. Depositional settings in this interval varied significantly depending on time and location. For example, the depositional settings listed above existed concurrently in different parts of the basin during the early stages, when pelagic to shallow lake conditions characterized the western part of the basin while perennial fluvial channels of the proto-Omo River dominated the eastern regions. The main tectonic episode in this time interval coincides with a significant reduction in depositional area and a shift in depositional patterns from widespread lacustrine sediments to prominent ephemeral channels entering from the basin margins.

Deposition of the KBS Member began with an airfall ash of the KBS Tuff that is now preserved as thin layer (normally <15 cm thick) of glass (local altered to bentonite) in many locations in the eastern, western, and southern parts of the basin (Harris et al., 1988a; Leakey and Harris, 2003; Brown et al., 2006). West of Lake Turkana, the airfall ash layer is altered to pink bentonite, but the remnant glass is compositionally indistinguishable from typical samples of the KBS Tuff from the rest of the basin. The airfall ash captures pelagic lacustrine settings in the upper part of the Lorenyang Lake deposits at many locations in East Turkana, for example the Bura Hasuma area in the west central Koobi Fora region (Brown et al., 2006; Gathogo and Brown, 2006), the area between Kaitio and Kang'aki in west Turkana (Harris et al., 1988a; Brown et al., 2006), and near Lothagam in southern Turkana (Leakey and Harris, 2003). From east to west, deposits that host the airfall ash in West Turkana (e.g., Kokiselei) change from pelagic to shallow conditions, and finally to floodplain deposits with well-developed paleosols that contain mammalian fossils. Farther south, at Lothagam, the KBS Tuff is an airfall deposit within shallow lacustrine strata (Feibel, 2003). Along the southern part of Koobi Fora Ridge, the airfall ash was laid down in pelagic lake conditions followed shortly afterward by another airfall ash (the Brown Tuff) before the fluvially transported KBS Tuff was deposited in a shallow lake at this location (Brown et al., 2006; Gathogo and Brown, 2006). Near Loiyangalani, the KBS Tuff occurs in a few locations (Gathogo et al., 2008) as lenses that lack pumice clasts, and there it is associated with abundant rhizoliths along with pedogenic carbonates—all features that suggest that it is a reworked airfall ash that settled on local alluvial plains, but near Eruth Lobokolei, it lies within shallow lacustrine strata.

Channel-form deposits of the KBS Tuff with pumices that represent fluvial settings of the Omo River are best represented in the eastern part of the basin, but are absent in West Turkana. A fluvial sequence of the lower KBS Member time interval occurs in the northernmost and easternmost parts of East Turkana, particularly east of Il Eriet where pumiceous channel-form KBS Tuff deposits indicate paleoflow to the south-southeast. There the sequence is represented by at least 23 m of upward-fining units of floodplain mudstones and channel sandstones (Gathogo, 2003; Gathogo and Brown, 2006). Concurrent fluvial deposits farther west in the Ileret area are interrupted by shallow lake deposits with biolithic beds (typically ≤25 cm thick) containing abundant gastropods and some small bivalves. Pumiceous deposits of the river-transported KBS Tuff are conformably overlain by a biolithic bed in the central Ileret area. Similar depositional settings where fluvial-dominated conditions were interrupted by short-lived shallow lakes were common in the Lower Omo Valley (de Heinzelin and Haesaerts, 1983a) and in the East Turkana region (Brown and Feibel, 1986) within ~20 km of the modern shoreline of Lake Turkana. The latter includes the Bura Hasuma area (Gathogo and Brown, 2006) but excludes the Karari escarpment and eastern Ileret (Gathogo and Brown, 2006). In contrast, lacustrine settings ranging from pelagic to marginal conditions dominated west of Lake Turkana where algal stromatolites and thick (>2m) gastropod-rich deposits (Harris et al., 1988a) represent shallow lake and persistent beach conditions, respectively, adjacent to a continually subsiding part of the basin. However, in the easternmost exposures, at Kaitio, the KBS Tuff is an airfall tuff within a laminated siltstone sequence.

The lower interval of the KBS Member time deposits thus far described represents depositional settings that were primarily determined by the basin geometry, with subsidence rates generally increasing towards the basin center and along the western side of the basin where the thickest onshore section occurs. The pelagic lacustrine strata described above represent continued deposition in the Lorenyang Lake of Feibel (2011) and probably the large deep lake of Trauth (2005, 2007, 2009) and Maslin (2014; Fig. 3) described as starting about 1.9 Ma in the basin. Shallow lake conditions in fluvially dominated intervals throughout East Turkana and in the Lower Omo Valley to the north may well represent climatic cycles, particularly precession-scale astronomical forcing (Brown, 1995; Lepre et al., 2007) when the Lorenyang Lake expanded its margins to the shallow and relatively flat regions of the basin. The following description focuses on changes in depositional settings in the basin during and after tectonic episodes that took place between 1.79 and 1.75 Ma.

The main tectonic event in the KBS Member time interval is manifested in a manner significantly different from that for most other tectonic events described previously in underlying member-level intervals. For example, this event does not coincide with a pelagic lake sequence shortly before or after basalt intrusion. In the Loiyangalani region where basalt volcanism occurred, alluvial fans and ephemeral channel settings became increasingly dominant prior to intrusion and extrusion of the Balo Basalt. The intrusions occur in a narrow area along the Balo Fault, a structural feature towards which the basalt thickens (>5 m) and which is represented by multiple flows. This observation indicates that the fault was active during the volcanism, and also that the basalt flows flooded a newly created accommodation area along a subsiding fault block east of the fault. No younger deposits of the Omo Group have been identified in the Loiyangalani region east of the Balo Fault. However, the Balo Basalt is also present west of the Balo Fault, where its surface reaches elevations of ~630 m, in contrast to a maximum elevation of ~550 m elevation east of the fault. Thus, approximately 80 m of displacement has occurred since eruption of the Balo Basalt 1.79 Ma ago.

Eruption of the Balo Basalt is believed to be the beginning of a tectonic episode that lasted less than 40 ka during which extensive erosion and narrowing of depositional area in the basin occurred. It is probable that major longitudinal faults along the basin's central axis (now under Lake Turkana) and border faults along the western margin of the basin were activated or reactivated during this episode, leading to a general tilt towards the active faults and a higher subsidence rate in West Turkana than in other regions.

These relations are consistent with geologic observations, which show the thickest onshore Lower Pleistocene section in West Turkana. By contrast, in East Turkana, an erosional feature commonly known as the “post-KBS unconformity surface” (Isaac and Behrensmeyer, 1997) was associated with ephemeral channels that extended westward from the basin margin.

The timing of this erosional event can be well approximated based on the local occurrence of the Orange Tuff (1.76 ± 0.03 Ma) stratigraphically only a few meters above the Steel Gray Tuff, which is placed right above the start of the associated tectonic episode. The magnitude of the event can be estimated based on the local differences in stratigraphic intervals between the Steel Gray Tuff and the KBS Tuff. The thickness changes from about 52 m in central Ileret to about 7 m in the nearby Karari escarpment (Gathogo, 2003; Brown et al., 2006; Gathogo and Brown, 2006). Ten of the 15 tuffs known from the KBS Member time sequence lie in the typically thinner interval above the erosional surface (Brown et al., 2006).

Here it is proposed that the a tectonic episode during KBS Member time led to significant narrowing of the basin’s depositional area and cessation of deposition along the easternmost (e.g., Karari and eastern Ileret) and southernmost (e.g., Loiyangalani) regions of the basin where deposits of the lower KBS Member time interval had begun. Depositional settings of the upper interval are characterized by diminished lacustrine conditions (particularly pelagic) at the expense of high-energy ephemeral channels from the basin margins. The most obvious manifestation of this transition occurs in the eastern part of the basin between the upper and lower KBS Member that are separated by the ‘post-KBS unconformity surface’. This transition is muted, or does not exist, in the northern part of the basin between Member H and Member J of the Shungura Formation, nor does it exist in the Nachukui Formation west of the lake. In the Nachukui Formation, discontinuities in sedimentation occur at different times in different locations (Brown et al., 2006).

Comparing the regional climatic proxy record from the Mediterranean Sea (Fig. 12) with the geological record in the Omo-Turkana Basin reveals that the most pronounced eccentricity maximum during KBS Member time occurred between 1.85 and 1.83 Ma, when many parts of the basin were already dominated by lacustrine conditions, particularly in West Turkana. However, these lacustrine conditions were continuous with the Lorenyang Lake of tectonic origin that occupied many parts of the basin during upper Burgi Member time. Nonetheless, climatic signals associated with precessional astronomical forcing may be identifiable in these deposits, and both Brown (1995) and Lepre et al. 2007) have made attempts to do so.

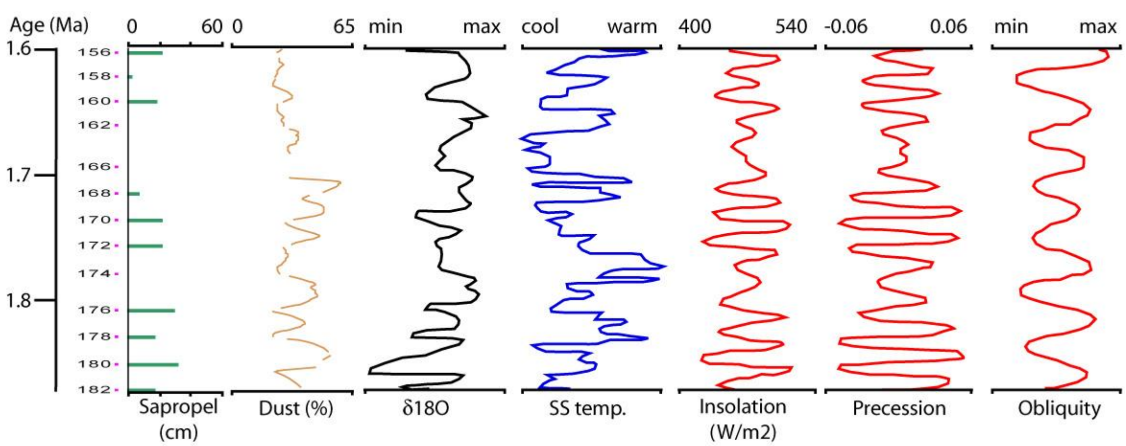

Figure 12. Climatic proxy records for KBS Member time (1.87–1.60 Ma) based on marine deposits from the Mediterranean Sea. Associated eccentricity cycles and predicted lake occurrence times are shown in Figure 4. See text for data sources and explanation.

The large deep lake discussed by Trauth (2005, 2007, 2009) and Maslin (2014; Fig. 3) whose beginning they place at about 1.9 Ma in the basin was already in existence because of tectonic movements by at least by 2.0 Ma when the global climatic conditions become favorable for lake formation (Fig. 4). They portray a large deep lake that persists to about 1.7 Ma, after which they show shallow alkaline lakes in the basin for the remainder of the KBS Member time interval. Work conducted for this dissertation found no field evidence for shallow alkaline lakes, but Feibel (2011) shows fluctuating lakes (his Fig. 3) following a major lake (Lorenyang Lake). The lake units perhaps correspond to the marginal lacustrine deposits described above among the fluvial dominated sections of East Turkana and in the Lower Omo Valley where they probably represent transgressive sequences of the Lorenyang Lake, subsequent shallow lakes, and/or expansive floodplain lakes. However, there is no evidence to show that these lakes were alkaline. Many shallow lacustrine strata of this interval contain abundant mollusks, which is reasonable evidence that the lakes were relatively fresh. These observations support the main hypothesis of this dissertation that major shifts in depositional facies in the basin took place only when tectonic activities allowed, including the development of prominent lake sequences such as the Lorenyang Lake.

The main tectonic episode during the KBS Member time interval produced a significantly different effect in relation to depositional settings compared to other Pliocene tectonic events. No tectonic lake developed in the basin during this time. Instead, lacustrine conditions were diminished and extensive erosion by ephemeral channels was amplified, particularly in East Turkana. It is possible that structural movements associated with this tectonic episode created a large but deep accommodation volume along a narrow zone within the basin's axial center (now under Lake Turkana) that was too voluminous to be rapidly filled by the perennial axial system of the proto-River Omo. Such a scenario is consistent with field evidence that the highest subsidence rate was in the western part of the basin. The thickness of the Omo Group under Lake Turkana increases rapidly (up to about 4,000 m) along north-south-oriented longitudinal faults, for which the timing of movements is not yet known (but see Bruhn et al. (2010) for discussion).

Okote Member Time: 1.60–1.38 Ma

Deposits from the Okote Member time interval are known for their richness in the archeological findings, especially stone tools, particularly at Karari and Ileret (Isaac and Isaac, 1997). Recent hominin findings from these deposits from Ileret and include a partial face attributed to *Homo habilis* that is younger (1.44 Ma) than previously known (Spoor et al., 2007) as well as footprints made by a hominid whose foot was essentially anatomically modern (Bennett et al., 2009). Mammalian fossils are not as common as they are in the underlying sequence. Strata of this time interval exhibit a unique change in depositional settings compared with other intervals of the Omo Group.

Two depositional settings are recognized in strata of the Okote Member time interval: i) delta plain settings dominated by tuffaceous silts locally interrupted by sandy to pumiceous distributary channels of the perennial axial system and minor ephemeral channels from the basin margins that grade locally into submerged delta or minor playa lakes; and ii) silt- and clay-dominated fluvial settings of the axial fluvial system alternating with minor deposits of fluvial lakes and/or ephemeral streams from the basin margins. Tectonic influences are subtle and climatic effects are fairly pronounced in these depositional settings.

In the Koobi Fora region, sedimentation during Okote Member time began with a deltaic phase characterized by an influx of volcanic ash and biotite mixed with silt and very fine sand, but typically poor in clay minerals. Pedogenic features are not well developed in strata of this age; instead, in many locations, the deposits are loosely packed and have microtubules (about 0.2 mm in diameter) that represent possible grass rootlets. Channel-form units normally lack coarse grains and basaltic clasts, but rhyolitic pumice pebbles are very common. So far as is known, deposition of this sequence took place only between the southern part of the Lower Omo Valley (de Heinzelin and Haesaerts, 1983a), the northern parts of West Turkana (Harris et al., 1988a), and the northern half of East Turkana (Brown and Feibel, 1986; Gathogo and Brown, 2006). Even within this limited depositional area of ~6000 km^2, sediment accumulation by the deltaic system of the perennial Omo River shifted episodically between subregions such as northern Ileret, southern Ileret, Karari, or Koobi Fora (Brown et al., 2006). For example, some tuffaceous parts of the Okote Member sequence in the Karari region (the “Okote Complex”; Brown and Feibel, 1986) older (or younger) than the strata present near Ileret (“Ileret Complex”). These three groups of tuffaceous beds are separated by the Black Pumice Tuff, which is approximately 1.53 Ma old. Most archeological sites in the Koobi Fora region (Isaac and Isaac, 1997) come from the “Okote Complex” on the Karari Ridge.

Southern exposures of the interval below the Black Pumice Tuff in the Koobi Fora region (Brown and Feibel, 1986) and also at Lowarengak (West Turkana; Harris et al., 1988a) contain thin beds of algal stromatolites, which are interpreted here as indications of clear shallow water, possibly on a perennially submerged delta plain or along the margins of a shallow playa lake. The younger “Ileret Complex” in the southern Ileret region is confined between thin beds (10–20 cm thick) of ferruginous carbonate concretions (Gathogo and Brown, 2006). The bottom carbonate bed is interpreted as representing a period of a persistently shallow phreatic surface when sediments of the deltaic system of the proto-Omo River bypassed this northern region for the Karari and Koobi Fora region to the south. (This paleogeographic arrangement may have been caused by minor local tectonic movements since older strata occur in the region north of Il Eriet.) Aquatic invertebrates are generally rare in the Ileret sequence, with occurrences of *Pila ovata* and *Etheria elliptica* (pers. obs.) that indicate the presence of intermittent to perennial freshwater, respectively. “Loessic” deposits that lack stratification are reported farther north at Errum, south of the main Shungura exposures (de Heinzelin and Haesaerts, 1983a).

Fluvial intervals deposited during the later part of Okote Member time have features similar to those of fluvial strata in the KBS Member, except that they do not interfinger with biolithic beds. These deposits have produced most of the mammalian fossils from the Okote Member time interval, including thousands of unpublished specimens from ongoing field research led by M. G. and L. N. Leakey in the Ileret area. This interval is dominated by clay-rich floodplain and channel deposits of the perennial system of the proto-Omo River, and preserves *E. elliptica* in many localities. These deposits are interbedded with ephemeral channel deposits from the basin margins typically poor in clay, but with coarse material that include fine pebbles of carbonatecemented silt and local basaltic clasts (Gathogo and Brown, 2006). Some sandstones from the perennial fluvial system grade into beds that contain scarce to abundant fish bones and traces of fish nests (*Piscichcnus*), for example the ‘Main Fish Bed’ in the Ileret area (Cerling, 1977; Brown and Feibel, 1986; Gathogo and Brown, 2006). Intervals with *Piscichcnus* are interpreted as floodplain lakes created by periodic flooding of the axial perennial river system of the proto-Omo River.

Field studies of outcrops of the Okote Member time sequence indicate that patterns of the distribution of lithologic facies were insignificantly influenced by tectonic developments in the Omo-Turkana Basin. Structural features such as syndepositional faults are very rare, minor, and localized (Gathogo, 2003). Lithologic features identified in various parts of the basin representing the Okote Member time interval seem to have been considerably influenced by astronomically modulated climatic cycles (Fig. 4). For example, eolian deposits, reduced sediment load, lack of clay, and limited channel elements of the axial perennial system coincide with a climatic period before 1.47 Ma when monsoon precipitation was reduced as indicated by paleoclimate proxies from the Mediterranean Sea (Fig. 13). The appearance of local fluvial lake deposits, particularly in the Ileret area, coincides with precession-scale climatic changes beginning 1.47 Ma ago when increased monsoonal precipitation is indicated by the occurrence of relatively thick sapropel layers in the Mediterranean Sea, and pronounced peaks in sea surface temperature. This period of sapropel formation follows a 90 ka interval during which no sapropels are known from the Mediterranean Sea (Kroon et al., 1998). These observations are consistent with the illustrations of Trauth et al. (2005, 2007, 2009) and Maslin et al. (2014) who show no lake sequences in the basin during this time interval.

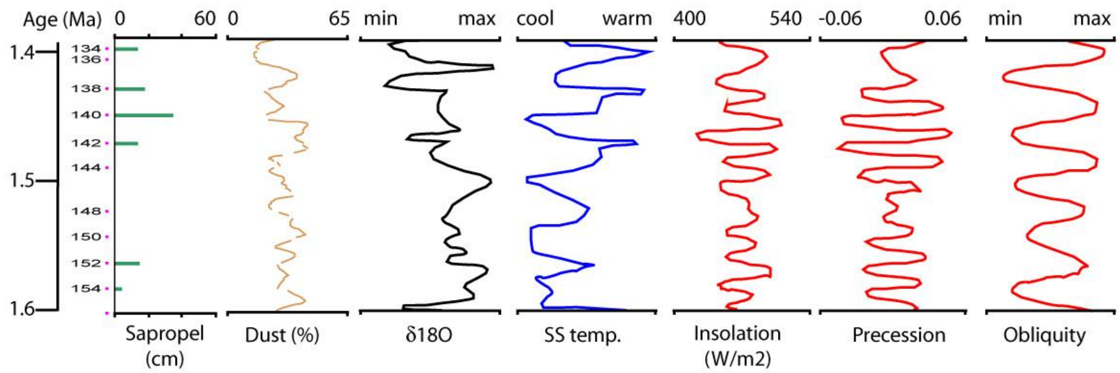

Figure 13. Climatic proxy records for the Okote Member time interval (1.60–1.38 Ma) based on marine deposits from the Mediterranean Sea. Associated eccentricity cycles and predicted lake occurrence times are shown in Figure 4. See text for data sources and explanation.

However, they do depict two eccentricity maxima between 1.48 and 1.35 Ma for which they do not suggest that lakes should have formed. On the basis of new information about the tectonic history of the basin (Fig. 4), it is predictable that no significant lacustrine facies developed during this time interval.

Chari Member Time: 1.38 Ma to ~0.7 Ma

The Chari Member time interval is the poorest represented geologic unit in the Omo Group sequence considering its duration (at least 700 kyr) and thickness (up to 93 m, in northern Ileret; Gathogo and Brown, 2006). Further, these deposits are poorly exposed, but generally represent depositional settings (mostly lacustrine, followed by deltaic) that were not favorable habitats for many mammals. The few mammalian fossils from this time interval make up the last appearance datum for many taxa, including the hominin *A. boisei* (or *Paranthropus*) (Feibel et al., 1989; Wood, 1991). The age assignments for these last occurrences have been coarse, either being the mean age (1 Ma) or the age of the youngest dated tuff (Silbo Tuff; about 0.7 Ma; see Table 1). The latter approach has been commonly used (Behrensmeyer et al., 1997) even though most mammalian fossils from this time interval are associated with fluvial deposits that underlie the Lower Nariokotome Tuff (1.298 ± 0.025 Ma; Table 1). Limitation of exposure makes it difficult to evaluate the effect of climatic cycles on deposition in the basin during this interval. At least one tectonic episode that generally coincides with major changes in depositional settings occurred between 1.3 Ma and 0.7 Ma, but more likely close to the older limit of this range.

Three prominent depositional settings are identified within the Chari Member time sequence: i) fluvial moderate energy deposits including sandy mudstone to brown clayey mudstone with well-developed pedogenic features; ii) high-energy alluvial to beach deposits including pebbly sandstones and conglomerates grading to bivalve-rich sandstones; and iii) low-energy, mostly pelagic lacustrine, pale olive claystones to siltstones interbedded with marginal lacustrine sandstones that are rich in quartz and ostracods (de Heinzelin and Haesaerts, 1983a; Brown and Feibel, 1986; Harris et al., 1988a; Gathogo and Brown, 2006). The latter are locally associated with beds of oolites and algal stromatolites.

The stratigraphy of these deposits is poorly understood since dated or correlated stratigraphic units (e.g., volcanic ash beds) are few. Most of the marker units occur near the bottom of the interval between the Chari Tuff (1.383 ± 0.028 Ma) and the Upper Nariokotome Tuff (1.230 ± 0.020 Ma), and at the top of the interval between the Silbo Tuff (0.751 ± 0.022 Ma) and the Kale Tuff (see Table 1). Most of the intervening section down to the Chari Tuff is preserved in only a few locations where the entire interval is relatively thin; for example, in the Lower Omo Valley (49 m; de Heinzelin and Haesaerts, 1983a) and in most of the Ileret area (45 m; Gathogo, 2003). The whole interval is represented by only the Chari Tuff in many locations in eastern Ileret and along the Karari ridge. This pattern of outcrop and stratal representation can be explained by syndepositional structural movements associated with at least one tectonic episode described below, and also by concentration of deposition in the central part of the Omo-Turkana Basin now occupied by Lake Turkana.

The stratigraphy and chronology of the strata laid down during Chari Member time has not been extensively discussed, but it is intricate. For this reason, a somewhat extended discussion is offered below, which is necessary to understand how lacustrine deposits in this interval may or not be related to tectonic or climatic causes.

In order to complete revision of the stratigraphy of the Koobi Fora Formation, Brown and Feibel (1986) defined the uppermost member of the formation (the Chari Member) to include the Chari Tuff and sediments above it up to the base of the Holocene Galana Boi Beds. In so doing, they included strata that were originally named the Guomde Formation (Bowen and Vondra, 1973). Brown and Feibel (1986) recognized a fluvial channel complex above the Chari Tuff that, in places eroded through the Chari Tuff. There are few tuffs within the Chari Member in most exposures east of Lake Turkana, but these include the Gele Tuff (1.326 ± 0.019 Ma), and the Silbo Tuff (0.751 ± 0.022 Ma) in outcrop areas near Ileret township, and the Lower and Upper Nariokotome tuffs farther north and east of Il Eriet.

West of Lake Turkana, the Chari Tuff is present in only a few sections, but the Lower Nariokotome Tuff is very continuous and widespread. For this reason, Harris et al. (1988a) defined the uppermost member of the Nachukui Formation as strata lying above the basal contact of the Lower Nariokotome Tuff, again excluding overlying Holocene strata.

In the Lower Omo Valley, Brown et al. (1970, p. 249) formally defined the Shungura Formation as:

> The series of fluvial, deltaic and lacustrine sands, silts, and clays with intercalated tuffs of late Pliocene to Early Pleistocene age, having a thickness of about 600 meters, and being typically exposed between latitudes 5º and 5º10’ N on the west side of the Omo River north of Lake Rudolf in Ethiopia….

The concept of the Shungura Formation was later extended to include higher stratigraphic levels by de Heinzelin (1971), and still later the formation was divided into members on the basis of widespread tuffs by de Heinzelin et al. (1973) who defined the highest member as Member L, which was divided into nine units above Tuff L by de Heinzelin and Haesaerts (1983a).

Above the Chari Tuff (1.383 ± 0.028 Ma), chronological control is poorer than is the case for older units, and sedimentation in the Ileret area was interrupted by at least one disconformity ~1.5 m above the Chari Tuff. The Gele Tuff (1.326 ± 0.019 Ma), and the Lower, Middle, and Upper Nariokotome tuffs (1.326–1.220 Ma), do provide quite tight control on the age of the section immediately above the Chari Tuff, but above that there is only a single unit, the Silbo Tuff, with a $^{40}Ar/^{39}Ar$ age of 0.751 Ma, higher in the sequence. Paleomagnetic polarity study of the Chari Member time interval is limited to that of Hillhouse et al. (1986), who showed that the lower part of the Chari Member at Ileret is of reversed polarity, but the upper part (including the Silbo Tuff) is of normal polarity. Hillhouse et al. (1986) interpreted the transition from reversed (below) to normal (above) as having recorded the Brunhes-Matuyama transition, which has an age of 0.781 Ma (Gradstein et al., 2004). They also noted that a disconformity above the Chari Tuff must thus have a duration on the order of 0.5 Ma.

The thickest sections of the Chari Member are at Ileret, but thicknesses vary significantly from one location to another. Brown and Feibel (1986) gave a thickness of 42.7 m in the type section of the member, and for the same general area, Gathogo (2003) recorded a thickness of 47 m. Gathogo and Brown (2006) discuss ostracod-packed beds, laminated to thin-bedded olive claystones, and a number of mollusk-rich layers in the section below the Silbo Tuff, and extending upward to slightly above the Kale Tuff. Laterally, diatomites are also present in the northern part of the exposures south of Il Eriet. These strata may be the evidence used by Maslin and Trauth (2009) to suggest the presence of a lake with an age of ~1 Ma, but the age of these beds is significantly younger (ca. 0.77 to 0.73 Ma), as is known from the date on the Silbo Tuff, and the fact that the section is of normal paleomagnetic polarity.

North of Il Eriet, the Chari Member is at least twice as thick (>93 m; Gathogo & Brown, 2006) as it is south of Il Eriet. The Lower and Upper Nariokotome tuffs are present low in this section, but the Silbo Tuff has not been identified. Thin layers rich in molluskan fossils, ostracodites, and oolites are interspersed in this section, but overall the record is one of fluvial deposition, perhaps with floodplain or deltaic lakes. No paleomagnetic work has been done on this part of the Chari Member, but as the Silbo Tuff is absent (though quite widespread over much of the northern part of the basin), it is probable that this section lies below the disconformity that separates the Chari Tuff from the section associated with the Silbo Tuff south of Il Eriet. Assuming that this is correct, there is then no evidence for a pelagic lake in this area for much of the time between 1.23 Ma and 0.78 Ma. Without question, additional work is needed to place this section more firmly temporally. Nonetheless, the discrepancy in thickness (~1.5 m below the disconformity south of Il Eriet vs. 93 m north of Ileret) is good evidence that much section was removed from the southern exposures of the Chari Member sometime between 1.38 and 0.75 Ma, and the reason for this erosion is most likely uplift of the southern exposures. Strata associated with the Silbo Tuff may be taken as evidence for a short-lived pelagic lake about 0.7–0.78 Ma ago.

West of Lake Turkana, the situation is different. There the Chari Tuff lies ~12 m below the Lower Nariokotome Tuff, which is immediately overlain by a sequence with oolites and ostracod-rich fine sandstones below the Middle Nariokotome Tuff. A well-cemented mollusk-packed sandstone lies above this tuff, followed by 12 m of fluvial section, a sandstone (~2 m thick) with decalcified mollusks and then a 57-m- thick section of alluvial gravels with minor interspersed ostracod-rich siltstones and stromatolite layers. Similar deposits are present in the Nachukui drainage, and these may be what prompted Feibel (2011) to label his Nachukui lake phase, although these deposits are not regarded here as those of a pelagic lake. Above the gravels, the section continues through mainly fluvial sandstones for another 20 m, up to a sandstone containing sparse mollusks at the top. The top 10 m of section are reddish brown mudstones, and included within them is the Silbo Tuff, identification of which has been confirmed by dating of anorthoclase from pumice clasts collected from this section (Ward et al., 2013). No paleomagnetic information has yet been published for this section, but eight samples taken from the Nariokotome Tuff upward to the mollusk bed are of reversed polarity, whereas the mudstones associated with the Silbo Tuff are of normal polarity (F. H. Brown, pers. comm.). The lowest of the intervals with abundant invertebrate fossils must be very near 1.25 Ma in age. In the northernmost known sections of the Nariokotome Member, there is a mollusk-rich sandstone associated with the Silbo Tuff that may correspond to similar strata associated with the Silbo Tuff at Ileret.

In the Lower Omo Valley, the record is also different. There, strata above Tuff L, which correlates with the Chari Tuff, are about 47 m thick (de Heinzelin and Haesaerts, 1983), and the section records a short normal magnetozone in submembers L-5 and L-6 (Brown et al., 1978). Mollusk-rich sandstones and ostracodites lie above this normal magnetozone in submembers L-7 and L-9. The age of these mollusk-rich layers can be either between 1.13 and 1.17 Ma if the normal magnetozone is correlated with the Cobb Mt. Event, or between 0.78 and 0.98 Ma if the normal magnetozone is correlated with the Jaramillo Event. No lacustrine strata are otherwise known from the base of Member L to submember L5, and submember L-4 contains patches of *Etheria* that are most likely record a river channel at that time. Thus there is no evidence for a lake in the Lower Omo Valley between 1.38 and 1.18 (or 1.07) Ma, and if the mollusk-rich sandstones are taken as evidence for a pelagic lake (not done here), then that lake must be either 1.13–1.17 Ma old or 0.78–0.98 Ma old.

In summary, evidence for a pelagic lake during Chari Member times is meager at best, but is more compelling for strata associated with the Silbo Tuff. If this reasoning is correct, then the Omo-Turkana Basin does not record pelagic lacustrine conditions suggested Maslin and Trauth (2009) at 1.0 Ma, instead possibly recording the presence of a lake at ~0.75 Ma or at 1.13–1.25 Ma, at which times Maslin and Trauth (2009) suggest that no lake should exist. On the other hand, the disconformity between the lower part of the Chari Member, and the great discrepancy in thickness of this member north and south of Il Eriet (a distance of ~ 5 km), certainly suggest that there was uplift of the southern part of this area sometime between 1.33 (the age of the Gele Tuff) and 0.78 Ma. Whether the possible pelagic lake associated with the Silbo Tuff can be related to this uplift remains to be seen.

Deposition during Chari Member time continues the preceding sequence of the Okote Member time interval and began in fluvial settings that were dominated by the axial perennial system of the proto-Omo River. A fining upward sequence with pale brown siltstones in northern West Turkana represents distal floodplains and levees of the main channel, although pumice clasts also locally occur in the Chari Tuff. In contrast, the channel-form Chari Tuff with large pumices in northeastern parts the basin at Ileret and on the Karari Ridge indicates proximity to the perennial river. The presence of riverine *E. elliptica* in sandstone associated with the channel-form Chari Tuff in the Ileret area confirms the perennial nature of this river. The correlative interval upstream in the Lower Omo Valley at Kalam is quite similar to the Ileret deposits, except that the correlate of the Chari Tuff there is very thin, and has not been found filling channels. Similar fluvial settings seem to have been common until shortly before deposition of the Lower Nariokotome Tuff (1.298 ± 0.025 Ma) based on the geology of the northern part of the Nachukui Formation in West Turkana (Harris et al., 1988a) and of the Koobi Fora Formation at Ileret (Gathogo, 2003).

Fluvial deposits of the lower Chari Member time sequence were eroded in many places following a tectonic episode that is believed (this dissertation) to have begun shortly before deposition of the Nariokotome Tuff suite (1.3-1.2 Ma) and that culminated shortly afterwards with substantial faulting and a regional disconformity surface that locally grades to an angular unconformity, for example in exposures in western Ileret. On the basis of this disconformity, early workers (Bowen and Vondra, 1973) assigned the upper Chari Member interval in East Turkana the ranking of a formation (Guomde Formation). The main evidence for this tectonic episode is described below.

The first evidence for tectonic movements in this time interval involves disparity in stratal thickness over short distances. For example, deposits forming the interval between the Chari Tuff and the Nariokotome Tuff suite attain at least 23 m in thickness at two locations in western and northern Ileret, but deposits are absent in locations only a few tens of meters away (Gathogo, 2003). An angular relationship between the Chari Tuff and deposits of the upper Chari Member interval is evident in a few places in western Ileret where the lower interval is missing. The correlative lower interval is 9 m thick in West Turkana where the Nariokotome Tuff suite is defined (Harris et al., 1988a). In the Lower Omo Valley, the lower interval is best exposed at Kalam where the sections measure between 11 m (at Naito) and 15 m (at Iryamanyang) between the bottoms of Tuff L (= Chari Tuff) and submember L-4 (de Heinzelin and Haesaerts, 1983a). The Gele Tuff and Nariokotome Tuff suite are missing, or have not yet been positively identified in the Lower Omo Valley, but structural and depositional settings that are indicated by submember L-4 tally well with observations made in Ileret and West

Turkana at the top of the lower interval. de Heinzelin and Haesaerts (1983a; p. 126) described submember L-4 as a "disconformable lake shore" at Naito.

The second line of evidence for tectonic disturbance comes from a rapid change in depositional settings accompanied by basaltic conglomerates deposited on ancient alluvial fans even closer to the basin's axial center. In East Turkana, the conglomeratic deposits with basalt clasts are exposed in the western and northern parts of Ileret where they lie above the Gele Tuff and just above the Nariokotome Tuff suite, respectively (Gathogo and Brown, 2006). In Area 131, a thick (~12 m) basaltic pebble conglomerate lies below the Chari Tuff. The western Ileret conglomerates mark the first occurrence of basaltic pebbles this close to the basin center, and these pebbles require considerable energy for transport. In the northern Ileret region, the conglomeratic deposits are overlain by strata containing the lacustrine-form of *E. elliptica* that are succeeded by ostracodites and oolites. Chari Member strata are represented by only the Chari Tuff in many locations in the northern part of East Turkana, including eastern Ileret and throughout exposures along the Karari Ridge. The missing section is a possible indication of areas that were relatively uplifted and isolated from deposition and/or affected by increased erosion following the tectonic episode. Farther northwest in the Lower Omo Valley, submember L-4 of the Shungura Formation consists of high-energy deposits, including thick, coarse conglomerates with rolled bones and *E. elliptica* reefs laid down on a disconformable lake shore (de Heinzelin and Haesaerts, 1983a). Correlative deposits in West Turkana consist of basaltic conglomerates interbedded with coarse-grained sandstones, particularly at Nariokotome (Harris et al., 1988a), and in the upper reaches of the Nachukui drainage. The proximity of some conglomeratic outcrops in West Turkana to basaltic hills along the western margins of the basin in part suggests an inherent geographic phenomenon; however, the abundance of basaltic clast conglomerates in coeval deposits in other parts of the basin that are distant from volcanic hills also suggests an increase in erosion and depositional energy related to tectonic adjustments (Harris et al., 1988a).

The descriptions presented above are consistent with tectonically related transformation in depositional settings. As discussed, the first change involved replacement of the axial fluvial system of the perennial proto-Omo River with local alluvial fan drainage systems in many locations. The next step was the introduction of shallow lake settings that were subsequently replaced by a pelagic lake (with *E. elliptica*) conditions that are described below.

Sections recording Chari Member time consist mainly of marginal lacustrine deposits that represent the 1.30–0.75 Ma interval between the Nariokotome Tuff suite and the Silbo Tuff. The geology of this interval is poorly understood because a large proportion of section is covered by later alluvium. Exposures of these deposits are thicker (at least 70 m) in northern Ileret (Gathogo and Brown, 2006) and West Turkana (Harris et al., 1988a) than in other areas (30–40 m), including southern Ileret (Brown and Feibel, 1986; Gathogo, 2003) and the Lower Omo Valley. Lacustrine strata in submember L-9 in the lower Omo Valley either postdate the Cobb Mountain Normal Subchron (1.173–1.185 Ma; Table 2; de Heinzelin and Haesaerts, 1983a) or the Jaramillo Normal Subchron (see above). The main pelagic lake settings are represented by olive green to olive brown claystones interlaminated with siltstones and with abundant fish remains. This interval is confined by biolithic units that are packed with mollusks, ostracods, algal stromatolites, or ooids, all of which represent shallow lake settings adjacent to a beach (de Heinzelin and Haesaerts, 1983a; Brown and Feibel, 1986; Harris et al., 1988a, b; Gathogo and Brown, 2006).

The final interval of the Omo Group began with resumption of deposition in deltaic to fluvial settings associated with the axial proto-Omo River system at least shortly before 0.751 ± 0.022 Ma, when the Silbo Tuff was deposited with pumices in various parts of the basin, including Ileret (Brown and Feibel, 1986) and near Shin (Cerling and Brown, 1982; McDougall and Brown, 2006) in southeastern Koobi Fora. The Silbo Tuff was deposited on a surface of eroded outcrops of various intervals of the Omo Group sequence. For example, near Ileret, the tuff with its associated pumices overlies deposits of shallow lake, pelagic lake, and deltaic settings in the middle Chari Member time sequence. Farther southeast, near Shin, the pumiceous part of this tuff overlies the Malbe (1.84 Ma) and Black Pumice (~1.53 Ma) tuffs and associated strata. Deposits of this younger interval are exposed over much smaller areas than is the rest of the Omo Group sequence. One such area is central Ileret where the upper interval begins in deltaic settings of the Omo River that were briefly interrupted by shallow lakes (Gathogo, 2003). Shortly after deposition of the Kale Tuff, the prevailing depositional settings were replaced by fluvial conditions in which expansive floodplains with well-developed soils contain mammalian fossils (e.g., bovids, equids, and elephantids) in Ileret and parts of West Turkana near Nariokotome. Shallow lake settings were reestablished at least in parts of the Ileret region before deposition of the Omo Group sequence ended in the basin, except in the area now covered by Lake Turkana. The end of Omo Group deposition appears to have been associated with the last major tectonic episode since the

Turkana Group is deposited on tilted beds of the Omo Group (Butzer, 1971, 1980; de Heinzelin, 1983; Isaac and Behrensmeyer, 1997).

Examination of proxy data for astronomical climatic forcing shown in Figure 14 reveals that the lower fluvial sequence (below the Nariokotome Tuff suite; >1.30 Ma) of the Chari Member time interval coincides with a period of extreme climatic conditions, including increased monsoonal rainfall as indicated by the high frequency of relatively thick sapropel layers. Lack of a lacustrine sequence in the basin during this period supports the main hypothesis of this dissertation: the right tectonic conditions are required in order for lacustrine settings to develop. Conditions changed probably because of tectonic deformation beginning shortly before deposition of the Nariokotome tuffs and at that time extensive lacustrine settings are established in many parts of the basin (Fig. 4).

Very limited geological data are available for this interval of the Chari Member time sequence, but the record indicates that the lacustrine sequence was continuous with shallow lake conditions associated with the Nariokotome Tuffs. There is evidence for these conditions in Ileret (Area 40) east of the lake, and in the western part of the basin from Nareng'ewoi northward. It is possible that mollusk-packed sandstones near Naito in the Lower Omo Valley, and at Iryamanyang still farther north, correlate with these, but the two probable chronological placements argue against this. These stratigraphic records imply that lacustrine settings prevailed in the basin and even developed into a large lake despite the subdued climatic conditions associated with reduced monsoonal rainfall at least between 1.3 Ma and 1.2 Ma (see Figs. 4 and 14). Lack of extensive outcrops makes it difficult to confirm whether the main lake existed during the extreme

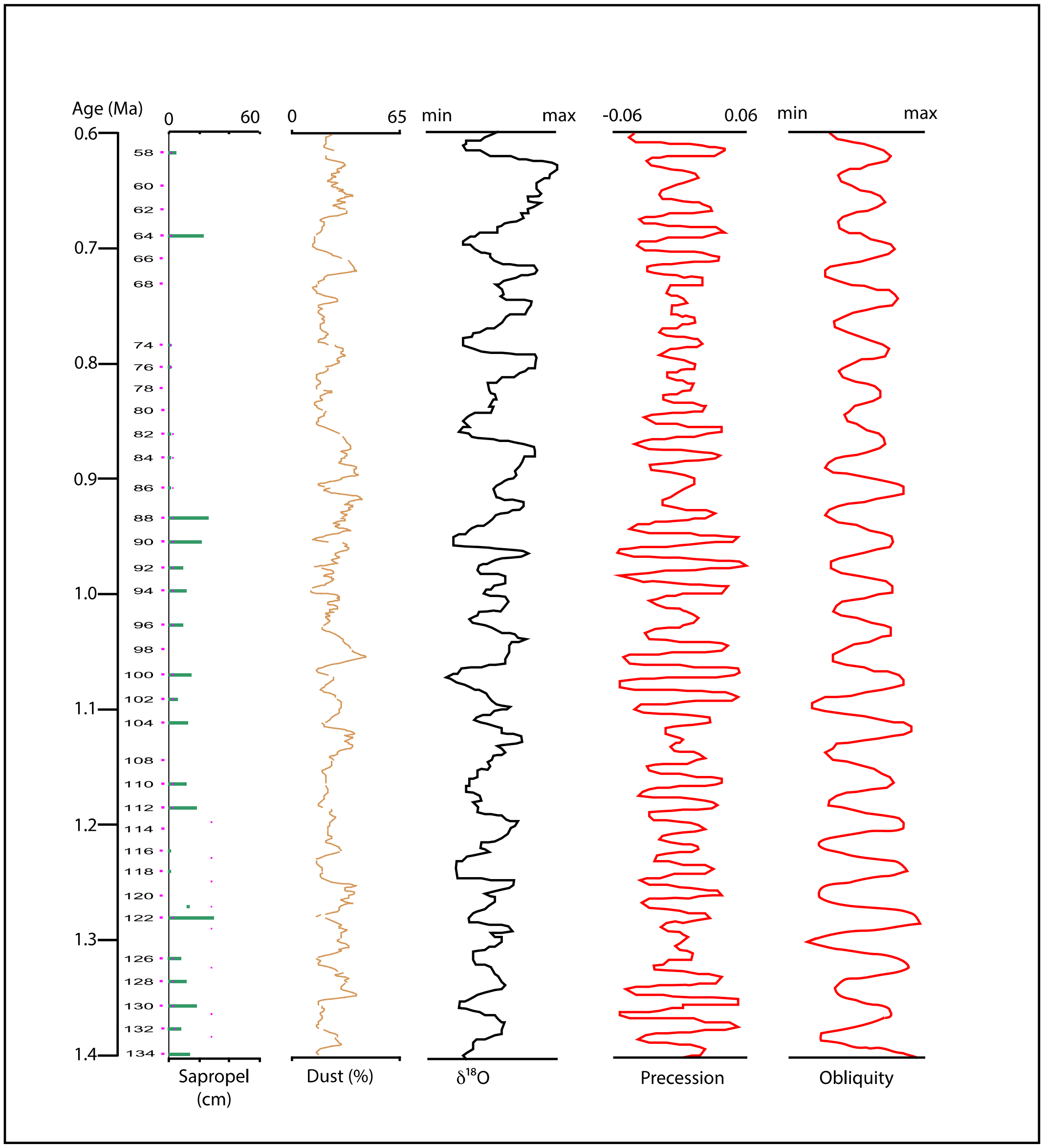

Figure 14. Climatic proxy records for the Chari Member time interval (1.38 Ma to Middle Pleistocene) based on marine deposits from the Mediterranean Sea. Associated eccentricity cycles and predicted lake occurrence times are shown in Figure 4. See text for data sources and explanation.

climatic conditions associated with a period of eccentricity maximum 1.09–1.04 Ma, but stromatolite layers are widely developed in the Nachukui Formation at about this time associated with volcanic conglomerates that only extend eastward into the basin after deposition of the Nariokotome tuffs.

Feibel (2011) places a major lake phase (his Nachukui Lake phase) stratigraphically between his Chari and Silbo floodplains (his Fig. 3). This lake corresponds in part to the pelagic lacustrine sequence described above between 1.3 Ma and 1.2 Ma. Trauth et al. (2005, 2007, 2009) and also Maslin et al. (2014) show a shallow alkaline lake starting slightly after 1.1 Ma and ending slightly after 0.9 Ma in the basin. Their chronological placement does not correspond with either possible placement of the deposits of Submember L-9 of the Shungura Formation, or with the age of lacustrine deposits associated with the Nariokotome Tuff. The scarcity of outcrops that represent the time between 1.2 Ma and 0.8 Ma makes it difficult to confirm the duration of any lake in this interval. Finally, Trauth et al. (2005, 2007, 2009) and Maslin et al. (2014) show no lake in the basin after 1.1 Ma, but that is incorrect because a pelagic lake occurred at least in the interval between 0.78 Ma and 0.70 Ma associated with the Silbo Tuff, which lies within the Bruhnes Normal Chron Ma).

Even with limited outcrop evidence for this time interval, the observations presented above support the main hypothesis of this dissertation that major shifts in depositional facies in the basin took place only when tectonic activities allowed. The main lacustrine sequence coincided with a major tectonic event, but favorable climatic conditions that occurred later may have helped sustain or expand the lake(s).

CHAPTER 3

DISCUSSION

This section focuses on the third and final objective listed in Chapter 1: a conceptual geologic model that incorporates both rift tectonics and climate to explain the depositional environments of the Omo Group sequence and its mammalian fossil record. The following discussion presents the conceptual geologic model under five related topics: i) onset of the Omo Group sequence; ii) mammalian fossil record, focusing on geographic distribution; iii) climatic controls, primarily precipitation; iv) geographic aspects, primarily regional physiography; and v) rift tectonics, primarily structural movements and basalt volcanism.

The conceptual model that is presented here focuses on the interpretation of the sedimentary record as a proxy for relative uplift and subsidence, particularly in association with surface faulting. It also draws inferences from previous studies. Detailed assessment of subsurface structural features will have significant implications for the mechanism or processes (Morley et al., 1999a) associated with rift tectonics, but this is beyond the scope of this study as such assessment would require a specialist in geophysical modeling and simulations. Structural studies focusing on subsurface geology have been conducted in surrounding regions including the Main Ethiopian Rift

(WoldeGabriel et al., 1990), Chow Bahir (WoldeGabriel and Aronson, 1987), and Kino Sogo (Cerling and Powers, 1977; Vétel et al., 2005). Integration of such regional studies with the geology of the Omo-Turkana Basin will require detailed studies on the sedimentary geology of the corresponding basins. This type of integration is recommended in order to develop a robust regional model.

Onset of the Omo Group Sequence

General physiography in the larger part of the Omo-Turkana drainage basin away from the present outcrop area of the Omo Group could have been quite similar to the present with major volcanic highlands to the north (Ethiopian highlands) and south (Elgon and Cherangani). Evidence includes the gorge of the Omo River north of the Lower Omo Valley, which predates the Omo Group. At some point in its history, the basin must have been connected to regions now (if not then) drained by the Nile, because there are fossil faunal elements (e.g., *Euthecodon, Lates*) common to the Omo-Turkana Basin and the Nile drainage, but generally absent from other fluvial systems in Africa. Field observations and related interpretations that are presented in this dissertation and elsewhere indicate that the Omo-Turkana basin became a coherent depositional feature in the early Pliocene. The proto-Omo River become a dominant, axial perennial fluvial system in the basin and fed an episodic lacustrine system (proto-Lake Turkana). Deposition in the basin remained relatively continuous while shifting to various regions up to 40 km away from the basin's longitudinal axis, with cumulative tectonic subsidence reaching up to 4,000 m. At present, the Omo-Turkana Basin is still a tectonically active structural trough along the northern part of the East Africa Rift System, as is Sanderson's Gulf, judging from the results of Mammo (2012).

Outcrop geology of rocks underlying the Omo Group sequence is poorly known, with most units being Miocene or older basalts (Brown and Feibel, 1986; Harris et al., 1988a; Gathogo and Brown, 2006). The sequence unconformably overlies basement metamorphics at a few locations (Gathogo et al., 2008). Upper Miocene and possible Lower Pliocene (> 4.3 Ma) fluvial sediments are described in the southern parts of the basin at Lothagam, possibly from a predecessor of the Kerio River (McDougall and Feibel, 1999, 2003). Moreover, deposits of both the Kerio and the Turkwel Rivers occur in the southern parts of the basin. Major structural features, including the Lokichar Fault along the western parts of the basin and the Balo Fault along the eastern part of the basin, could have been variably active (Gathogo et al., 2008) during the Pliocene and Pleistocene, as is shown by the presumed "O" horizon of Morley et al. (1992) in the North Lokichar Basin.

The geologic record indicates downwarping of the present area of Lake Turkana occurred ~4.5 Ma ago (Bruhn et al., 2010). Significant deposition in the Omo Group began in local drainage subbasins about 4.3 Ma based on age controls from the southern parts of the basin at Kanapoi and Lothagam, and also in the northeastern part of the Koobi Fora region (Kidney, 2012). The onset of deposition in the basin coincided with major structural movements in the basin, possibly marking the initial stages of basinwide tectonic inversion and reactivation of major structures (Morley et al., 1992; LeGall et al., 2005), including the Balo Fault at Loiyangalani (Gathogo et al., 2008). Other structural features that became active at this time include border faults along the western margin of the basin such as the fault at Kosia (a probable extension of the Lokichar Fault) and also the Lomekwi Fault. Findings presented in this dissertation support a mechanism suggested by Haileab et al. (2004) for the initial stages of basin development in the Early Pliocene. The mechanism considers a short series of major rift-related movements that led to rapid basin extension followed by extensive subsidence along the axis of present Lake Turkana. High local topography during this initial stage is indicated by alluvial fan deposits with conglomerates and pebbly channel gravels that record the proximity to fault scarps. An increased deposition rate was triggered by movement along the faults. The increased rate is represented by fine-grained deltaic to lacustrine sequences, with the latter being intruded and locally flooded by basalt flows.

Results presented here indicate that Pliocene and Pleistocene tectonic activity in the Omo-Turkana Basin has been quite variable, but with distinct episodes locally marked by syndepositional faulting and basalt volcanism, at least during the Pliocene (Gathogo et al., 2008). These episodes also coincide with systematic changes in depositional settings. For example, low-energy settings such as pelagic lake deposits occur adjacent to high-energy settings such as alluvial fan and stream deposits. Changes in local depositional settings are manifestations of readjustment of hydrologic patterns as dictated by tectonic-induced modification of the general physiography. Some prominent ephemeral rivers that presently drain basaltic and metamorphic terrain along the eastern and western margins of the basin developed in the early Pliocene coinciding with episodic tectonic events. Modern ephemeral rivers that are recorded in local sequences include the Topernawi (Harris et al., 1988a; Leakey et al., 2001) in West Turkana and Il Eriet (Gathogo, 2003; Gathogo and Brown, 2006) in the Ileret region of East Turkana. Other prominent ephemeral rivers in West Turkana (e.g., Nariokotome, Kokiselei, Kataboi, and Kosia), East Turkana (e.g., Il Kimire, Il Alia, and Il Lokochot), and Loiyangalani (e.g., Balo) also represent local drainage systems that developed primarily due to the relative uplift of basalt landscapes in the source areas. These ephemeral rivers commonly cut prominent gorges across basalts, but the lack of basaltic gravels in local sections indicate late timing (late Pleistocene and younger) particularly in the southern parts of the basin (Brown and Gathogo, 2005). In the eastern part of the basin, basalt pebbles and cobbles are common in modern ephemeral streams, but deposits of such streams in local Pliocene and Lower Pleistocene sections are confined to areas near the the basin margin.

Fieldwork at Loiyangalani in combination with studies conducted in many areas of the basin as part of this dissertation reveals common features in cycles of geologic events. Most of these events involved faulting (local angular unconformity) and the development of alluvial fans and channels adjacent to pelagic lacustrine sequences that are locally intruded by basalt flows. The intrusions include the Kankam Basalt (3.2–3.3 Ma), the Lenderit Basalt (2.02 to 2.18 ± 0.02 Ma), and the Balo Basalt (1.79 ± 0.02 Ma) (Gathogo et al., 2008). Exceptions, at least in the Loiyangalani region, include a tectonic event at about 3.5 Ma with no intrusive basalt, and the lack of pelagic lake deposits directly associated with eruption of the Balo Basalt. Three more tectonic events are proposed during the Pleistocene occurring about 1.7 Ma, 1.2 Ma, and 0.3 Ma. The last two tectonic events are poorly characterized due to limitations of the sedimentary record, which is present only in scarce and widely separated outcrops. Overall, the most expansive lake sequences and high-energy alluvial deposits that indicate major changes in depositional environments in the basin developed when tectonic settings allowed (see a summary of events in Fig. 4).

Mammalian Fossil Record

Mammalian taxa from the Omo Group are represented by many tens of thousands of specimens housed at the Ethiopian National Museum in Addis Ababa and in the National Museums of Kenya in Nairobi. Table 3 lists most genera of macromammals (primates, carnivores, proboscideans, perissodactyls, and artiodactyls) that have been retrieved from the Omo Group. Information on most of these specimens is recorded on digital catalogues accessed for this dissertation. Field studies and data analysis reveal significant patterns relating the outcrop geology to fossil occurrences. The fossil abundance and taxonomic representation appear to be strongly influenced by local as well as basin-scale geology.

Data used in this dissertation mostly include published records (Leakey and Leakey, 1978; Harris, 1983; Harris et al., 1988a; Feibel et al., 1989; Harris, 1991; Wood, 1991; Leakey et al., 1995; Isaac and Isaac, 1997; Bobe and Eck, 2001; Harris and Leakey, 2003; Leakey and Harris, 2003; Jablonski and Leakey, 2008). Data from the more recent unpublished collections were examined but are not included in estimates reported below. However, general patterns revealed by the recent findings are likely to be comparable to the observations reported previously, since fieldwork for this dissertation involved geologic placement of many new (upbublished) fossil finds. Unpublished collections include specimens from the eastern and western parts of the basin where M. G. Leakey and L. N. Leakey have conducted field expeditions in collaboration with the National

Museums of Kenya, the Kalakol Research Project, the Koobi Fora Research Project, the West Turkana Paleontological Project, and the Turkana Basin Institute. Leakey (2001) describes trends in some of the more recently found fossils from the Nachukui Formation in her doctoral dissertation on body weight estimation of Bovidae and faunal change in the Pliocene and Pleistocene assemblages. Several thousands of mammalian specimens representing Pliocene and Pleistocene assemblages were retrieved that are associated with hominin specimens from the Lomekwi area of West Turkana and Ileret area of East Turkana, respectively (Leakey et al., 2001; Spoor et al., 2007).

The abudance of mammalian fossils in many parts of the study area is strongly associated with particular depositional settings. Fossils from outcrops in the southern parts of the basin at Kanapoi, Lothagam, and South Turkwel represent at most 2% of catalogued mammalian fossil specimens. Exposures at these locations are mostly Lower and Middle Pliocene strata, which have produced the oldest fossil assemblages in the Omo Group sequence. Upper Pliocene to Middle Pleistocene outcrops are sparse and poorly studied. Mammalian fossils are rarely associated with these deposits. Mammalian fossils are most commonly associated with fluvial deposits. A majority of the mammalian fossil specimens (at least 70% of total) of the Omo Group sequence come from the Lower Omo Valley, primarily the Shungura Formation. This is expected since the area has the thickest exposures of the sequence, with extensive outcrops of fluvial deposits. Depositional settings that represent alternating fluvial and marginal lacustrine beds in the Omo Group are also rich in mammalian fossils. Outcrops in East Turkana (Koobi Fora Formation) that are dominated by alternating fluvial and marginal lacustrine beds have yielded a majority of mammalian fossil specimens (at least 15% of total).

Most fossils from East Turkana come from extensive outcrops of the KBS Member that dominate the Koobi Fora region, and in which fluvial beds interfinger with marginal lacustrine deposits. The West Turkana region (Nachukui Formation) has yielded significantly less mammal fossil specimens (about 5% of total), which mostly come from scattered outcrops. In contrast with East Turkana, the West Turkana outcrops are characterized by high-energy beach deposits with thick mollusk-packed sandstones adjacent to pelagic lake deposits.

Deposits of the Lonyumun Member (ca. 4.3–3.97 Ma) are widely exposed and record one of the thickest intervals (about 110 m) of the Omo Group, but have yielded relatively few mammalian fossils because they generally represent lacustrine facies. Published fossils from this time intervals are associated with fluvial to deltaic deposits that are exposed in only a few locations in the basin. Most fossils come from fluvial deposits that underlie the main lacustrine sequence (about 4 Ma) and that form isolated outcrops in the northern (Mursi) and southern (Kanapoi, Lothagam), and western (Kosia) parts of the basin. These fluvial deposits record depositional settings of local rivers before the perennial axial system of the Omo River became established in the basin. Only a few mammalian fossils have been discovered in the Lonyumun Member time sequence above the lake. The best known assemblage from these younger deposits comes from the base of the perennial axial fluvial system of the Omo River at a location in the Allia Bay region, East Turkana. Hominin fossils include *A. anamensis*, which is represented in both the lower and upper fluvial sequences at Kanapoi and Allia Bay, respectively.

The local depositional settings that are represented by outcrops at various parts of the basin significantly dictate the taxomic representation of the mammalian fossil data. For example, Harris et al. (1988a) describe pronounced geographic trends from east to west in fossil assemblages from the Pliocene sequence in the Lomekwi region of West Turkana. They observe that the trends are predictable based on paleogeographic settings. For example, with bovids that prefer woodland and forest edge (tragelaphines and impala) being dominant in densely vegetated ranges along the west; bovids that prevail in open grassland and scrub (alcelaphines and antilopines) being particularly common farther east near lake margins or along floodplains of the Omo River; and bovids that favor wet grassland (reduncines) dominating east-draining rivers that may have been more permanent in the past than they are at present. The geographic distribution seen in the Lomekwi region fossil assemblages is compatible with temporal change in depositional settings; a gallery forest of the Omo River dominated the western part of the area during the early part of Lomekwi Member time.

Climatic Controls

The dynamics of modern climate in eastern Africa are well described (e.g., Sepulchre et al., 2006; Nicholson, 1996). Various proxies have been used to reconstruct Pliocene and Pleistocene climatic conditions in this region (deMenocal, 1995; Vrba et al., 1995; Maslin et al., 2014). The main proxies that are utilized include carbon isotopes in carbonate nodules from fossil soils (Nicholson, 1996; Levin et al., 2011) of the Omo Group, results from which compare with results from oxygen isotopes of marine invertebrates (Wynn, 2004) and terrigenous dust from Africa deposited in the Arabian

Sea (deMenocal, 1995). The main perennial fluvial systems draining into the Omo-Turkana Basin (Omo, Turkwel, Kerio) and the Mediterranean Sea (Blue Nile, White Nile) have had a common drainage area in East Africa at least since the Early Pliocene. Both areas are therefore expected to record similar climatic signals with respect to fluvial inputs (water and/or sediments). For example, sapropelic organic layers in the Mediterranean Sea indicate intensified monsoonal rainfall in the catchment area of the Nile River (Rossignol-Strick, 1985; Lourens et al., 1996; Emeis et al., 2000; Rohling et al., 2002), which has its main sources in the Ethiopian highlands (Blue Nile) and Lake Victoria (White Nile). Sapropel layers in the Mediterranean result from increased freshwater from the Nile during periods of increased precipitation in eastern Africa, whereas intervening deposits composed of high amounts of eolian dust are linked to the Sahara during dry periods (Emeis et al., 2000). These wet-to-dry patterns also correspond to expected changes in oxygen isotope and sea surface temperatures calculated from foraminiferal assemblages in the Mediterranean Sea and the open ocean (Lourens et al., 1996). Results from most eastern African paleoclimate studies (cited above) agree on the broad changes with long-term trends such as progression towards drier and cooler conditions from the Early Pliocene to the Late Pleistocene. McDougall et al. (2008) and Brown et al. (2012) have shown that there is good correspondence of the sapropel record with that of deposition of the Kibish Formation for at least the past 200 ka.

Detailed geochronological work on the Omo Group sequence including identification of volcanic ash layers in deep-sea cores (Sarna-Wojcicki et al., 1985; Brown et al., 1992; deMenocal and Brown, 1999; Feakins et al., 2007) enables correlation between local depositional settings and global climatic changes including Milankovitch cycles as recorded in the marine rocks. Despite the excellent chronostratigraphic record in the Omo Group sequence, some disputes exist about the interpretation of paleoclimatic data with respect to the east African Pliocene and Pleistocene sequences and global climate change. For example, Trauth et al. (2008) state that large-amplitude variations in tropical Africa climate are unrelated to the onset and amplification of high-latitude glacial cycles as previously suggested by deMenocal (1995, 2004). Trauth et al. (2008) and deMenocal are in favor of the 400 kyr eccentricity maxima as periods when lake levels in East Africa oscillated with the largest amplitude at precessional/half-precessional cycles, citing related studies (Trauth et al., 2005, 2007). However, results presented in this dissertation support deMenocal (1995, 2004) by demonstrating that most of the prominent lake sequences in the Omo Group sequence coincide with tectonic events in the basin (e.g., the Lonyumun Lake of about 4.0 Ma, the Lokochot Lake of about 3.5 Ma, Waru Lake of about 3.2 Ma, the Lorenyang Lake of about 2.0 Ma, and the Nachukui Lake of about 1.3 Ma).

This dissertation demonstrates that climate-related lake fluctuations in the basin generally involved minor lakes with thin mollusk-packed sands alternating with fluvial mudstones possibly at precessional scales (Brown, 1995; Lepre et al., 2007; Brown and Fuller, 2008). Fluvial cycles deposited by the Omo River that range in thickness from 5 to 10 m may be additional indications of precessional-scale orbital climatic forcing. In some well-dated continuous sequences in the Shungura Formation, for example, the average time needed to produce a fining-upward sequence is about ~20 kyr (de Heinzelin, 1983), which closely approximates the average duration of precessional cycles. Figure 4 illustrates that major perennial rivers (ancient Omo River) transported sediments to many parts of the basin without significant restriction by climatic conditions. These findings indicate that the impact of climate forcing and cycles on depositional settings of the Omo-Turkana Basin was locally muted or amplified by geographic factors such as physiography. Note also that the temperature of the region has not changed appreciably over the past 4 Ma (Passey et al., 2010).

Climate change has become the main focus of many studies on the evolution of modern mammals, particularly humans (Sepulchre et al., 2006). It is also commonly assumed that climate was the main cause of pronounced shifts in depositional settings within the study area and the surrounding regions where it caused concurrent development of thick lake sequences (Trauth et al., 2005, 2007, 2009; Maslin et al., 2014). Many studies (e.g., Behrensmeyer et al., 1997; Bobe et al., 2002; deMenocal, 2004; Hernández Fernández and Vrba, 2006) have focused on interpreting temporal trends in the fossil assemblages of the Omo-Turkana Basin. Such trends may reflect large-scale changes due to climate, but they are also strongly affected by local depositional settings that occurred due to tectonic activity (this dissertation) and the preservational or taphonomic bias of some depositional settings (Alemseged, 2003; Gathogo, 2003). Detailed understanding of local geology in the context of regional tectonic settings and global climate changes is critical in this type of analysis as described below.

Geographic Aspects

Geographic aspects such as location and physiography significantly affect modern depositional environments in the Omo-Turkana Basin by controling hydrological features, flora, and fauna. Figure 15 shows the general physiography of the African continent, and Figure 16 highlights physiographic features in the large drainage area of the Omo-Turkana Basin, which spans 9° of latitude. The main hydrologic features in the basin include one of the world's largest desert lakes (Lake Turkana) and a perennial river system (Omo River). The lake and river primarily exist because of rift-related physiography, tectonic depression, and volcanic highlands, respectively. Global climatic conditions play a secondary role in supporting the lake and river. When combined with the local physiography and global climate, hydrologic features are the key determining factors of the type and distribution of vegetation and wildlife in many parts of eastern Africa. The highest species diversity and endemism among living mammals and birds occurs in eastern Africa (Groombridge and Jenkins, 2002). It is not known how long the modern east African faunal phenomenon has existed or whether it extends in the fossil record as far back as the Early Pliocene. Lack of fossils in many parts of Africa, particularly in the central and western regions, makes it difficult to verify any postulated distribution of taxa. Nevertheless, local and regional geologic records enable reasonable reconstructions of habitat-related factors such as physiography and hydrology, which are known to be uniquely associated with the modern eastern African flora and fauna. Results presented in this dissertation provide a geologic framework for local and regional reconstruction of paleogeographic features, including drainage and general physiography related to faulting and basalt volcanism.

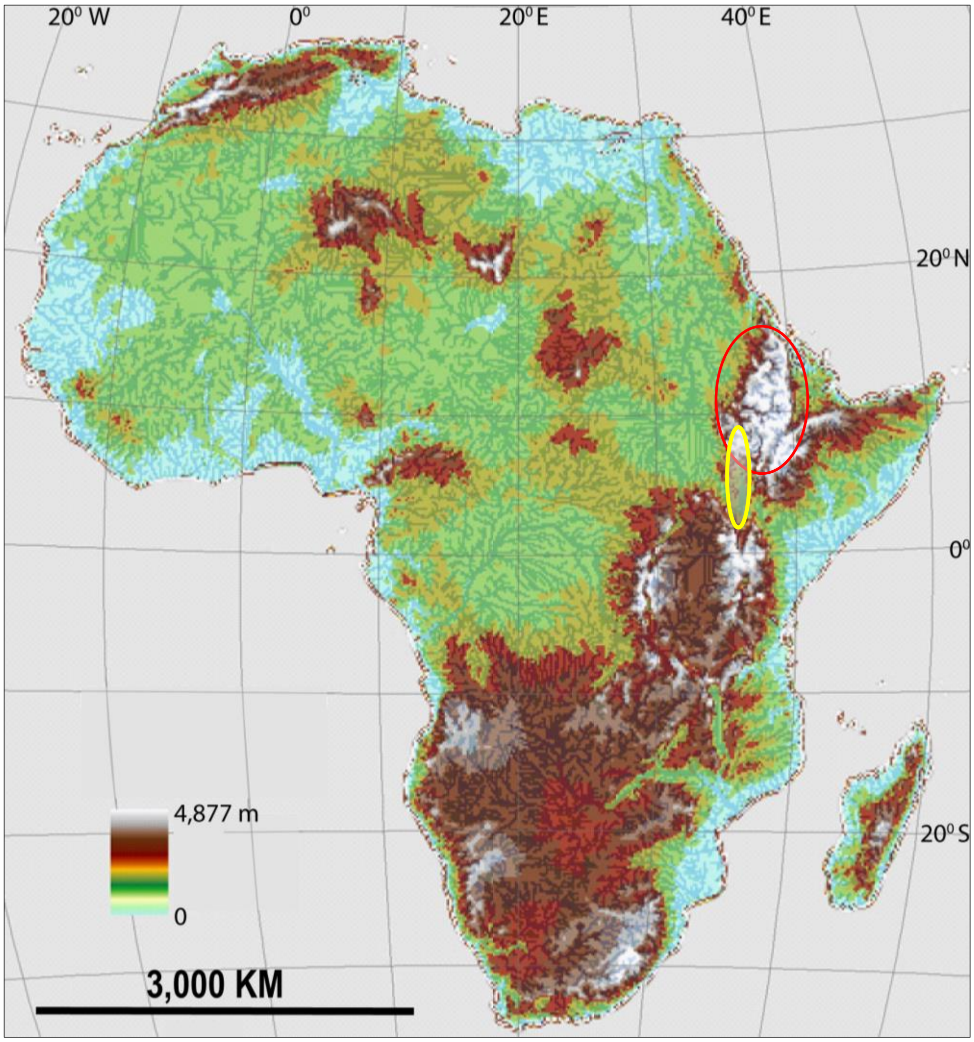

Figure 15. Digital elevation map of Africa showing general physiographic features. The eastern African region is characterized by high topography associated with tectonic uplift and volcanic highlands that are mostly associated with activity along the East African Rift System. The volcanic rocks include thick and extensive Paleogene and Neogene basalt flows forming the Ethiopian Highlands (red oval) north of the drainage area of the Omo-Turkana Basin (yellow oval).

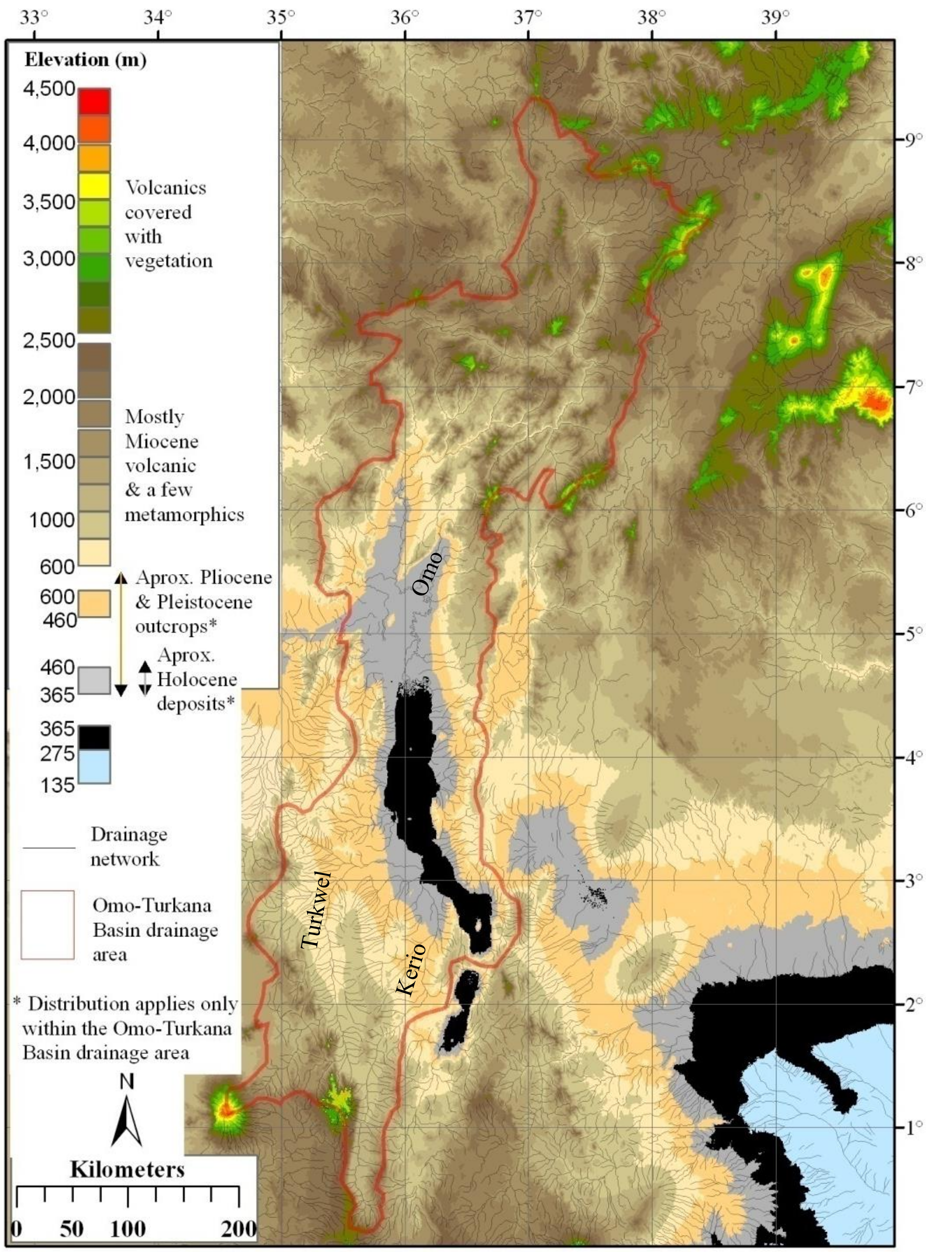

Figure 16. Generalized elevation map of the Omo-Turkana Basin illustrating general physiography. High terrain (greater than 600 m ASL) is mostly defined by Miocene to Pliocene volcanic (primary basaltic) and metamorphic rocks that have been faulted and uplifted along the eastern and western margins of the Omo-Group depositional area. The main drainage systems are Omo, Turkwel, and Kerio Rivers, with the former being perennial and supplying the bulk of the sediments the main depocenter(s).

Various studies (Rapoport, 1975, 1982; Groombridge and Jenkins, 2002; Wynn, 2004; Lisiecki and Raymo, 2005;) have attributed the modern east African faunal phenomenon primarily to the coupled effect of two geographic controls: i) proximity to the equator and ii) physiographic features related to rift escarpments and volcanic highlands as discussed below. Proximity to the equator is generally related to increases in much of the world's biodiversity in terrestrial (Groombridge and Jenkins, 2002) and marine (Stevens, 1989) environments. In marine environments, for example, Jablonski et al. (1986) suggest an "out of the tropics" model in which the tropics are both a "cradle and a museum," with taxa preferentially originating in the tropics and expanding over time into higher latitudes without losing their initial tropical distributions. One possible mechanism involves Rapoport's rule (or effect) that describes how species diversity generally increases as favorable geographical ranges narrow towards the tropics (Rapoport, 1975, 1982; Groombridge and Jenkins, 2002). The Rapoport effect and biomic specialization in African mammals are reported by Hernández Fernández and Vrba (2005). The effect of physiography is discussed by Vrba (1992), who notes that in mammalian evolution, topographically diverse areas are more likely to be sources of species diversity because there is constant fragmentation of habitats and distributions. No studies have investigated whether the superior species diversity and endemism in modern mammals seen in eastern Africa also characterized the Pliocene and Pleistocene fossil record in the Omo-Turkana Basin. Lack of a fossil record in most regions of Africa makes outcrop-based geologic interpretations in the study area critical.

Modern Lake Turkana and the perennial Omo River system have their equivalents in the Pliocene and Pleistocene sequence of the study area where they alternately dominated the depositional settings. Observations suggest that, at its origin in the Early Pliocene, the Omo-Turkana depositional basin tapped major drainages from an already existing physiography that characterizes the larger Omo-Turkana drainage area, which extends from near the equator (source of the Kerio and Turkwel Rivers) to about latitude 9° N (source of the Omo River). The geologic record indicates that the three main drainage systems in the basin--the Omo, Kerio, and Turkwel Rivers--could be as old as the Miocene (Powers, 1980; Leakey and Harris, 2003). However, it was not until the early Pliocene that a significant accumulation of water and sediment from these drainages began in a coherent manner (in the Omo-Turkana Basin) forming the Omo Group. Although inception of deposition within the basin is not the focus of this dissertation, it is almost certain that downwarping of the present area of Lake Turkana occurred ~4.5 Ma ago (Bruhn et al., 2010). The resulting increased gradient perhaps resulted in capture of the Omo River, and possibly also the Turkwel River, which may formerly have been part of the Nile drainage system (see Fig. 16).

High topography associated with the source of the Omo River in the Ethiopian highlands dates back to the Miocene (Hoffman et al., 1997; Kieffer et al., 2004; Gani et al., 2007), which suggests that the river may have remained perennial during dry climatic conditions at least since the Early Pliocene. Thus it is not surprising that geological observations show that fluvial deposits of the proto-Omo River dominate the Omo Group sequence in most parts of the basin since that time. The Omo River system drains an area that is equivalent to 58% of the basin's catchment, and it supplies at least 90% of the water in Lake Turkana (Gani et al., 2007). The source of the river system is the Ethiopian Highlands, which have relatively dense vegetation supported by cool conditions with reduced surface evaporation and transpiration. These conditions result in excess groundwater storage at shallow depths following precipitation. The groundwater subsequently translates to effluent flow into the headwaters of the Omo River. As a result, the river maintains a perennial base flow to the basin throughout the year even during times of reduced rainfall when surface flow stops in all other major drainages including the Kerio River and the Turkwel River.

The Turkwel and Kerio Rivers are the main ephemeral fluvial systems in the basin and originate from volcanic or metamorphic highlands slightly north of the equator and east of Mt. Elgon. These rivers maintain limited perennial flow near their sources, but at least the Kerio River maintained a perennial flow farther north during Miocene. For example, the Miocene sequence at Lothagam shows evidence of perennial flow from the proto-Kerio River with *E. elliptica*, but the overlying Pliocene and Pleistocene sequences suggest that neither the Kerio nor the Turkwel Rivers were perennial (Powers, 1980; Leakey and Harris, 2003). Fluvial deposits attributed to these rivers intercalate with proto-Omo River deposits at least in the southern parts of the basin at Lothagam and South Turkwel (Powers, 1980; Leakey and Harris, 2003).

Changes in physiography resulting from the development of the East African Rift Valley including tectonic uplift and associated volcanic highlands may have produced local effects similar to those resulting from global change in climate, particularly when considering precipitation and associated vegetation. For example, Sepulchre et al. (2006) in their eastern Africa climate model demostrated increased preciptation by 15% and 30% after reducing the elevation of the Ethiopian highlands to 2,000 m and 400 m respectively. The highlands range in elevation from about 1,500 m to slightly above

4,500 m (Fig. 16). There is evidence of significant uplift of the Ethiopian highlands during the Pliocene (Gani et al., 2007). Pik (2008) place uplift of the mountains separating the Omo Turkana Basin from the Chew Bahir Basin in the middle Miocene, about 20 Ma ago. The changes in precipitation will significantly affect local hydrologic systems such as rivers and lakes as well as the type, abundance, and distribution of vegetation and associated mammals.

Rift Tectonics

Many studies acknowledge the potential for tectonic activities along the East African Rift System to shape Pliocene and Pleistocene depositional settings in various parts of the Omo-Turkana Basin, including the Lower Omo Valley (Brown and de Heinzelin, 1983), West Turkana (Patterson et al., 1970; Harris et al., 1988b), and East Turkana (Vondra and Bowen, 1976; Cerling, 1977; Brown and Feibel, 1986). However, as Feibel (2011) notes, recognition of structural bounds and active subbasins from specific episodes of the past is often complicated. Recent studies (Haileab et al., 2004; Gathogo and Brown, 2006; Gathogo et al., 2008), including results presented in this dissertation, focus on the timing and relative magnitude of syndepositional tectonic activities such as faulting, basalt volcanism, and associated changes in topography, drainage systems, and depositional settings as summarized below.

Evidence from outcrop geology is used as a proxy for subsurface structural features that have been described by various workers (Vondra and Bowen, 1976; Dunkleman et al., 1988; Morley et al., 1992, 1999) largely based on geophysical methods. For example, seismic data in combination with the identification of the 4 Ma old basalts of the Gombe

Group (Morley et al., 1992; Haileab et al., 2004) across the basin indicate that the Omo Group sequence has subsided by up to 4,000 m under Lake Turkana since the Early Pliocene (Fig. 17). Even more recently, a gravity survey in the lower Omo Valley has shown that similar thicknesses (up to 5 km) of sedimentary strata are located beneath the Omo River, immediately east of outcrops of the Shungura Formation (Mammo, 2012). The temporal and spatial magnitudes of tectonic subsidence in the study area can be closely estimated by observing stratal thicknesses and changes in depositional settings between dated horizons as discussed below.

Previous work and new data presented here indicate that the rate of tectonic subsidence in the Omo-Turkana Basin has been highly variable in different parts of the basin and through time since the Early Pliocene. General rules about relative subsidence rates in relation to the basin center and margins can be misleading without defining geographic limits (location of major faults) and time intervals (timing of major tectonic activity). That said, thickness in exposed sections of the Omo-Group sequence generally increases towards the basin center, which is defined by Lake Turkana. The Lower Omo Valley, being near the central axis, records the thickest sequence (at least 766 m), but even this package of strata may be thin compared to that identified immediately east of the outcrops (Mammo, 2012). Reasonable estimates can be made about total subsidence or vertical displacements along major faults in various parts of the basin based on composite thicknesses of exposed strata. For example, in the Shungura region, the vertical motion of faults observed in outcrops of the Omo Group sequence is at least 500 m, and may be up to twice as much (Brown and de Heinzelin, 1983).

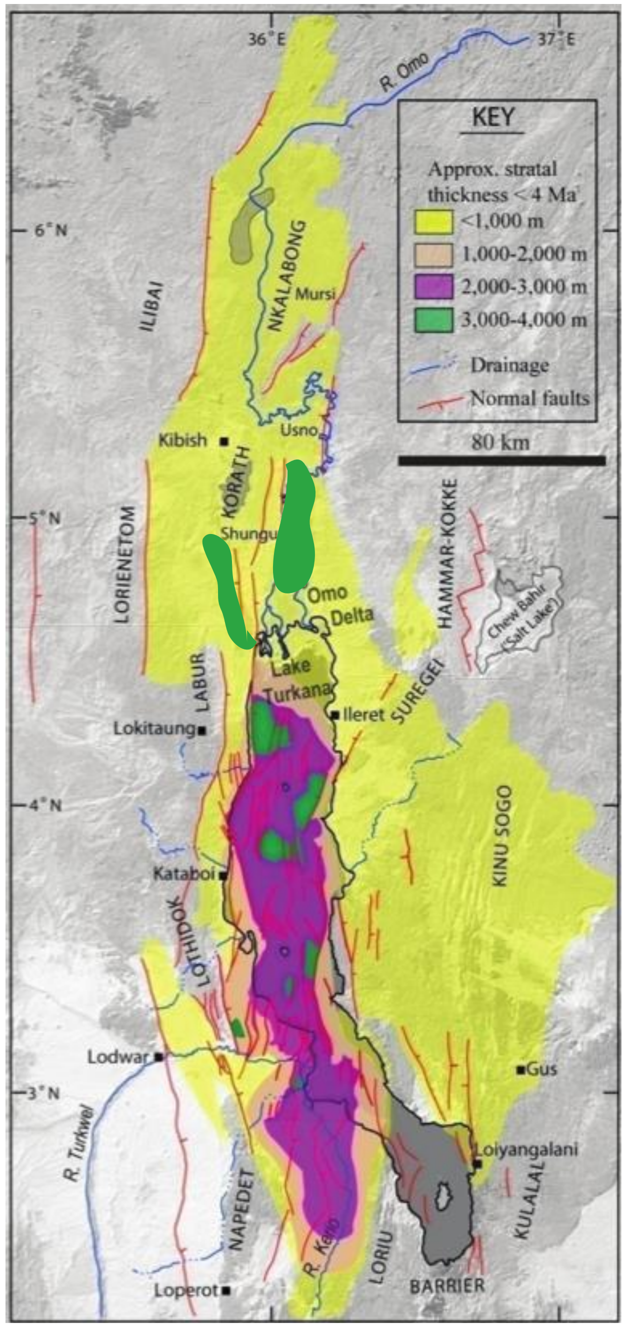

Figure 17. A map showing major structural features (hills and ranges labeled in all upper case), prominent faults, and stratal thicknesses of the Pliocene and Pleistocene sequence that reflect tectonic subsidence. The base of the Omo Group is chosen at the depth of the "Orange Horizon," which correlates with the Gombe Group Basalts (ca 4.0 Ma; Haileab et al., 2004), Mapping is based on seismic reflection data (Morley et al., 1999).

Faults with vertical displacement of up to about 490 m occur in the Ileret area (Gathogo, 2003; Gathogo and Brown, 2006), which is consistent with the stratal composite thickness (about 560 m) between Koobi Fora and Allia Bay regions. The magnitude and timing of syndepositional faulting in association with the Omo-Group are briefly described in the Lower Omo Valley (de Heinzelin, 1983), West Turkana (Harris et al., 1988a), East Turkana (Gathogo, 2003), and Loiyangalani (Gathogo et al., 2008). A comprehensive analysis that includes local and regional implications of depositional settings is detailed above in Chapter 2 and summarized below.

The main rift-related structures that were variably active in the study area between early Pliocene and Recent include longitudinal faults in the Lower Omo Valley (Brown and de Heinzelin, 1983; Mammo, 2012), West Turkana (Harris et al., 1988a), East Turkana (Findlater, 1976; Gathogo, 2003), southwestern Turkana (Powers, 1980; Leakey and Harris, 2003), and the southeastern region north of Loiyangalani (Gathogo et al., 2008). Some of these faults define prominent topographic features where metamorphic rocks form the footwall along scarps, block-faulted mountains, and prominent ranges. Examples of such topographic features include the Hammar Highlands, Kacheriangor Range, Lokwanamoru Range, Lorienetom Range, Maji Escarpment, Maji Highlands, and Nkalabong Range along the Lower Omo Valley, and the Labur Hills in the northern part of West Turkana (Brown and de Heinzelin, 1983). Other major longitudinal rift structures associated with metamorphic rocks as footwalls but without significantly topographic expression include the Kosia-Lokichar Fault that defines the western margin of the basin in West Turkana and the Balo Fault (Gathogo et al., 2008) that defines the local western margin of the Pliocene sequence in the Loiyangalani region.

Overall, many major tectonic episodes are poorly manifested near the axial center of the basin including the Lower Omo Valley in the Shungura region. The record of the most stable and continuous sedimentation exposed of the Omo Group occurs in the northern part of the basin at Shungura (de Heinzelin, 1983). Disconformity surfaces representing long durations (hundreds of thousand years) due to a break in sedimentation or erosion are rare in the Shungura Formation (Brown and de Heinzelin, 1983). A correlative sequence in the Ileret Area, for example, exhibits unconformities that represent up to ~2 Ma along an east-west transect in Pliocene deposits (Gathogo and Brown, 2006), although Kidney (2012) was able to show that one of these is a composite, with hiatuses of 1.3 and 0.7 Ma on two surfaces that merge toward the basin margin. Nevertheless, tectonic effects are still recognizable in the Lower Omo Valley. For example, Brown and de Heinzelin (1988) identify systematic disparities in local stratal thicknesses in the northern part of the basin in the Lower Omo Valley. In their Figure 4, they compare the thicknesses of selected intervals or members of the Omo Group among different locations including the Shungura Type area, Kalam, and Usno. Their methodology involves not only thickness differences of the same members in different areas, but also a comparison in the thicknesses ratio among different members from one area to another. The study infers a syndepositional subsiding trough situated along the present axis of Lake Turkana. With the recent work of Mammo (2012), it appears that this is only partly correct, and that the Shungura Formation was deposited on a horst that separated two basins, one located east of the fault that bounds the Shungura Formation on the east, and the other located west of the fault that bounds the Shungura Formation on the west.

CHAPTER 4

SUMMARY

The potential influences of global climate and local geological features on the fossil record of the Omo-Turkana Basin are extensively described in the literature, particularly in reference to habitat evolution (Hernández Fernández and Vrba, 2005, 2006; Spoor et al., 2007). Many studies cite rift tectonics as being key to understanding the geology of eastern Africa (Haesaerts et al., 1983; Morley et al., 1992; Morley et al., 1999; Hernández Fernández and Vrba, 2005; Bennett et al., 2009), but the contributions of rift tectonics in shaping the ancient habitats of ancestral mammals including hominins are poorly understood (Maslin and Christensen, 2007). The tectonic history of the Omo-Turkana Basin during the Pliocene and Pleistocene Epochs is poorly understood in terms of the timing and magnitude of structural movements, but now this dissertation reports (Chapter 2) the main tectonic events that are recorded in this well-dated and extensive sedimentary sequence. This combination of evidence makes the basin an ideal area to study the tectonic evolution of part of the East African Rift System (Chapter 3). Findings from this dissertation consist of possible associations among depositional environments, tectonic movements, climatic influence, and distribution of mammalian fossils in the Omo-Turkana Basin during the Pliocene and Pleistocene Epochs as summarized below.

Rift Tectonics and Deposition

Data and field evidence presented in this dissertation indicate that deposition of the Pliocene and Pleistocene Omo Group sequence (~800 thick in exposed sections) in the Omo-Turkana Basin has been modulated primarily by tectonic activities. The Omo River and Lake Turkana, and their ancient equivalents, have been the main hydrologic features accounting for deposition of most of the sequence. The river represents an axial perennial system that drains the majority of the basin’s catchment area and supplies at least 90% of the water in the lake. The inception of major lake sequences in the basin has been determined primarily by the availability of accommodation space generated by tectonic movements. Based on new data presented in this dissertation (Chapter 2), the most prominent lake sequences in basin coincide with major tectonic events. Examples include the Lonyumun Lake (~4.0 Ma), the Lokochot Lake (~3.5 Ma), the Waru Lake (~3.2 Ma), the Lorenyang Lake (~2.0 Ma), and the Nachukui Lake (~1.3 Ma). Rift-related basalt eruptions are also associated with the major Pliocene lakes, except for the Lokochot Lake and the Nachukui Lake. Ancient lakes in the basin have ranged from freshwater to alkaline, depending on whether or not the basin had an outlet, commonly to the Indian Ocean. Accumulation of sediments in the basin occurred at many locations within 40 km of the longitudinal axis of the basin, but the depositional area may have extended farther east (where Gombe Group basalts are exposed) in the early Pliocene. Ephemeral rivers developed from the eastern and western margins of the basin as early as the middle Pliocene and commonly influenced local depositional settings, particularly during pronounced tectonic episodes. The occurrence and local dominance of these ephemeral systems act as indicators of tectonic movements within or at the margins of the basin, and they became increasingly common features in the local geologic record in many parts of the basin beginning in the Pleistocene. There has been a general reduction in the depositional area towards the north and axial center. The end of Omo Group deposition appears to be associated with a last major tectonic episode in Late Pleistocene (at least about 0.3 Ma).

Climate and Depositional Facies

The well-dated Omo Group sequence and paleoclimatic data from the Mediterranean Sea or deep sea provide an excellent opportunity for correlating local to regional depositional settings since early Pliocene. Field evidence provided in this dissertation (Chapter 2) indicates a strong relationship between precessional-scale climatic forcing and the deposition of main fluvial sequences as well as associated marginal lacustrine sequences representing intermittent fluctuations in lake levels. However, none of the widespread perennial lacustrine sequences in the basin coincide with the 400 ka eccentricity maxima as suggested by Trauth et al. (2005, 2007, 2009), among others. These new findings resolve the long-standing disputes about the implication of large-amplitude and small-amplitude climatic variations (deMenocal, 1995, 2004; Trauth et al., 2007, 2009; Maslin et al., 2014) at least in the Pliocene and Pleistocene of the Omo Group sequence in eastern Africa. The independent contributions of climate and tectonics in modifying regional habitats during the last 5 million years when many mammals including humans evolved have been poorly understood. The scarcity of information on tectonic history led many studies to ascribe a climatic origin to changes in depositional settings and associated fossils, for example the origin of major lake sequences and changes in habitats that enabled evolutionary changes of large mammals including humans. This dissertation demonstrates that most prominent changes in depositional environments of the Pliocene to Pleistocene sequence (Omo Group) in the Omo-Turkana Basin (eastern Africa) coincided with rift-related tectonic episodes.

Mammalian Fossil Record and Local Geology

Pleistocene and Pliocene paleontological sites in the Omo-Turkana Basin region are rich in mammalian fossils that include hominids. This fossil record is key to understanding human evolution as many of the earliest appearances and diversification of hominid taxa have examples in the basin. Global climate has been cited as the main factor that defined the evolution of hominids and other African mammals, particularly based on the paleontological and geological record from the Pliocene and Pleistocene sequences in the basin. This dissertation shows how fossil occurrences in the basin correlate locally with geology, and how major changes in depositional environments have determined the distribution of fossil-rich sediments. Also shown is the role of tectonics in masking the influence of global climatic cycles (e.g., eccentricity) on the environments in which ancient mammals including early human lived. Preservational and taphonomic biases related to depositional settings also are reported (Bobe and Eck, 2001; Alemseged, 2003; Gathogo, 2003). Comprehensive interpretation of the fossil record in relation to climatic change requires better understanding of local settings including the effects of tectonic movements on local habitats. Consideration of factors other than climatic changes can have far-reaching implications because the Omo Group has produced a widely referenced fossil record that includes many members of the hominin lineage (or lineages). Results presented here illustrate the importance of isolating climatic signals from those of tectonics for evolutionary studies, at least when dealing with a tectonically active basin. This dissertation also demonstrates that geographic aspects and tectonic-related physiography should be considered when evaluating for the role of climate on the evolution of mammals including humans in eastern Africa and other regions.

CHAPTER 5

CONCLUSION

Findings presented in this dissertation lead to three main conclusions:

i) Rift-related tectonic movements caused the most prominent changes in depositional environments of the Pliocene and Pleistocene Omo Group sequence of eastern Africa. Major tectonic episodes (e.g., ~4.0 Ma, ~3.5 Ma, ~3.2 Ma, ~2.0 Ma, and ~1.3 Ma) characterized principally by extensive faulting, local subsidence associated relative uplift, and basaltic volcanism are associated with widespread lacustrine deposition.

ii) Global climatic forcing and cycles do not coincide with the most pronounced changes in depositional environments of the Omo Group. Orbital forcing or Milankovitch cycles that are recorded in the deep sea, particularly the Mediterranean, generally correspond to precessional-scale fluvial sequences and minor fluctuations in local ephemeral lakes. Eccentricity maxima that occur on a 400 ka scale mostly correspond to perennial fluvial sequences, but not to lacustrine sequences as suggested by Trauth et al. (2005, 2007, 2009), among others.

iii) Field observations and associated interpretations presented in this dissertation suggest that deposition of the Pliocene and Pliocene Omo Group sequence began after major structural movements including downwarping, faulting, and basalt volcanism (e.g., Haileab et al., 2004; Gathogo et al., 2008; Bruhn et al., 2010). Relatively similar tectonic episodes repeated and significantly affected the depositional system dominated by an axial perennial fluvial system corresponding to proto-Omo River and also lacustrine systems (corresponding to proto-Lake Turkana). Studies of subsurface structural features using geophysical modeling and simulations are required in order to understand the mechanism or processes associated with rift tectonics. Integration of such studies with interpretations of outcrop geology will produce a robust numerical model that may find application to many other regions.

Overall the new findings demonstrate that deciding whether influences on deposition are due to tectonic or climatic factors can be difficult to determine without information from outcrop geology and a very robust temporal framework. Therefore, when interpreting the paleontological record, climatic and tectonic signals may need to be weighted differently depending on geographic location and age.

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