MIDDLE PALEOLITHIC HOMININ LAKE ENVIRONMENTS IN SAHARAN NORTH AFRICA

by

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The following individuals read and discussed the thesis submitted by student Cynthia Anne Bradbury, and they evaluated her presentation and response to questions during the final oral examination. They found that the student passed the final oral examination.

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The start of every project brings the hope that this one will lead to the feeling of finally knowing. The end of each project ends brings the realization that Feynman was right: "I was born not knowing and have had only a little time to change that here and there." This project has ended with only that little change leaving more questions than answers. However, even that little change has occurred with significant assistance from a number of individuals.

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ABSTRACT

Contemporaneous with the transition to biologically modern humans was the episodic change from wetter to drier environments in the Egyptian Sahara. At Bir Tarfawi, White Lake sediments represent a wet phase occurring prior to the last interglacial in the now hyperarid Egyptian Western Desert. One hypothesis for the development of Western Desert Pleistocene lakes was that the northward migration of the Intertropical Convergence Zone (ITCZ) provided a path for summer-constrained, Atlantic-sourced precipitation resulting in local precipitation. Oxygen and carbon stable isotope analysis of climate proxies such as the gastropod, *Melanoides tuberculata,* indicate precipitation and groundwater sources as well as the ephemeral or perennial character of the surface water. Unaltered diagenetically, *Melanoides* from a coquina in White Lake sediments had shell carbonate oxygen isotope values from -7.7‰ to 0.8‰. Along shell oxygen isotope values varied from 3‰ to 6‰ with no indication of highly depleted values indicative of Atlantic-sourced monsoon. Also, the shell oxygen isotope values were higher than previously published perennial lake sediment carbonates (-8.3‰ to -5.7‰). The enriched oxygen isotope values for White Lake remnant *Melanoides* suggest the coquina may have formed in shallow waters sensitive to evaporation and multiple precipitation sources, e.g. Indian Ocean, contradicting the hypothesis that the Atlantic monsoon was the sole contributor to the perennial White Lake.

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CHAPTER ONE: INTRODUCTION

The evolutionary transition of biologically modern *Homo sapiens* (BMHS) and movement out of East Africa are largely contemporaneous with the existence of White Lake at Bir Tarfawi. The sediment sequence, along with Middle Paleolithic lithics and invertebrate fossils, dates prior to the last interglacial, which began 130 ka (Basell, 2008; Lahr and Foley, 1998; McDougall et al., 2005; Schild and Wendorf, 1993; Schwarcz and Morawska, 1993; White et al., 2003). In addition to the evolution of BMHS, it is important to examine the types of landscapes that may have facilitated hominin movement (Basell, 2008; Lahr and Foley, 1998; Osborne et al., 2008; Stringer, 2000; Vermeersch, 2001). While no hominin fossils have been found associated with White Lake that could be used to directly evaluate the physical evolution of BMHS, the presence of Middle Paleolithic lithics (Schild and Wendorf, 1993) indicates that the environment associated with White Lake was a hominin resource. Evaluating the habitat of the lake is therefore critical in the discussion of BMHS movement out of Africa.

White Lake, dated to before 130 ka, consists of a sedimentary sequence indicating that sufficient moisture once existed in this now hyperarid region to allow the formation of a lake (Schild and Wendorf, 1993; Schwarcz and Morawska, 1993; Wendorf et al., 1993). The main hypotheses for sources of water in the Egyptian Western Desert include 1) Atlantic-sourced monsoons bring local precipitation and 2) groundwater potentially is recharged farther south in the Sudan. Identifying the source of water along with addressing the question of the seasonal variation of the White Lake are key to understanding the perennial or ephemeral nature of the lake and therefore its availability as a resource for hominins. The gastropod *Melanoides tuberculata* present in the White Lake sediments is a robust biological proxy for isotopic composition of the lake water. Evaluations of the oxygen and carbon stable isotope values of the gastropod shell are used in this study to examine potential source water and seasonal conditions in the lake (Abell, 1985; Li and Ku, 1997; McKenzie, 1993).

Investigating environmental conditions associated with the presence of White Lake, including the seasonality of the lake, provides hints about the way the region was used by hominins. One view suggests hominin movement was coincident with migrating herbivores across open grassland (Basell, 2008; Finlayson, 2005). Another view suggests the Pleistocene hominin preference was associated with a sparsely treed savannah and that movement would occur when environmental conditions supported the preferred environment (Basell, 2008).

Biologically modern *Homo sapiens* use of Egypt's Western Desert during the Holocene wet phase provides an example of hominin behavior in an area subject to rapid climate change and therefore to ecological variation (Bubenzer and Reimer, 2007; Kuper and Kröpelin, 2006; Nicoll, 2004). Holocene wet phase information from Egypt is potentially relevant to the context of White Lake since the climate and environmental changes may be connected to similar physical processes that produced the White Lake wet phase. Traditionally, lifeway patterns associated with Holocene hunter-gatherers have been used to develop models for Paleolithic hominin ecology (Marlowe, 2004). There is artifactual evidence that Neolithic hominins incorporated pastoralism into their lifeway during the Holocene (Kuper and Kröpelin, 2006, Wendorf and Schild, 1998).

Another example of land use response is hominin site data from the Pleistocene of East Africa (Basell, 2008). The East African data apply to hominins living at the same time as the White Lake wet phase. Thus, a comparison of Neolithic and Paleolithic hominin behaviors and resources provides some potential insights regarding Middle Paleolithic adaptations in the Western Desert. Regardless of the specific habitat type, evidence from both the Holocene Egyptian Western Desert wet phase and the Pleistocene East African sites implies that water sources are a key to land use by hominins (Basell, 2008; Bubenzer and Reimer, 2007; Kuper and Kröpelin, 2006; Nicoll, 2004). Therefore, a clearer understanding of the regional transitions from wet to arid environments as well as seasonality could provide direct evidence that can be used to model potential hominin resources at the White Lake.

White Lake at Bir Tarfawi represents only one of many wet phases documented in the Western Desert. Changes in the surface availability of water, particularly in the form of lakes and streams in the Western Desert, have been postulated to result from movement of the Intertropical Convergent Zone (ITCZ), and changes in the ITCZ have been correlated with climate transitions associated with northern hemispheric glaciations (Osborne et al., 2008; Rohling et al., 2002). Northward movement of the ITCZ brings western monsoon rains into the Western Desert, supplying not only local precipitation but also replenishing the regional aquifer (Ghoneim and El-Baz, 2007; Gossel et al., 2004; Hoelzmann et al., 2000; Sultan et al., 1997).

The western monsoon movement implies that the Atlantic Ocean is the water source of precipitation during summer months. This moisture has important implications for hominin habitation in the region; if rainfall were restricted to the summer months, then the surface water in the Western Desert may not have been perennial unless the lakes were sustained by groundwater. Evidence of hand-dug wells in many areas of the Western Desert during the Holocene Wet Phase as well as at Bir Sahara during the Middle Paleolithic attests to the ephemeral nature of surface water during this period (Campbell, 1993; Wendorf and

Schild, 1998; Close, 1990). Thus, determining whether the water source for the White Lake was monsoonal precipitation and/or groundwater provides clues to the type of resources available for hominin use during the Pleistocene.

Information concerning the water source and seasonality of lakes can be obtained through oxygen and carbon stable isotope analysis from gastropod shells such as *Melanoides tuberculata* (Abell, 1985; Leng et al., 1999; McKenzie, 1993; Smith et al., 2004). *Melanoides* is a ubiquitous gastropod found in a variety of environments with perennial fresh to slightly saline water sources, including lakes, rivers, and streams (de Kock and Wolmarans, 2009; Pointier et al., 1992). While this gastropod has a variable life span of up to 5 years, two years are more commonly reported, with attained shell lengths of 30 to 36 mm (Pointier et al., 1992).

The shell of *Melanoides* is secreted year round in equilibrium with the surrounding water, preserving the carbon and oxygen stable isotope ratio of the water source (Abell, 1985; McKenzie, 1993; Pointier et al., 1992; Shanahan et al., 2005; Smith et al., 2004). The isotope ratio of source water is changed by interaction among solid, liquid, and gas phases. Water evaporation and condensation alter the oxygen isotope ratio so that the isotopic composition of the precipitation differs from that of its source (Dansgaard, 1964). Additionally, the gastropod shell carbon stable isotope values preserve information on lake productivity (Leng and Marshall, 2004; McKenzie, 1993).

Specimens of *M. tuberculata* from Tarfawi White Lake trench 10/86 sediment layer 3 were selected for analysis in this study. The samples were taken from a shallow water marl layer with dates ranging from 152 to 265 ka and an average of 220 ± 60 ka using U series dating (Schwarcz and Morawska, 1993). Since gastropod shell and sediment carbonates from previous studies were co-located, *Melanoides* from the deep and shallow areas of the Bir

Tarfawi ca. 125 ka remnant were analyzed. Previously analyzed sediment carbonates from trench 14/86 of White Lake and trench 7/74 of southern Bir Tarfawi ca. 125 ka remnant were incorporated into this study to understand the relationship between shallow water gastropods and deep water sediments (McKenzie, 1993).

Carbon and oxygen stable isotope analysis of *M. tuberculata* shells is used to examine White Lake characteristics and ultimately infer details of the habitats likely surrounding the lake and available for use by Middle Paleolithic hominins. One question involves the water source for White Lake and whether the surface water was derived from direct precipitation, runoff, or groundwater. A second question is can stable isotope data help to determine whether this surface water was perennial (available year round) during the wet period. Hypotheses with respect to the White Lake are the following:

H_0 = Stable oxygen isotope compositions from *Melanoides tuberculata* shell carbonates are isotopically consistent with compositions from Source X.

H_a = Stable oxygen isotope compositions from *M. tuberculata* shell carbonates are not isotopically consistent with compositions from Source X.

Source X = Atlantic Ocean monsoon, Indian Ocean monsoon, Mediterranean precipitation, Nubian Aquifer, Nile flooding

H_0= Carbon and oxygen stable isotope compositions are the same along the shell of the animal.

Ha = Carbon and oxygen stable isotope compositions are not the same along the shell of the animal.

The first set of hypotheses provides information concerning the water source of the lake. If oxygen isotope values are concordant with Atlantic monsoonal precipitation, then the lake may have been ephemeral since this precipitation is presently confined to summer months. Differences in expected values of Atlantic precipitation may be associated with additional water sources such as the Indian monsoon or the Nubian Aquifer. The second hypothesis set concerns variation of the stable isotope values along the length of the gastropod shell. Inconsistent isotopic composition implies a potential change in the water source from local precipitation or groundwater or lake conditions such as stable versus regressive during the life of the organism (Abell, 1985; McKenzie, 1993).

These hypotheses can be used to model the environmental context, i.e., habitat and resource availability, for BMHS in northeast Africa. Additionally, there is the possibility that the White Lake *M. tuberculata* assemblage represents a coquina or mass die-off and, therefore, a transition from a humid to a more evaporative state or saline environment initiating a terminal lake phase or a regression within the perennial lake evolution (Hailemichael, 2000).

CHAPTER TWO: BIR TARFAWI AND CLIMATE CHANGE

The interaction of astronomical and earth processes cause variation in insolation prompting the atmospheric and oceanic changes responsible for the movement of the ITCZ. Northward movement of the ITCZ creates a path for Atlantic monsoonal precipitation to penetrate into the Egyptian Western Desert causing an increase in surface moisture. Sedimentary sequences, such as Bir Tarfawi, are evidence of the resulting episodic changes from wetter to dryer environments prior to the last interglacial. Increasing moisture in the Western Desert created conditions suitable for hominin movement into the area.

Bir Tarfawi Description

The Eastern Sahara, a hyperarid region (<20 mm/yr precipitation), spans an area of more than 2 million km^2 and includes the Egyptian Western Desert, Northwest Sudan, and sections of Chad and Libya (Figure 2.1, Kuper and Kröpelin, 2006; Sultan et al., 1997). The Egyptian Western Desert is an extreme environment even within the Sahara. It presently receives less than 1 mm/yr precipitation because it currently is beyond the reach of winter Mediterranean moisture and summer Atlantic monsoons. The Western Desert (Figure 2.1) is a peneplain in the northeast corner of the Selima Sand Sheet (Hill, 2009), bordered on the east by the Libyan plateau, the north by the Dakhleh escarpment, the northwest by the Sand Sea, and the west by Gilf el Kebir massif. To the south, the desert continues well into the Sudan (Wendorf et al., 1993). It is within this plain, interrupted only occasionally by sand dunes and wadis, that the Bir Tarfawi deflational basin is located (Wendorf et al., 1993a; Hill, 2009). Bir Tarfawi is notable in this landscape due to the presence of tamarisk, acacia, and date palms, which survive due to near-surface groundwater (Wendorf et al., 1993a). Home to foxes, mice, lizards, and insects, Bir Tarfawi at present provides a small survival zone in the midst of a barren landscape (Kowalski, 1993; Wendorf et al., 1993a).

Paleoanthropological interest in Bir Tarfawi, currently a desolate region uninhabited by humans, is prompted by the presence of Acheulian and Middle Paleolithic lithics embedded in sedimentary sequences indicating a much different environment from present. These previous environments contained sufficient moisture to produce surface water and thereby resources that could support hominins. The wet phases are distinguishable in the deflational basin as mesa-like areas of indurated sediments (Schild and Wendorf, 1993). Sequences dating from a minimum age of ~ 300 ka associated with Acheulian lithics to ~ 60 ka associated with the Middle Paleolithic have been recognized (Hill, 1993, 2009; Wendorf et al., 1993a). It is the latter of these sequences corresponding with the movement of biologically modern *Homo sapiens* (BMHS) to the Levant that has been more thoroughly studied (Wendorf et al., 1993). The slightly older (220 ka to 150 ka) White Lake sequence, with the earliest Middle Paleolithic lithics in the area, is the subject of the current study intended to understand details of the earlier environment and its connection with Pleistocene hominin ecology.

Saharan Climate Change

North African wet/dry events have been correlated with glaciation cycles of the northern hemisphere. Glacial cycles, through differing extent and duration, consist of a slow build-up of ice sheets and rapid deglaciation followed by an interglacial or mild period (Osborne et al., 2008; Rohling et al., 2002; Lahr and Foley, 1998; Carto et al., 2009). The chronology of the Saharan wet phases suggests a link to the transition from glacial to interglacial conditions. Some models suggest that during early interglacials, precipitation increases effective moisture in the Sahara as the Intertropical Convergence Zone (ITCZ) moves north, bringing Atlantic-sourced monsoons into the region. The Atlantic monsoon, confined to the summer months, not only supplies local precipitation but also replenishes the regional Nubian aquifer (Crombie et al., 1997; Ghoneim and El-Baz, 2007; Gossel et al., 2004; Hoelzmann et al., 2000; McKenzie, 1993; Sultan et al., 1997; Tisserand et al., 2009).

Increasing moisture in Egypt's Western Desert creates conditions for expansion of savanna flora and fauna from the Sahel in the south (Osborne et al., 2008; Rohling et al., 2002). Northern ITCZ excursion and precipitation amounts vary with each cycle. The glacial/interglacial cycle has been correlated with changes in the earth's orbital parameters (Imbrie et al., 1984).

The solar energy that fuels the earth's climate system is not a constant. Rather it is a function of the distance between the earth and the sun, which according to Milankovitch astronomical theory is controlled by three factors: eccentricity, obliquity, and precession (de Noblet et al., 1996; deMenocal and Rind, 1993; Gasse, 2000). The shape of the earth's orbit or eccentricity varies from nearly circular to slightly oval in a 100,000-year cycle. As eccentricity changes, the distance between the earth and sun increases and decreases with respect to perihelion (closest distance to the sun) and aphelion (farthest distance from the sun). All points of a low eccentricity orbit are equidistant while the more oval orbit has a greater difference in the earth-to-sun distance between perihelion and aphelion. The tilt of the earth's axis, obliquity, changes with a periodicity of about 40,000 years. Obliquity controls

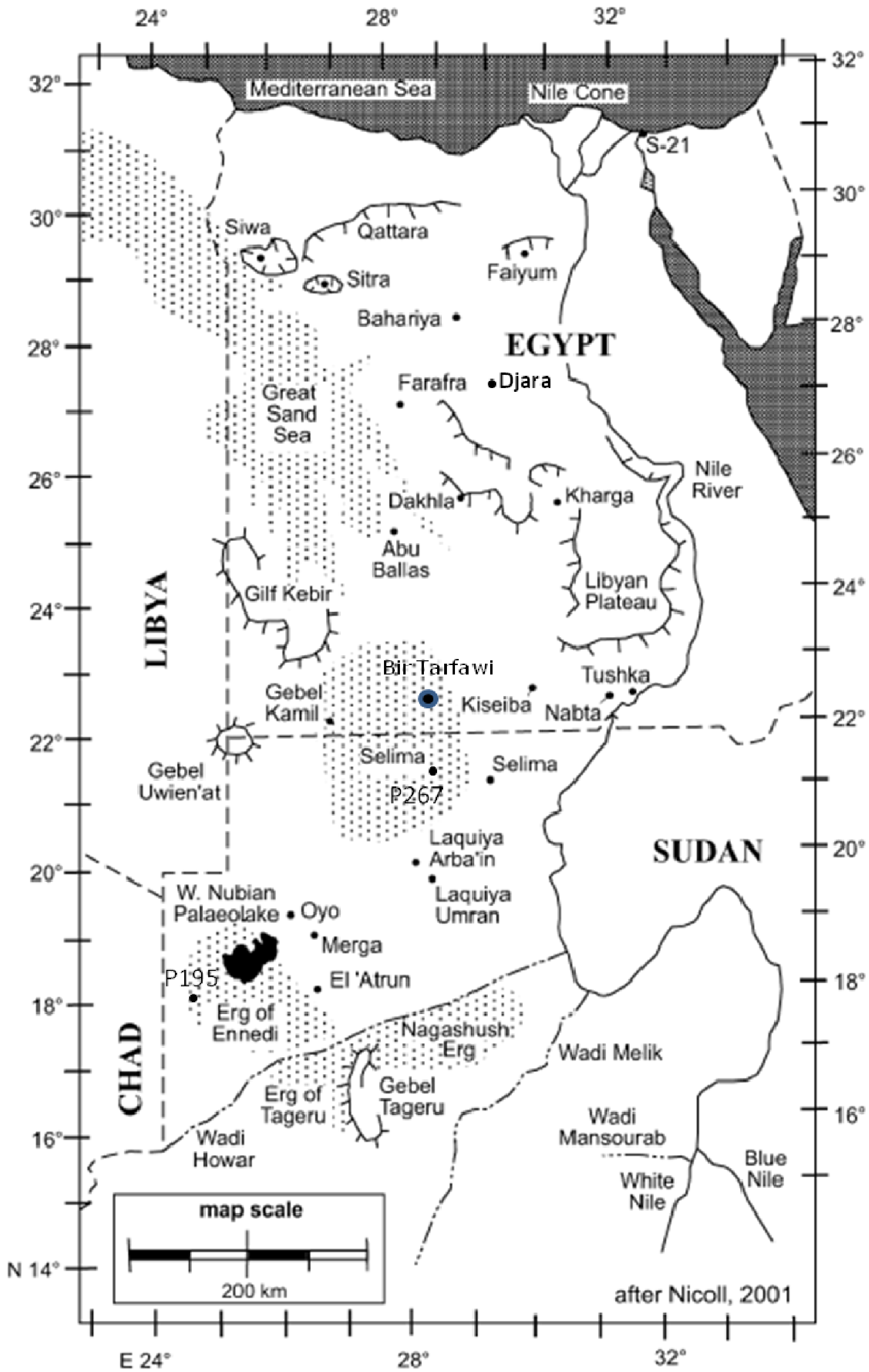

Figure 2.1 Map of eastern Sahara including the Nubian aquifer region of the Egyptian Western Desert, northern Sudan and eastern sections of Chad and Libya (modified from Nicoll, 2004; Szabo et al., 1995; and Wendorf et al., 1993).

the hemispheric differential heating by reducing the northern or southern hemisphere to sun distance. Precession changes the point in the earth's orbit at which the equinoxes occur and can amplify seasonality. A full precessional cycle occurs every 20 ka. Each of these cycles is independent but predictably combines to increase or decrease the amount of solar radiation affecting the atmosphere (de Noblet et al., 1996; deMenocal and Rind, 1993; Gasse, 2000; Hays et al., 1976; Imbrie et al., 1984; Zachos et al., 2001).

Once the sun's energy strikes the atmosphere, a number of chemical and physical processes initiate insolation use and distribution. These processes include interacting with atmospheric, biospheric, cryospheric, geospheric, and oceanic material parameters such as albedo, heat capacity, and roughness, as well as geographic parameters such as elevation and latitude. The results of these interactions create atmospheric circulations such as the Intertropical Convergence Zone (ITCZ) (Figure 2.2). This is the region near the equator where the southern and northern hemisphere tropical easterly winds converge, creating a band of tropical precipitation. With differential heating in each of the hemispheres due to obliquity, the ITCZ migrates slightly north by the summer solstice and slightly south by the winter solstice. With this movement of the ITCZ, associated rains also migrate and are responsible for the seasonality of precipitation throughout most of Africa.

While northern glaciations and Saharan wet phases have been correlated with these orbital parameters, variations between and within each cycle cannot be completely explained by changing insolation (Imbrie et al., 1984; Rossignol-Strick et al., 1998; Rossignol-Strick and Paterne, 1999). Analysis of marine cores off the west African coast suggest the ITCZ moves farther south during the colder parts of the glacial cycle with increased northward movement during the warm interglacial, demonstrating a direct effect of solar insolation on the ITCZ (Sultan et al., 1997; Tisserand et al., 2009).

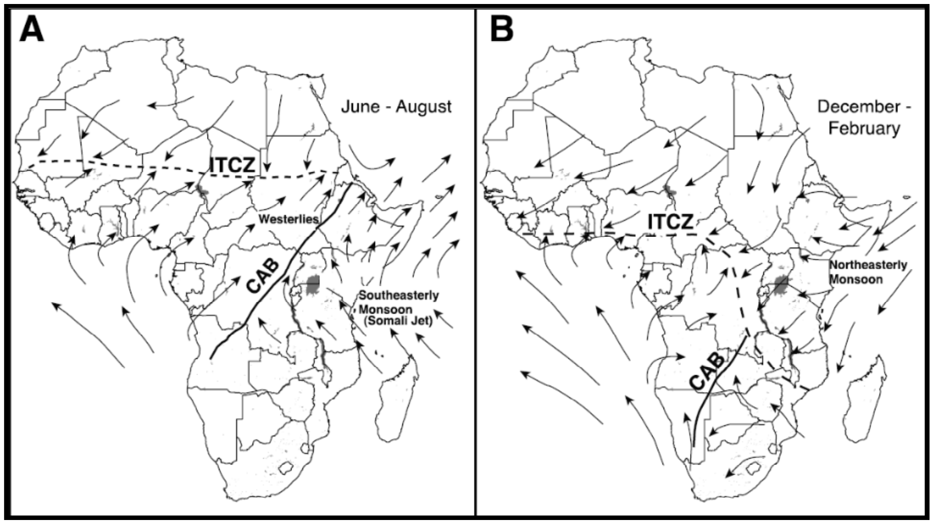

Figure 2.2 Location of the major air boundaries ITCZ (north and south) and Congo Air Boundary (CAB, westerly and easterly) during northern summer (A) and southern summer (B) (Levin et al., 2009).

The movement of the ITCZ creates a series of wet and arid phases, each lasting 1 to 10 ka with rapid century scale transitions between each state (deMenocal, 2008). While there appears to be a strong connection between orbital parameters and wet phase cycles, wet phases do not span the same time intervals as the glacial-interglacial cycles and their abrupt transitions cannot be easily explained by gradual changes in insolation (deMenocal et al., 2000; deMenocal, 2008; Tjallingii et al., 2008). Climate modeling indicates complex interactions in the climate system parameters, especially sea surface temperature and nonlinear vegetation feedbacks, which affect the extent of penetration of the ITCZ and the abrupt return to arid conditions (deMenocal, 2008; de Noblet et al., 1996; Foley et al., 2003; Hoelzmann et al., 2000; Kröpelin et al., 2008; Liu et al., 2009; Tjallingii et al., 2008; Tisserand et al., 2009). While details of the control of the conditions affecting the strength and duration of each African Humid Phase have not been fully identified, these wet phases provide brief changes in the hyperarid conditions generally found in the Sahara (deMenocal et al., 2000; deMenocal, 2008).

The strong zonal nature of the Atlantic monsoon means local precipitation is confined to a narrow belt defining the tropical, Sahelian and Saharan vegetation zones (Figure 2.3 and Gasse, 2000). Thus, as the ITCZ moves to the north creating the northern wet phases, there is only a narrow band of precipitation. This implies that surface water can only form in this narrow precipitation band if water can be delivered to the area by other means. Two such water delivery systems include runoff and groundwater.

Figure 2.3 African map of major environmental regions: Sahara (tan), Sahel (orange), rain forest (green), southern desert (grey) (CSELS, 2011)

The Gilf Kebir rises to the west of Bir Tarfawi (Figure 2.1). Depending upon the northern extent of the ITCZ and the exact track of the Atlantic monsoon, this highland area can intercept precipitation before it reaches the basin. Remote Sensing Images indicate that paleodrainage channels flowed from Gilf Kebir toward Bir Tarfawi sometime in the past (Figure 2.4, Ghoneim and El-Baz, 2007). Paleochannels in the region may have been active during the Tertiary, which would significantly predate the period of interest in this study

(Maxwell et al., 2010; McHugh et al., 1988). The features documented by Ghoneim and El-Baz et al., (2007) mimic present-day topography and are attributed to the Quaternary.

Bir Tarfawi is located within the confines of the Nubian aquifer, which extends over northern Sudan, eastern Libya, and western Egypt (Sultan et al., 1997). Nubian aquifer groundwater flows generally from southwest to northeast and is contained within four sandstone horizons partially constrained by shale layers and granite basement (Patterson et al., 2005; Robinson et al., 2007; Sultan et al., 1997). Deep horizons with long residence times are recharged during wet phases by local precipitation seeping through southwestern sandstone outcrops in wadis and mountainous areas (Gossel et al., 2004; Patterson et al., 2005; Sultan et al., 1997). Shallow horizons are recharged by local precipitation (Gossel et al., 2004; Patterson et al., 2005). Faults and folds create local areas for groundwater to move to the surface (Sultan et al., 1997).

Insolation variation caused by astronomical and earth processes drives episodic wet and dry phases across the Sahara. Increasing moisture in the Egyptian Western Desert resulted in surface water and environments amenable to hominins. The latter Pleistocene and early Holocene have evidence of wetter environments in the Western Desert and artifactual evidence of the presence of hominins.

Holocene Wet Phases in the Egyptian Western Desert

The Holocene African Humid Period provides a record of hominin movement prompted by the ebb and flow of regional precipitation through the presence of artifacts in datable contexts. During the terminal Pleistocene, the Sahara is thought to have extended 400 km farther south than present so that about a third of the African continent was covered by desert (Kuper and Kröpelin, 2006).

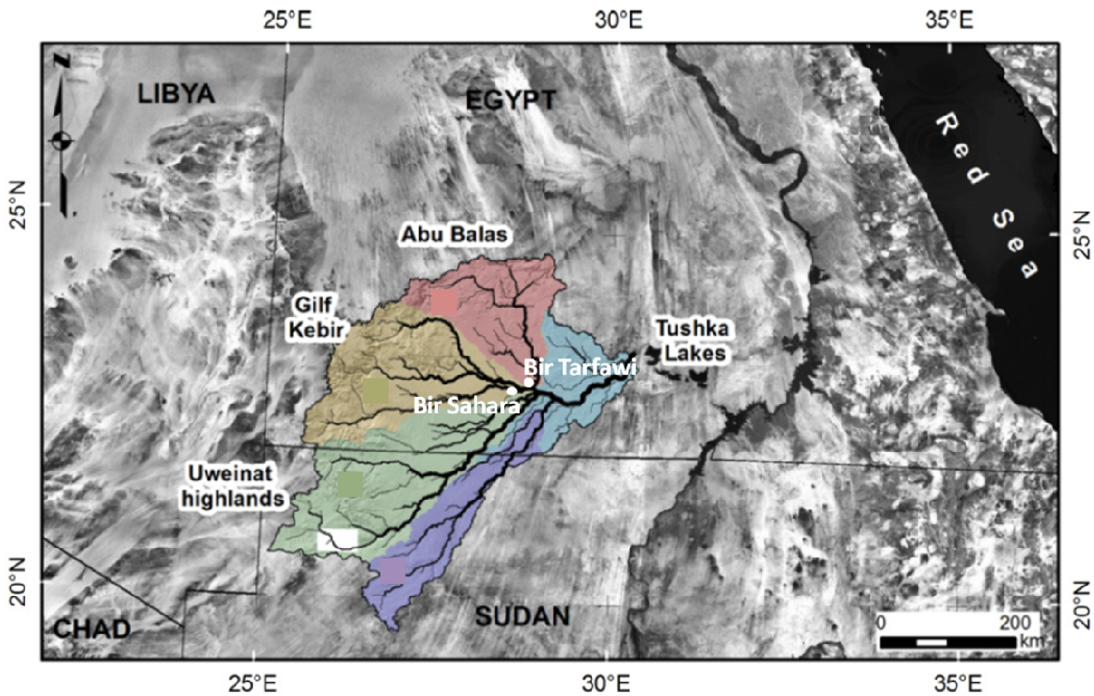

Figure 2.4 Paleodrainage channels between Gilf Kebir and Bir Tarfawi (Ghoneim and El-Baz et al., 2007).

The onset of the African Humid Period is synchronous with the end of cold, glacial conditions in Europe and North Atlantic Heinrich event 1. At this time, solar energy increased 8% over modern conditions due to the 20 ka precessional coincidence of perihelion and northern summer (deMenocal et al., 2000). Radiocarbon dates suggest moisture moved into the Egyptian Western Desert during the 10th millennium, providing sufficient rainfall, runoff, and aquifer recharge to provide surface water for human and faunal migration as far north as Farafra (28°N) (Kuper and Kröpelin, 2006; Nicoll, 2004).

Within just a few hominin generations, the ITCZ reached the farthest northern position, Monsoonal rains followed, resulting in an increase in surface water and Sahelian type vegetation (Kuper and Kröpelin, 2006). Based on the absence of mollusks and vegetation, regional precipitation was not uniform and in many areas produced only ephemeral lakes and runoff (Bubenzer and Reimer, 2007; Linstadter and Kröpelin, 2004). The northern areas of the monsoonal expansion such as Nabta, Dahkla, and Kharga (Figure 2.1) had surface water in the form of playa and artesian-fed lakes respectively (Wendorf and Schild, 1998; Smith et al., 2007; Kieniewicz and Smith, 2007). Northern Sudan has evidence of deep freshwater lakes such as Gebel Rahib (Kuper and Kröpelin, 2006; Hoelzmann et al., 2001). Farther south, Wadi Howar was carrying sufficient water to become a tributary of the Nile River (Kuper and Kröpelin, 2006; Pachur and Wünnemann, 1996). Kuper and Kröpelin (2006) describe this latitudinal gradient as semi-humid in the south (that is, in northern Sudan), and semi-arid in the central area corresponding to the Western Desert.

About 9.5 ka, the Selima region (20.7 to 22.5°N) of Northern Sudan (Figure 2.1) may have been similar to a wooded steppe desert with Sahelian plants typical of the modern day Ennedi area of Wadi Howar (16.5 to 19.0°N) (Haynes et al., 1989; Nicoll, 2004). Lakes formed at El 'Atrun (Wadi Howar region) and Selima (Figure 2.1) during the early Holocene, leaving records such as freshwater carbonates, sands, muds, and salts (Pachur and Wünnemann, 1996). It has been proposed that the West Nubian paleolakes (Figure 2.1) resulted from western-mountain drainage, local precipitation, and through a recharged aquifer (Hoelzmann et al., 2001; Pachur and Wünnemann, 1996). Farther north, within Egypt, Gilf Kebir (22.5 to 24°N) at 9.4 ka may have had vegetation indicative of steppe type habitat, including Rhamnaceae (buckthorn) and Tamaricaceae (tamarisk). Just northeast of Gilf Kebir, in Abu Ballas and Dahkla, evidence exists of semi-desert animals such as gazelle and elephant (Nicoll, 2004). The presence of hippopotamus in the Dahkla region may provide additional evidence for perennial artesian fed lakes. Playa type activity was present at Nabta and Wadi Tushka. By 8.7 ka, playa lakes were present at Dahkla (>25°N) (Brookes, 1989; Wendorf and Schild, 1998).

Western Desert moisture became less reliable beginning by about 7.0 ka. The declining availability of surface water after about 6.0 ka may have led to humans and animals using refugia such as Gilf Kebir and Nabta. A continuing decrease in the availability of surface water may be indicated by the exhaustion of the shallow aquifer after 4.5ka; this situation may have resulted in a movement of biota southward into the Sudan and eastward into the Nile Valley. This movement may have occurred at a rate of 35 km per year (Bubenzer and Reimer, 2007; Pachur and Wünnemann, 1996). The return to hyperarid conditions in the Eastern Sahara is marked by the lack of archaeological dates, indicating the complete lack of human viable habitats. Even Northern Sudan experienced reduced precipitation with the Nile tributaries Wadi Melik and Wadi Howar drying out (Kuper and Kröpelin, 2006; Nicoll, 2004).

In summary, paleoecology and archaeological evidence indicates that beginning at the glacial to interglacial transition and ending at the Middle Holocene, 5000 years of wet and dry cycles created environmental changes in the eastern Sahara. Hominin movement into and out of the Egyptian Western Desert followed the ecological change.

Pleistocene Wet Phases

<u>Bir Tarfawi</u>

A series of wet phases prior to the Holocene African Humid Period occurred from about 130 to 70 ka (Wendorf et al., 1993). It has been postulated that these climate changes were driven by the 20 ka precessional cycle coinciding with a maximum in orbital eccentricity creating a strong summer insolation maximum (deMenocal, 2008; deMenocal et al., 2000; Imbrie et al., 1984; Rossignol-Strick et al., 1998; Rossignol-Strick and Paterne, 1999; Rohling et al., 2002). This pluvial phase appears to have been particularly pronounced throughout the

Western Desert, with lakes forming at Bir Tarfawi and at Dakhleh (Wendorf et al., 1993; Smith et al., 2007; Kieniewicz and Smith, 2007; Osborne et al., 2008). The ca. 125 ka depositional sequence at Bir Tarfawi had a duration estimated between 17.4 and 5.8 ka years based on an average open ocean sedimentation rate (McKenzie, 1993). The duration of ~ 6 ka estimated by McKenzie (1993) would be similar to that of Holocene lakes. The Bir Tarfawi ca. 125 ka depositional sequence consists of three stages (Hill, 1993; McKenzie, 1993). The initial period characterized by highly evaporative conditions was capable of supporting Sahelian-type fauna such as giraffe (*Giraffa camelopardalis*), rhinoceros (*Ceratotherium simum*), antelope (*Kobus kob*), and gazelle (*Gazella dama, G. dorcas, G. rufifrons*). Fossils of crocodiles (*Crocodylus sp.*), hare (*Lepus capensis* and *L. whytei*), jackal (*Canis aureus/C. adustus*), and hyena (*Crocuta crocuta* and *Hyaena hyaena*), along with fish (*Pelusios* and *Geochelone*), were found in the Bir Tarfawi ca. 125 ka depositional sequence (Gautier, 1993). The presence of these faunal types implies wooded and savanna landscape within the vicinity of the lake (McKenzie, 1993; Wendorf et al., 1993b). Finally, the body of water became more ephemeral and eventually the area returned to arid conditions (Hill, 1993; McKenzie, 1993).

Earlier dates (150 to 220 ka), as well as a higher position above sea level, White Lake stratigraphic sequence precedes the Bir Tarfawi ca. 125 ka depositional sequence and contains the oldest Middle Paleolithic artifact assemblage in the area (Figure 2.5, Schild and Wendorf, 1993; Close and Wendorf, 1993). Middle Paleolithic lithics were found embedded in a shore segment on the western edge of the lake remnant (Close and Wendorf, 1993). No vertebrate fossils have been found, possibly due to limited excavations as well as severe erosion of shoreline areas (Schild and Wendorf, 1993; Close and Wendorf, 1993). The remnant consists of a depositional sequence capped by white limestone (indurated carbonate). Stratigraphic trenches were excavated into the White Lake remnant at its

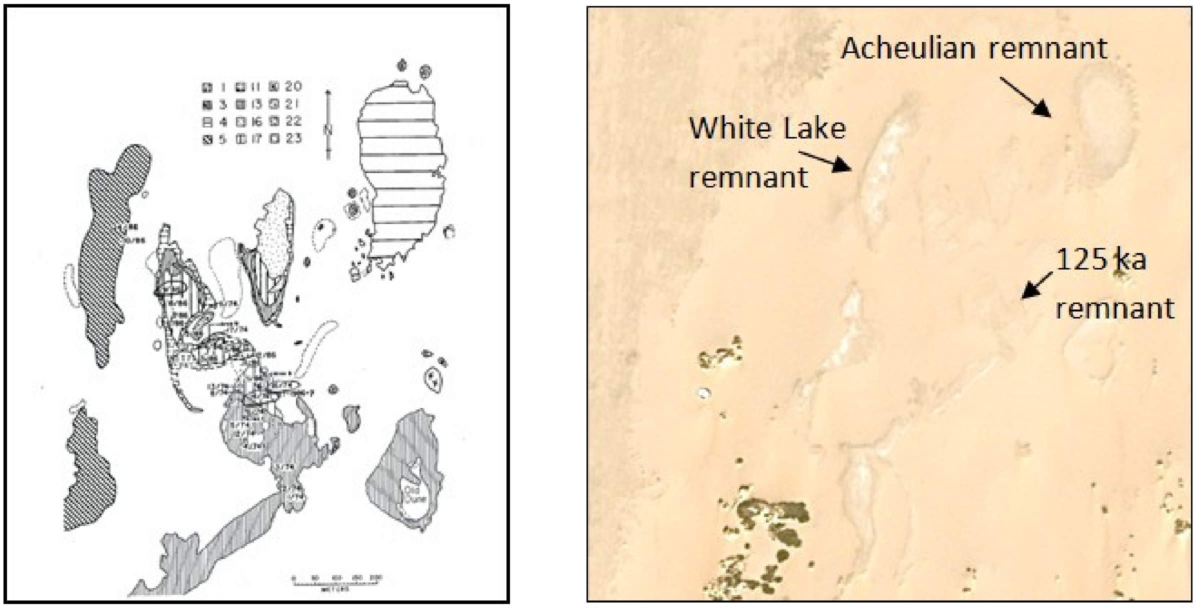

Figure 2.5 The Bir Tarfawi deflational basin Acheulian, White Lake and ca. 125 ka remnants (Wendorf and Schild, 1993; Google Earth).

center and edge, providing information on central lake and shoreline environments. The central lake sequence had a depositional pattern consistent with a long period of chemical sedimentation. It had a thickness similar to that of the ca. 125 ka depositional sequence, possibly implying a similar duration (Close and Wendorf, 1993; Hill, 1993).

Based on U-series measurements, the deposits of the White Lake are thought to be older than those of Oxygen Isotope Stage (OIS) 5. However, the large standard deviations of the U-series age from the White Lake sequence make it impossible to place this particular wet phase in either OIS 6 or 7 (Schwarcz and Morawska, 1993; Szabo et al., 1995). Marine core, vegetation tracers, and speleothem data from 180 ka suggest that during OIS 6.4 and 6.5 that the Sahel boundary moved northward in a manner similar to that of the OIS 5 and Holocene humid period (Tisserand et al., 2009). White Lake is considered too extensive to be part of a minor wet period during this arid phase, and investigators have instead placed this pluvial sequence in the beginning of a wet phase assigned to OIS 7 (Wendorf et al., 1993b).

CHAPTER THREE: HOMININS AND CLIMATE

The Western Desert often provided an impassible barrier to Pleistocene and Holocene hominins due to hyperarid conditions and associated lack of resources. However, there is sedimentalogical and ecological evidence of a number of wet periods providing a window for foraging range expansion (Derricourt, 2005; Lahr and Foley, 1998). Saharan arid periods roughly corresponding to northern glaciations lacked sufficient precipitation to support hominin resources outside of scattered oases. However, periods of waning glaciations brought rain and vegetation across the Sahara, providing resources for birds, mammals, and hominins (Finlayson, 2005; Kuper and Kröpelin, 2006; Wendorf et al., 1993; Wendorf and Schild, 1998). Even during wet phases, mobility would have been potentially critical for hominin survival due to a north to south gradient of surface water availability and a corresponding mosaic of vegetative types (Nicoll, 2004; Haynes et al., 1989). Saharan wet phases may have been short lived, and the return to arid conditions during the Holocene at least appears to have been swift (deMenocal, 2000; Tisserand et al., 2009; Vermeersch, 2001; Wendorf et al., 1993). This would have left much of northern Africa in a hyperarid and resource-free state. Ecologically this would have forced humans to return to previous ranges.

Holocene Hominins in the Western Desert

During the early Holocene, the apparent northward movement of precipitation into the Western Desert put water into wadis and deflational basins and established areas capable of sustaining vegetative, animal, and fish resources. Sudanese foraging populations with flexible survival strategies likely including fishing would have extended their ranges as similar resources became available just to the north in the Western Desert (Garcea, 2006; Kelly, 1992; Kuper and Kröpelin, 2006). The conditions along the Nile River throughout the early Holocene were difficult since the river was in a high stage (Kuper and Kröpelin, 2006). Cemetery sites contain skeletal evidence of violence, indicating difficult social conditions that may have encouraged population movement from the Nile Valley into the Western Desert (Kuper and Kröpelin, 2006; Wendorf, 1968).

Western Desert resources, even at the maximum of water availability, were concentrated on the banks of rivers and shore zones of lakes and in drainage systems, attesting to the patchiness of resources (Garcea, 2006). Artifacts including rock art, pottery, Neolithic tool kits, and anchoring stones are found mainly in water-related environments such as oases, playas, and lakes. However, open desert sites have been noted (Close, 1990; Kuper and Kröpelin, 2006). House foundations (Nabta Playa) and clusters of stone structures (Dakhleh) provide evidence for long term or multiple occupation sites in the Holocene Western Desert exclusively located above high water levels (Bubenzer and Reimer, 2007; Close, 1990; Garcea, 2006; McDonald, 1991; Wendorf and Schild, 1998). Hand dug wells are thought to have been used to access water in dry seasons and/or as the region returned to an arid state (Close, 1990; Nicoll, 2004). In the southern Western Desert, sand pan locations containing only a hearth and/or a few stone artifacts are interpreted as overnight sites utilized by nomadic herders (Close, 1990). Herding was a key addition to the survival strategy of BMHS during the African Humid Period (Garcea, 2006; Kuper and Kröpelin, 2006; Nicoll, 2004).

Between 6.3 and 4.7 ka, there was movement from the Western Desert back into the Nile Valley, into southern Sudan, and to a few highland refugia, leaving the area uninhabited by BMHS (Bubenzer and Reimer, 2007; Kuper and Kröpelin, 2006). The movement of

Holocene hominins out of the Western desert may reflect the advancing arid conditions as the ITCZ returned to its normal location.

Pleistocene Hominins in the Western Desert and East Africa

Late Pleistocene wet phases in the Western Desert are associated with key events in human evolution, including the transition between archaic and BMHS as well as the movement of BMHS out of Africa. Until approximately 200 ka, there were multiple species/subspecies of archaic hominins in Africa (McDougall et al., 2005; Grun, 2006; Lahr and Foley, 1998; Stringer, 2000; Stringer and Barton, 2008; White et al., 2003). Fossils of BMHS from Omo (195 ka, McDougall et al., 2005) and Herto (160 ka, Clark et al., 2003), Ethiopia provide evidence that a transition to modern morphology occurred between approximately 200 and 150 ka. Additionally, fossil discoveries at Singa, Sudan (OIS 6 > 130 ka) (Spoor et al., 1998) and Djebel Irhoud, Morocco (190-130 ka) (Hublin, 2001) demonstrate the movement of BMHS out of East Africa shortly after the transition associated with biologically modern humans. Sometime before 130 ka, conditions are postulated to have been so severe that only one hominin species (or subspecies) survived (Basell, 2008; Lahr and Foley, 1998; Mellars, 2005). Further, the surviving BMHS population is hypothesized from genetic studies to have been reduced to perhaps 10,000 individuals (Horai et al., 1995). During the subsequent wet period (130-70 ka) at Bir Tarfawi, the presence of BMHS fossils at Skhul, Israel (100-70 ka) show that BMHS were mobile (Mellars, 2005). While the Western Desert is not the only possible route between East Africa and the Levant, lake remnant artifacts demonstrate that this region was likely an integral part of BMHS movement.

East African Sites

Well-dated Pleistocene East African sites can be firmly placed in wet or arid climate periods. They provide clear evidence of archaic and biologically modern hominin response to climate change. The dominant Middle Paleolithic lithics that have been preserved are lithics, so site characterization is based on lithic assemblage types, material, composition, and site location (Basell, 2008).

One of the earliest BMHS is from Herto in the Ethiopian Highlands and dated to 160-154 ka (OIS 6). Although this is considered an arid period, this site was located along a shallow freshwater lake (Clark, 2003). The presence of extinct bovine, horse, and hippopotamus imply the region contained grassland habitats as well as aquatic resources (Clark, 2003). Middle Paleolithic East African sites assigned to arid cycle of OIS 7 and 6 were located across all types of vegetation zones, but always within fluvial and lacustrine margins similar to those of Herto. Trees or some type of cover were important but proximity to water appears to have been the critical factor in determining site location (Basell, 2008). Cover may have been achieved by locating sites at the edges of ecotones so that resources of both could readily be utilized. The sites of Singa and Abu Hugar in Sudan assigned to OIS6 (Basell, 2008; Grun, 2006) are located in the Ethiopian Highlands. Extreme desert conditions are thought to have prevailed in this region during OIS 6. The locations were habitable during this period because the sites were situated in the fluvial environments of the Blue Nile (Basell, 2008).

A substantial change in East African Middle Paleolithic hominin sites occurred after OIS 6 when caves, rock shelters, hillsides, and calderas were added to river and lakeshore site locations as habitation settings (Basell, 2008). Artifact differences after this period include decreases in the number of core axes and choppers and an increase in the variability of retouched pieces. After OIS 6, there is also an increase in the use of non-local resources (Basell, 2008).

Movement Out of East Africa

The route out of East Africa to the north and/or west likely leads through the eastern Sahara since an alternate route, a water crossing at Bab el-Mandab (Africa to Arabian peninsula over Red Sea), was probably not used until after 100 ka (Armitage et al., 2011; Derricourt, 2005; Forster and Matsumura, 2005). Even with reduced water levels, the gap between Africa and the Arabian peninsula is thought to be a significant barrier without boat technology and there is currently no evidence of a land bridge (Derricourt, 2005). A coastal site exists at Abdur Eritrea (ca. 125 ka) where tools are embedded in a raised coral bed (Carto et al., 2009; Walter et al., 2000; Stringer, 2000). Arid conditions could provide the motivation for coastal movement, but modern high sea levels may be a deterrent to acquiring significant evidence for this choice (Carto et al., 2009).

If a sea crossing were problematic during the Pleistocene, there were three likely Egyptian routes: the Nile Valley, the Eastern Desert, and the Western Desert (Derricourt, 2005; Lahr and Foley, 1998; Vermeersch, 2001; Van Peer, 1998). The Nile Valley has little archaeological evidence of use during the transition between archaic and BMHS*,* and data available at present is difficult to interpret due to lack of a reliable chronology (Vermeersch, 2001). While the Nile Valley may seem the most direct route out of the Sahel, periodic high waters would have provided a barrier to passage, and floods may have destroyed any artifacts left by hominins. Sodmein Cave in the Eastern Desert (Figure 2.1) has evidence of use over long periods for short intervals with dates correlating only to those of the last Pleistocene wet period and the BMHS movement to the Levant (Derricourt, 2005). However, the

Eastern Desert has the same climatic constraints as the Western Desert, that is, passage is likely only during wet phases (Derricourt, 2005). To date, the best evidence for a desert route exists in the Western Desert. Wet period surface water in the Western Desert provided resources that could lead to both western Africa and the Levant, reducing the need for coastal routes or use of the Nile Valley for movement out of East Africa during the Middle Paleolithic.

Bir Tarfawi Sites

The White Lake sequence (220-150 ka, McKenzie, 1993; Schwarcz and Morawska, 1993), containing the earliest Middle Paleolithic artifacts at Bir Tarfawi, potentially coincides with the transition period from archaic to BMHS (Basell, 2008; Lahr and Foley, 1998; Mellars, 2005). White Lake artifacts were embedded in dark grey sand associated with the beach or near shore of a lake (Close and Wendorf, 1993). Cores and debitage were not collected so assignment to a particular techno-typological lithic assemblage was not attempted. Artifacts were predominately composed of quartzite sandstone. This material is available from "granite outcrops on the eastern edge of the Tarfawi depression" (Close and Wendorf, 1993). Hominin presence may have occurred during an early lake phase (Wendorf et al., 1993). No vertebrate remains were found in the limited excavations at the White Lake.

The subsequent sequence of sedimentary deposits and Middle Paleolithic artifacts at Bir Tarfawi is associated with OIS 5 (and possibly OIS 6) and the postulated movement of BMHS out of Africa. Population pressure to extend foraging ranges was not likely during this wet phase since the preceding arid period corresponds to the loss of hominin species/subspecies and a possible reduction in genetic diversity of BMHS (Lahr and Foley, 1998). Arid conditions in tropical Africa have been proposed as the impetus for populations to move ranges north during this time (Olszewski et al., 2010). Additionally, modern foragers are known to move to resources, so the availability of new resources in the Western Desert may have been part of typical foraging strategies and sufficient reason to move into the Western Desert (Kelly, 1992).

Patterns of stone tool use associated with this wet phase have been used to propose that an arid landscape strategy persisted throughout the Bir Tarfawi ca. 125 ka sequence (Wendorf et al., 1993a). Extraction of lithic raw materials and initial shaping occurred at quarries, while lakeside areas appeared to be used for secondary tool working (Wendorf et al., 1993a). The generalized typology of the lithics implies the site was used for a wide range of activities. Lithic raw material is mainly from the outcropping about 3 km east of Bir Tarfawi and this is considered the likely source for raw materials used during the White Lake wet phase. While there are thousands of artifacts at Bir Tarfawi ca. 125 ka remnant, little change is noted in morphology over the time of the lake's existence. The presence of lithics in waterlain sediments suggests the area was used in dry seasons and/or that the lithic assemblages have been affected by sedimentalogical processes (Hill, 1993).

The Bir Tarfawi ca. 125 ka lake was used by hominins and animals likely due to a concentration of resources near the lake. The large number of distal fragments of gazelles and the fact that animal fossils have only been found in conjunction with artifacts is interpreted as hominin responsibility for animal bones through either scavenging or hunting (Gautier, 1993). No hearths were found in the area of the lake, leading to the interpretation that Bir Tarfawi (BT14) was a day use site, while at night the population would occupy areas away from the lake for protection from predators (Close and Wendorf, 1993).

Discussion

Until the latter part of the Holocene wet phase, where evidence suggests by 8-9 ka that herding was part of survival strategy, hominin populations present in the Western Desert were likely hunter-gatherers. A key characteristic of hunter-gatherers is mobility (Kelly, 1992). Hunter-gatherers move to resources and while each group may have a home range, the expansion of that range to utilize new resources would require resource mapping but not any significant change in strategy (Kelly, 1992; Marlowe, 2004). Additionally, increased precipitation in the Western Desert would expand the landscape in which these hunter-gatherers were already utilizing resources.

It has been proposed that Holocene and Pleistocene Western Desert hominin sites were mainly tied to wadis and lake edges predicated on resources concentrated near these areas (Garcea, 2006). These are the only locations for long duration sites during the Holocene. Lithics have been found in the intervening sand pans in both periods (Close, 1990; Garcea, 2006; Kuper and Kröpelin, 2006; Nicoll, 2004; Olszewski et al., 2010).

The types of lithics and lithic assemblages are different for Holocene and Pleistocene hunter-gatherers. Pleistocene hunter-gatherers have left behind predominately lithics. Holocene hominins also used lithics; their mark on the Western Desert includes rock art, pottery, and anchoring stones, as well as more distinct archaeological features such as wells, huts, and hearths. Anchoring stones are a hallmark of the North African pastoralists and document this change in hominin lifeways (Basell, 2008; Kuper and Kröpelin, 2006; Nicoll, 2004; Wendorf et al., 1993).

The inherent mobility in the lifeways of Pleistocene and Holocene hunter-gatherers makes the use of the Western Desert a natural extension of their adaptive strategy as conditions improve. This same mobility and range extension also explains how hominins reached the western edges of Africa and the caves of the Levant, making the Western Desert key in these seminal events of hominin evolution.

CHAPTER FOUR: STABLE ISOTOPE ANALYSIS

For each atom type or element, the number of protons and electrons is fixed. The number of neutrons for each element varies, creating different varieties known as isotopes. In the example of oxygen, there are three isotopes, oxygen 16 (^{16}O), oxygen 17 (^{17}O), and oxygen 18 (^{18}O), where the superscript value refers to the atomic mass of the particular isotope. The atomic mass is equal to the number of protons plus the number of neutrons. Oxygen has eight electrons and eight protons; however, the number of neutrons can be 8, 9, or 10 (Pauling, 1970). This difference in the number of neutrons is critical in the behavior of the element since the mass difference between isotopes that creates differentiation during processes such as evaporation, condensation, and photosynthesis results in a change in the isotopic composition of a material. This isotopic mass differentiation is known as fractionation and is the basis for stable isotope analysis (Clark and Fritz, 1997; Hoefs, 2004; Kendall and Caldwell, 1998).

Stable isotope analysis uses a mass spectrometer to measure the number of each isotope type, capturing these small mass differences resulting from natural processes. The hydrologic cycle is one such process, and therefore changes in the oxygen isotopic composition of water provide information such as source water and details of evaporation and condensation of particular storm paths. Carbon isotopes are also commonly studied due not only to the prevalence of biological systems but also because biological systems can be proxies for palaeoclimate studies. In this study, carbon and oxygen stable isotope values from the shells of the gastropod *Melanoides tuberculata* are used as a proxy for climatological information such as precipitation source and perennial versus ephemeral lake characteristics.

Fractionation

The differences in mass between isotopes create differences in the behavior of molecules. Of primary importance is a change in the dissociation energy or the amount of energy to separate the atoms in a molecule. The energies involved in isotope mass differences are greatest for elements with low mass since each neutron addition is a change of only one mass unit and that change is most pronounced in the low mass range (Chacko et al., 2001; Kendall and Caldwell, 1998). Water evaporation provides a good example of the fractionation process; water molecules composed of ^{1}H and ^{16}O are more likely to evaporate than more massive molecules containing ^{2}H (or D) and ^{18}O. (D refers to deuterium, which is the isotope of hydrogen with two neutrons.)

Dissociation energy is affected by changes in mass since it is governed by quantum mechanical effects and to a first order the molecule can be thought of as a simple oscillator (Chacko et al., 2001; Kendall and Caldwell, 1998). Quantum mechanical theory implies that the energies available are discrete and that there is not a zero energy state. Known as zero point energy, all molecules even at absolute zero have a small vibrational energy ($1/2\ h\nu$ where h is Planck's constant and ν is the vibrational frequency of the molecule) (Chacko et al., 2001). Changes in molecular mass cause changes in the zero point energy of the molecule because the vibrational frequency (ν) is dependent upon the reduced mass of the molecule.

Simple oscillators can be described using Hooke's Law. Hooke's Law states that the force between the masses of an oscillator is equal to the negative of the force of the spring and the distance between the masses ($F= -kx$). The molecular bond strength (k) is a result of the electron configuration, and this is not significantly affected by isotopic differences (Bauman, 1962; Chacko et al., 2001). The potential energy of this oscillator or molecular system is defined as $kx^2/2$ and is represented by a parabola. Since the spring constant is defined by the electron configuration, the energy of the molecular system defined by the parabola is the same regardless of the mass of the isotopes.

Finally, the frequency of oscillation in the molecular system is a function of the bond strength and the reduced mass $\nu = (k/\mu)^{1/2}/2\pi$. Changes in dissociation energy are primarily a result of the reduced mass (μ) component. Reduced mass is equal to the product of the atomic masses divided by the sum of the atomic masses, $(m1*m2)/ (m1+m2)$ (Chacko et al., 2001; Hoefs, 2004). Energies in the simple oscillator are represented by a parabolic energy well (Figure 4.1), negative potential energies, and a minimum or zero point energy for each isotope configuration. Heavier masses have energy levels that are lower in the potential well with correspondingly lower zero point energies with shorter bond lengths, implying maximum stability. Correspondingly, lighter molecules have a higher zero point energy or a smaller absolute potential energy and are thus more likely to be dissociated and exchanged. In general, processes that require less energy or have a product with a lower energy state are favored.

There are two processes that result in fractionation: equilibrium isotope exchange reactions and kinetic processes (Chacko et al., 2001; Gat, 1996; Hoefs, 2004). Equilibrium isotope exchanges result in no change in materials but rather a change in isotopic distribution of materials or phases or molecules. The oscillator model predicts that heavier molecules are more stable and that heavier isotopes are attracted to stronger bonds. Higher oxidation states attract the heavier isotope (Hoefs, 2004; Kendall and Caldwell, 1998).

Temperature is a factor in these isotope exchanges. As temperature increases, the differences in zero point energies are reduced since the molecules are no longer in the ground state and higher energy states tend to have smaller differences, reducing the energy reduction incentives for isotopic partitioning (Chacko et al., 2001).

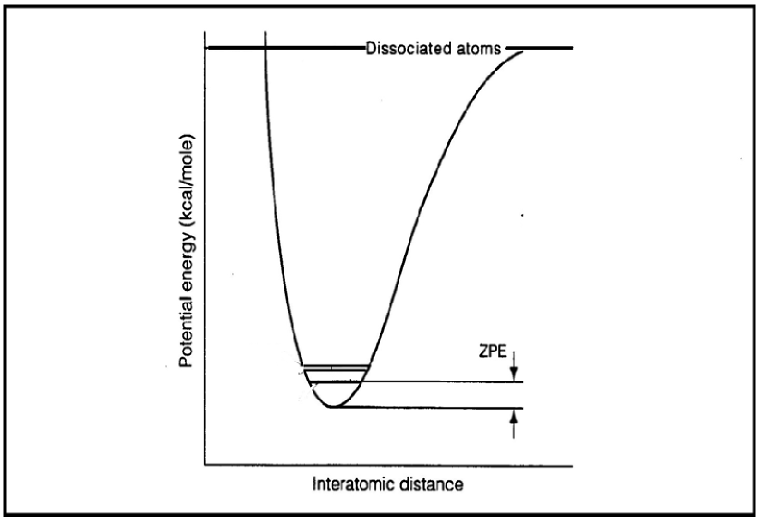

Figure 4.1 Potential energy relationship to interatomic distance (adapted from Kendell and Caldwell, 1998).

Kinetic processes are often characterized by a rate of reaction that is sensitive to atomic mass. This is the case for processes such as evaporation, many biological reactions, and diffusion. In these processes, isotopic composition is governed by the fact that lighter isotopes have bonds that are dissociated more easily and thus react faster. This results in an increase in the concentration of the lighter isotope in reaction products (Hoefs, 2004; Kendall and Caldwell, 1998).

The isotopic composition of most materials is measured using a mass spectrometer. It is a fact of measurement that the accuracy of absolute isotope composition is poor but relative isotopic compositions can be made with very good precision (Hoefs, 2004).

Therefore, stable isotope measurements are made using relative measurements defined by the δ (delta) value. The δ value is defined as

$$\delta\ (‰) = \left[\frac{\left(\frac{R}{C}\right)_{samp} - \left(\frac{R}{C}\right)_{std}}{\left(\frac{R}{C}\right)_{std}}\right] x10^3 = \left[\frac{\left(\frac{R}{C}\right)_{samp}}{\left(\frac{R}{C}\right)_{std}}\right] x10^3 - 1000$$

R refers to the rare or heavier isotope. C represents the common, lighter or most abundant isotope. R/C is the measured ratio of the isotopes for a sample and the standard (Hoefs, 2004). Per mil is the unit for the δ value as the differences in isotope ratios are small and by convention are multiplied by 1000. Materials with higher heavy isotope concentrations relative to the reference standard have positive δ values.

Standards for oxygen in water are SMOW (Standard Mean Ocean Water) or VSMOW (a synthetic version of SMOW certified by IAEA (International Atomic Energy Agency). The δ value for ^{18}O in SMOW is 0 (zero). Carbonates are measured against the PDB (PeeDee Belemnite) Standard. The PDB standard for carbonates is used for both ^{13}C and ^{18}O with δ values of 0 and 30.4‰ (SMOW) respectively (Hoefs, 2004).

Hydrologic Cycle

Continental precipitation is the result of evaporation from surrounding oceans at the beginning of the hydrologic cycle. In the basic hydrologic process, water evaporates from the ocean surface and is transported into the atmosphere. As the water vapor rises in the atmosphere, it cools and condenses into clouds and moves over the continent or other land mass and falls as precipitation. The processes of evaporation and condensation cause fractionation or a change in the ratios of the stable isotope composition. Dansgaard (1964) noted the consequences of these processes in the hydrologic cycle: ocean water has a somewhat consistent isotopic value; fresh water has a lower oxygen isotopic value than seawater; oxygen isotope values for fresh water are inversely proportional to latitude and altitude; precipitation isotope values vary with source, temperature, and amount.

The change of phase of water from liquid in the ocean to vapor in the atmosphere is the first fractionation process in the hydrologic cycle. Isotopically depleted water molecules are preferentially evaporated in the vapor phase as predicted by both equilibrium exchange and kinetic process theories. Conversely, the remaining ocean reservoir will have a reduction in the depleted molecular species and have increased concentrations of enriched water molecules (Gat, 1996; Hoefs, 2004; Kendall and Caldwell, 1998). Evaporation is controlled by a diffusion of vapor across the water/air boundary layer, by temperature and by humidity. When humidity is 100%, evaporation essentially occurs under closed conditions where the back flux of the atmosphere and upward flux of the ocean surface are equal; therefore, isotopic equilibrium allows isotope exchange conditions. At humidity levels less than saturation, a flux condition is set up, and the diffusion across the boundary layer is the determining step. While the details of the evaporation process change with specific conditions, the result is always a decrease of the ^{18}O isotope value in the vapor phase. Measurements of marine moisture by the International Atomic Energy Agency (IAEA) show that water vapor over the ocean is not uniform and the $\delta^{18}O$ values are more negative than expected since in reality there is a combination of isotope exchange and kinetic effects (Hoefs, 2004; Gat, 1996, 2000). These differences in the isotopic value of oceanic water vapor are an indication of the effects of temperature and humidity on the evaporation process.

By the time the atmospheric water vapor reaches a land mass, mixing of non-uniform moisture packets will have occurred and resulting precipitation will have a $\delta^{18}O$ value that lies along the Global Meteoric Water Line (GMWL) defined by $\delta(D) = 8\delta(^{18}O) +$

10% (Dansgaard, 1964; Gat, 1996, 2000; Hoefs, 2004). Meteoric waters refer to "air moisture and equilibrium precipitation (Gat, 2000). There are numerous local meteoric lines depending upon the source and the evolution of the precipitation.

Three parameters control the isotopic composition of precipitation over landmasses: temperature, rainfall amount, and size and morphology of the land mass (continental effect) (Dansgaard, 1964). Temperature and type of cooling within clouds will control the fractionation of the phase change from vapor to solid and to liquid. Regardless of the cooling conditions (i.e., constant pressure, volume or heat), precipitation is well modeled as a Rayleigh process. Rayleigh processes are characterized by the continuous removal of the condensate, which results in increased fractionation over equilibrium processes since the vapor phase will have a continuous reduction in the ^{18}O isotope concentration and, so, is represented by curves with a continuously changing slope (Dansgaard, 1964).

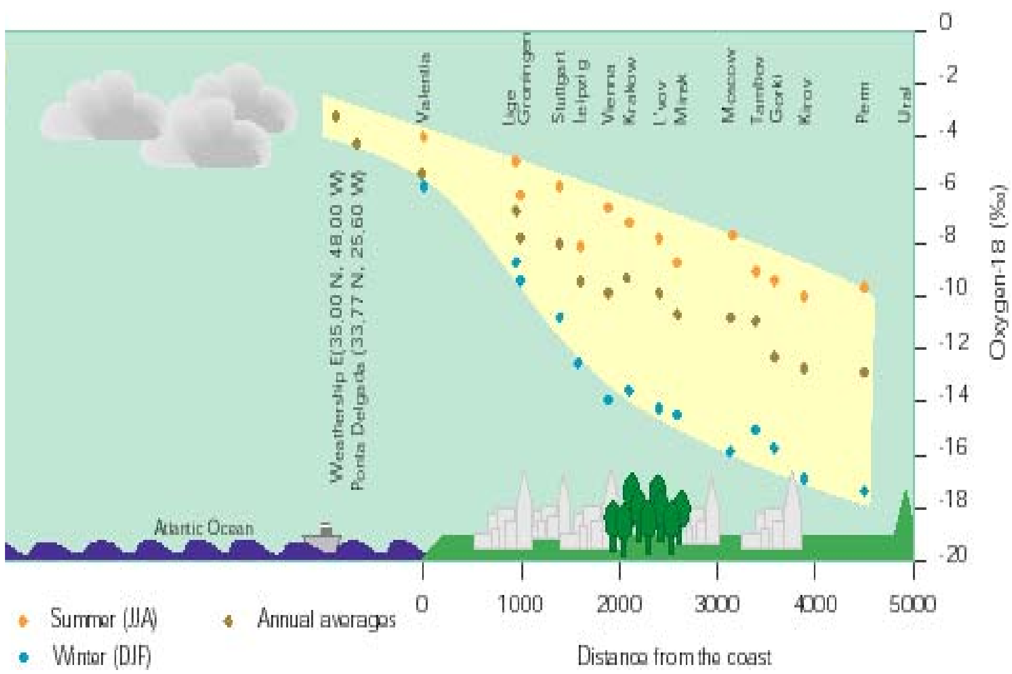

Figure 4.2 The continental effect (IAEA, 2006).

As a result of the condensation and precipitation process as a storm track continues over land, the heavy isotope deficit increases. Evaporation of continental lakes, rivers, and forests can replenish cloud vapor and increase isotope value (Figure 4.2). The condensation process begins immediately with this new infusion of water vapor, again resulting in the reduction of the isotope value. As the air parcel tracks over land, these processes repeat and ultimately result in a reduction in the oxygen isotopic composition that is a function of distance from the coast and replenishment sources. An additional factor in this continental traverse of the air mass is orography or altitude. As the storm moves over the landmass, mountains push the parcel higher into the atmosphere, causing a decrease in temperature and a decrease in the ^{18}O isotope value of the vapor. While altitude is often considered the parameter affecting the isotopic value, a decrease in temperature is the actual driver. Taken together, changes in isotope values due to distance from the original ocean source and altitude are known as the continental effect (Dansgaard, 1964).

Local precipitation isotopic values are controlled by atmospheric conditions at the site of the precipitation, mainly temperature and humidity. Normally, there is an exchange between precipitation and the moisture of the atmosphere through which it is falling, causing an increase in the ^{18}O isotope value for the precipitation (Dansgaard, 1964; Gat, 1996, 2000). The isotopic exchange between precipitation and the atmospheric column also implies that the isotopic values correlate with ambient air temperature (Gat, 2000). Where there is low humidity, evaporation of precipitation can occur as it falls through the air column. Evaporation always causes isotopic enrichment in the denser phase through selective removal of the lighter isotope. In the case of precipitation through dry air, the $\delta^{18}O$ value will decrease further. The final controlling parameter of isotopic composition for precipitation is the amount effect (Dansgaard, 1964). This is the only factor controlling isotope composition that is not strictly temperature-dependent. When precipitation is heavy, that is, falls in large amounts, there is insufficient time for atmospheric vapor/precipitation exchange and therefore the precipitation remains highly depleted. The amount effect is associated with thunder storms, cold fronts, and tropical convergence zones (Gat, 2000).

Even with all of these variables, storms tracking over the same path in the same seasons produce precipitation with isotopic compositions that are highly predictable (Dansgaard, 1964; Gat, 1996; Joseph et al., 1992). This predictability provides a means of deciphering past storm tracks in a specific area as long as proxies for the isotopic conditions of the precipitation can be found. Animals such as gastropods, which secrete shells in isotopic equilibrium with the water they inhabit, maintain such a record. Shells are often well preserved, and therefore isotopic information recorded in shell material can act as a proxy to help understand past storm systems (Abell, 1985). However, gastropods such as *Melanoides tuberculata* require standing surface water such as lakes to survive, although hibernation in soft sediments at temperatures below ~21°C has been noted (de Kock and Wolmarans, 2009; Pointier et al., 1992; Livshits and Fishelson, 1983).

How well lake water represents the isotopic composition of the precipitation or groundwater from which it is formed depends mainly on the volume, residence time of the water, seasonality, and temperature (Leng and Marshall, 2004; Li and Ku, 1997). These characteristics tend to act in tandem to determine the isotopic composition of a lake. Small-volume open lakes with an inlet and outlet generally have short water residence times; therefore, water in the lake will have isotopic compositions reflecting the precipitation or the groundwater that replenishes it. This means there can be significant differences in the isotopic composition based on season. Storms may track from different source bodies based on the time of the year and local seasonal temperatures will modify the precipitation isotopic composition. Small lake isotopic signals can reflect seasonality, temperature, and source water inputs (Leng and Marshall, 2004; Li and Ku, 1997).

Slightly larger open lake volumes are likely to have increased residence times, perhaps longer than a year (Leng and Marshall, 2004). This will reduce or eliminate the seasonal isotopic signal since the incoming water will mix with water from previous seasons. There is still likely to be a strong temperature signal in the medium to small lake as well as preservation of the source water isotopic composition. As lakes increase in volume, the ability to distinguish signals from temperature or isotopic composition of specific precipitation events diminishes. Large lakes have volumes sufficient to mask the isotopic values of individual storms. In large lake systems, isotopic values are controlled by the ratio of precipitation to evaporation. Evaporation can become the primary controller of isotopic composition, causing values to increase significantly as the ratio of $^{18}O/^{16}O$ increases in the denser liquid phase (Leng and Marshall, 2004).

CHAPTER FIVE: MATERIALS AND METHODS

Stratigraphic trenches were excavated into the White Lake remnant at its center and its edge. The deep lake section (trench 14/86) had a depositional pattern consistent with a long period of chemical deposition and was analyzed by McKenzie (1993) (Figure 5.1). The shoreline deposits contain six units of both sandy and muddy strata consistent with the ebb and flow of a lake beach (Hill, 1993). *Melanoides tuberculata* gastropods analyzed in this study were selected from the coquina in the sandy deposits of the unit 3 trench 10/86 shoreline depositional sequence (Hill, 1993).

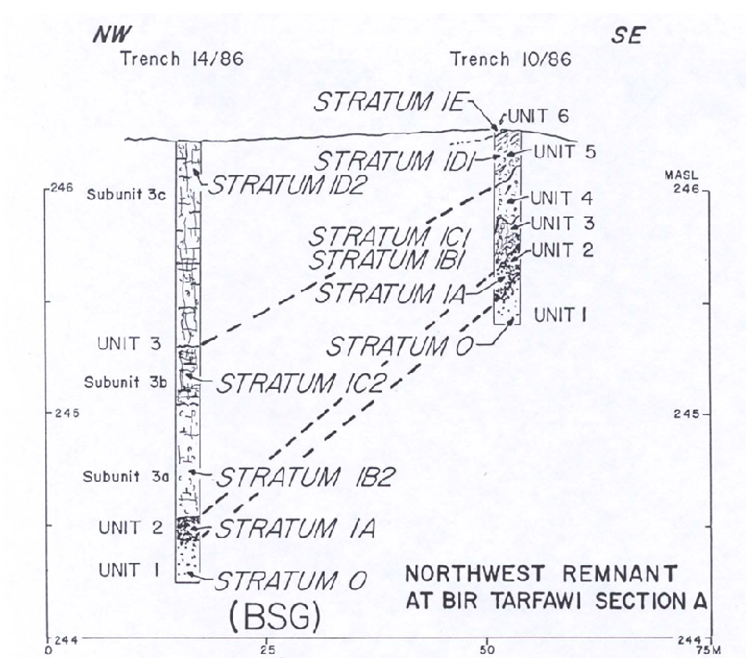

Figure 5.1 Cross-section of White Lake remnant excavations trenches 14/86 and 10/86. *Melanoides tuberculata* coquina trench 10/86 unit 3 (Hill, 1993).

Melanoides tuberculata was taken from two Bir Tarfawi ca. 125 ka remnant trenches near BT-14. Trench 7/74 had a long depositional period indicating deep water and the second trench had a short depositional pattern consistent with a shallow water area. Gastropods from trench 7/74 or deep water were present in 10 of 14 strata, while *Melanoides* from the shallow water trench was found in only one stratum. *Melanoides* shells from the ca. 125 ka remnant were analyzed to provide comparison with White Lake remnant gastropods.

The gastropod *Melanoides tuberculata* lives in sandy, muddy, or sandy-muddy sediments, as well as in stony-sandy strata of rivers, lakes, and streams with salinity less than 35 ppt and a temperature range of 16°C to 30°C (de Kock and Wolmarans, 2009). *Melanoides* has been observed to hibernate by burrowing into soft sediments at temperatures below 21°C (Livshits and Fishelson, 1983). Life span and size are variable and dependent upon environment. Growth is controlled by pH, calcium, and nutrient availability, as well as by temperature (Livshits and Fishelson, 1983; Pointier et al., 1992). Nutrients are supplied by dissolved carbonate in the water and decaying flora such as aquatic plants, algae, epiphytic organisms, and adhering detritus (de Kock and Wolmarans, 2009; Livshits and Fishelson, 1983; Pointier et al., 1992). *Melanoides* reaches reproductive maturity at a length of 15 to 16 mm within 6 months at temperatures between 26°C and 32°C (Livshits and Fishelson, 1983; Pointier et al., 1992). After reaching sexual maturity, shell length growth slows with shell widening taking precedent to increase brood size (Pointier et al., 1992). The aragonite shell is secreted throughout life in equilibrium with surrounding waters preserving the water's isotopic composition.

Melanoides tuberculata specimens (Figure 5.2) from the coquina of unit 3 trench 10/86 White Lake and shallow and deep water areas of Bir Tarfawi ca. 125 ka remnant were in excellent preservation, with unabraided, unbleached surfaces suggesting little postdepositional disturbance. *Melanoides* shell is composed of aragonite, a metastable form of carbonate that can recrystallize into calcite, altering the isotopic composition to conditions contemporaneous with the recrystallization. X-ray diffraction analysis (Bruker AXS D8 Discover X-Ray Diffractometer at Boise State Center for Materials Characterization) of powdered shell samples had only aragonite peaks. The lack of calcite peaks indicates that the shell's isotopic composition is unaltered and therefore represents the environmental conditions that existed during the secretion of the aragonite shell.

Figure 5.2 *Melanoides tuberculata* from the coquina of unit 3 trench 10/86 of White Lake remnant.

Melanoides tuberculata shells from 10/86 unit 3 coquina and the ca. 125 ka remnant were ultrasonically washed and rinsed with distilled water. Cleaned shells were roasted at 200°C for 2 hours. Small (1.0 to 1.2 cm) and medium (1.5 to 1.8 cm) shells were crushed using a mortar and pestle for whole shell analysis. Large (1.9 to 2.8 cm), medium shell, and small segments were subsampled using a thin hand-held blade and then powdered. Weighed powdered samples of 0.4 mg were reacted in a He gas-purged borosilicate exetainer with

0.05 mls of concentrated phosphoric acid until completely dissolved (~1 hr). Evolved CO_2 samples were analyzed using CF-IRMS (continuous flow isotope ratio mass spectrometry) on a Thermo Delta V Plus IRMS with a Gasbench II headspace sampler. Carbon and oxygen stable isotope values are reported using δ notation ($\delta^{13}C$ and $\delta^{18}O$) as per mil VPDB. A correction of -0.6‰ was made to aragonite oxygen isotope values since aragonite is enriched in ^{18}O with respect to calcite (Kim and O'Neil, 1997). This correction allows a direct comparison of oxygen isotope values between sediment carbonate (calcite) and gastropod shell (aragonite).

Oxygen stable isotope values were compared with published results from water sources and palaeoclimate proxies from the region including precipitation from the Atlantic and the Indian Oceans and the Mediterranean Sea, as well as deep aquifer groundwater (Hoelzmann et al., 2001; McKenzie, 1993; Smith et al., 2004; IAEA, 2006). In addition, it has been proposed that water delivery to the White Lake may have occurred from paleochannels emanating from the western mountains (Ghoneim and El-Baz, 2007) or from the highlands of the Nile headwaters (Maxwell et al., 2010).

Precipitation oxygen isotope values for the Mediterranean Sea and Indian Ocean have not been measured at Bir Tarfawi. Estimates for these source waters were made using modern precipitation values for the locations nearest Bir Tarfawi representing moisture from these sources: Khartoum, Sudan -2.09‰ SMOW (Indian Ocean source; Joseph et al., 1992) and Sidi Barrani, Egypt -3.4‰ SMOW (Mediterranean source; IAEA, 2006). An oxygen isotope continental effect relationship for Africa was calculated by Joseph et al., (1992), based on moisture transported west from the Indian Ocean across the Sahalo-Sudanese zone, to be -0.08‰ per 100 km. Assuming this correction applies to air masses traveling from Kharga or Sidi Barrani to Bir Tarfawi, the Indian Ocean and Mediterranean Seasourced precipitation values at Bir Tarfawi would be -2.9‰ SMOW and -4.2‰ SMOW, respectively.

Determination of potential oxygen isotope values for equilibrium precipitation of sediment carbonate from specific water isotopic compositions (source water) and temperatures were made using the equation of Kim and O'Neil (1997):

$$1000\ln\alpha_{\text{(Calcite-H2O)}} = 18.03(10^{3}T^{-1}) - 32.42$$

The Grossman and Ku (1986) equation was used to evaluate the equilibrium precipitation of shell carbonate (aragonite) and water oxygen isotopic compositions at various temperatures (Figure 6.14):

$$1000\ln\alpha_{\text{(aragonite-H2O)}} = 18.07(10^{3}T^{-1}) - 31.08$$

Conversions between VPDB and VSMOW utilized the relationship from Friedman and O'Neil (1977). This conversion is necessary to utilize the Kim and O'Neil (1997) equation:

$$\delta^{18}O_{VSMOW} = 1.03086 * \delta^{18}O_{VPDB} + 30.86$$

CHAPTER SIX: RESULTS AND DISCUSSION

The results from carbon and oxygen stable isotope analysis of whole and subsampled *Melanoides tuberculata* from unit 3 trench 10/86 Bir Tarfawi White Lake remnant are shown in Table 6.1, Figure 6.1, and Appendix A. $\delta^{13}C$ overall values vary over a smaller range (4.8‰) with lower standard deviation (0.8) than $\delta^{18}O$ overall values (8.5‰, 1.9). $\delta^{13}C$ values range between a high of +1.0‰ and a low of -3.7‰. $\delta^{18}O$ values vary between -+0.8‰ and -7.7‰. NBS 19 standard deviations were 0.13 and 0.23 for $\delta^{13}C$ and $\delta^{18}O$, respectively.

White Lake *Melanoides tuberculata* were taken from the shallow remnant area. Sediment carbonates measured by McKenzie (1993) were from the deep area trench 14/86. In previous studies (Table 6.4), gastropods and sediment were co-located. To better understand the relationship between deep water sediment carbonate and shallow water gastropod shell carbonate, *Melanoides tuberculata* shells from deep and shallow water Bir Tarfawi ca. 125 ka remnant were measured. Trench 7/74 deep water sediment carbonate was measured by McKenzie (1993).

Bir Tarfawi ca. 125 ka remnant shallow and deep water *Melanoides* shell carbonate results are presented in Tables 6.2 and 6.3, Figures 6.2 and 6.3, Appendixes B and C. Overall $\delta^{13}C$ means for the deep and shallow gastropod carbonates were -3.7‰ and -3.3‰, respectively. $\delta^{13}C$ deep section values ranged from -6.2‰ to 0.2‰ while the shallow values maximum and minimum were -2.1‰ and -5.6‰. The $\delta^{18}O$ overall mean value for deep section gastropod carbonates was -5.2‰ and the shallow section value was -3.2‰. Ranges

Table 6.1 Summary of Oxygen and Carbon Stable Isotope Compositions for White Lake Remnant *Melanoides tuberculata* shells.

Sample	Number of samples	$\delta^{13}C$ (‰, VPDB)					$\delta^{18}O$ (‰, VPDB)				
		min	**mean**	max	std	**range**	min	**mean**	max	std	**range**
Overall	99	-3.7	**-2.4**	1.0	0.8	**4.8**	-7.7	-4.3	0.8	1.9	**8.5**
Whole Shell											
medium	3	-3.4	**-2.8**	-2.2	0.6	**1.2**	-4.9	**-4.0**	-3.4	0.8	**1.5**
small	3	-3.5	**-2.7**	-2.2	0.7	**1.3**	-5.5	**-3.6**	-1.0	2.3	**4.5**
Subsampled Shell											
large 1	11	-3.4	**-2.4**	-1.2	0.7	**2.2**	-7.1	**-5.2**	-3.6	1.0	**3.5**
large 2	11	-3.3	**-2.9**	-2.2	0.4	**1.1**	-6.6	**-5.0**	-2.9	1.4	**3.6**
large 3	12	-3.4	**-2.1**	-1.5	0.5	**1.8**	-7.7	**-5.6**	-3.9	1.1	**3.8**
medium 2	8	-2.8	**-2.2**	-1.8	0.4	**1.1**	-2.8	**-2.2**	-1.8	0.4	**1.1**
medium 5	12	-3.7	**-2.2**	-0.4	1.0	**3.3**	-7.3	**-3.1**	0.8	2.4	**8.1**
medium 6	7	-2.9	**-2.6**	-2.1	0.3	**0.8**	-6.1	**-4.3**	-3.1	1.2	**3.0**
medium 7	9	-3.4	**-2.9**	-2.1	0.5	**1.3**	-7.2	**-4.8**	-1.1	2.3	**6.1**
medium 8	9	-3.3	**-2.6**	-2.0	0.4	**1.2**	-7.5	**-3.9**	-1.7	2.0	**5.8**
small 2	2	-0.9	**-0.7**	-0.5	0.3	**0.5**	-2.5	**-1.8**	-1.1	1.0	**1.4**
small 3	3	-2.8	**-2.6**	-2.4	0.2	**0.4**	-7.4	**-6.6**	-6.1	0.7	**1.3**
small 4	3	-3.3	**-3.3**	-3.2	0.1	**0.1**	-4.2	**-3.4**	-2.4	0.9	**1.8**
small 5	3	-0.7	**-0.1**	1.0	1.0	**1.7**	-4.6	**-4.1**	-3.0	1.0	**1.7**
small 6	3	-3.3	**-2.7**	-2.4	0.5	**0.9**	-3.5	**-3.0**	-2.6	0.5	**1.0**

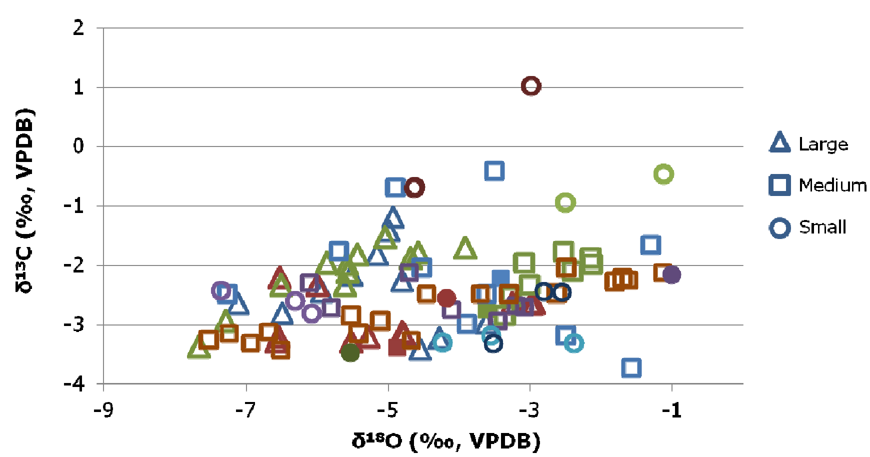

Figure 6.1 Stable isotope values for *Melanoides tuberculata* shell carbonates from a coquina at White Lake remnant Bir Tarfawi.

Table 6.2 Summary of Oxygen and Carbon Stable Isotope Compositions for a shallow area of Bir Tarfawi ca. 125 ka Remnant *Melanoides tuberculata* shells.

Sample	Number of samples	$\delta^{13}C$ (‰, VPDB)					$\delta^{18}O$ (‰, VPDB)				
		min	mean	max	std	range	min	mean	max	std	range
Overall	30	-5.6	**-3.3**	-2.1	0.9	**3.5**	-6.0	**-3.2**	0.5	1.5	**6.5**
Subsampled Shell											
large	8	-3.8	**-3.0**	-2.5	0.5	**1.3**	-3.6	**-2.6**	-0.5	1.1	**3.1**
medium 1	6	-3.3	**-2.7**	-2.1	0.5	**1.1**	-6.0	**-4.8**	-3.5	1.0	**2.5**
medium 2	7	-3.6	**-3.1**	-2.7	0.3	**0.9**	-5.4	**-2.9**	0.5	2.1	**5.9**
small 1	5	-5.6	**-5.0**	-4.1	0.6	**1.5**	-4.1	**-3.1**	-1.4	1.1	**2.7**
small 2	4	-3.3	**-2.9**	-2.5	0.4	**0.8**	-3.6	**-2.9**	-2.0	0.8	**1.6**

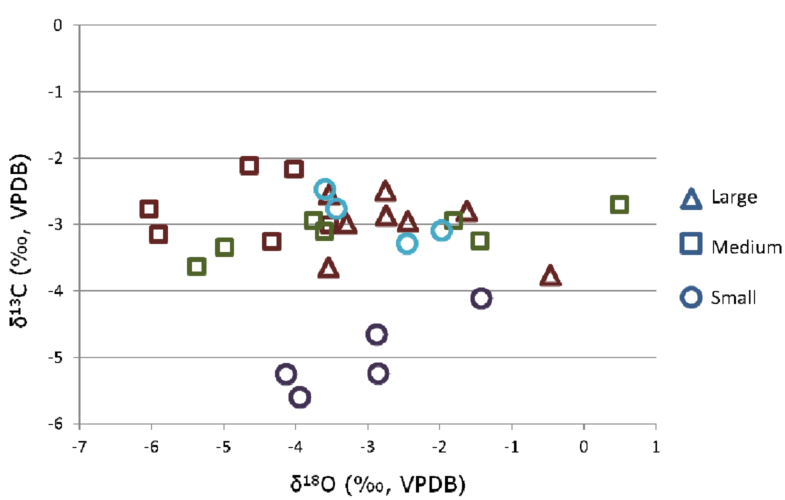

Figure 6.2 Carbon and oxygen stable isotope values for Melanoides tuberculata shell carbonates from a shallow area of Bir Tarfawi ca. 125 ka remnant.

Table 6.3 Summary of Oxygen and Carbon Stable Isotope Compositions for trench 7/74, deep area of Bir Tarfawi ca. 125 ka Remnant *Melanoides tuberculata* shells.

Sample	Number of samples	$\delta^{13}C$ (‰, VPDB)					$\delta^{18}O$ (‰, VPDB)				
		min	**mean**	max	std	**range**	min	**mean**	max	std	**range**
Overall	48	-6.2	**-3.7**	0.2	1.5	**6.4**	-8.0	**-5.1**	-0.5	1.6	**7.5**
Whole shell											
small	2	-4.2	**-3.5**	-2.8	1.0	**1.5**	-5.8	**-5.3**	-4.7	0.8	**1.1**
Subsampled shell											
large	5	-4.2	**-4.0**	-3.9	0.1	**0.3**	-7.7	**-5.6**	-4.5	1.4	**3.3**
medium 1	4	-3.2	**-1.7**	-0.4	1.2	**2.8**	-5.6	**-3.4**	-0.5	2.3	**5.1**
medium 2	4	-5.0	**-4.6**	-3.7	0.6	**1.4**	-5.7	**-5.3**	-4.2	0.7	**1.6**
medium 3	3	-4.6	**-4.3**	-4.0	0.3	**0.6**	-7.6	**-6.6**	-5.9	0.9	**1.7**
medium 4	3	-3.2	**-2.7**	-2.5	0.4	**0.7**	-4.3	**-4.1**	-3.6	0.4	**0.8**
medium 5	3	-5.6	**-5.4**	-5.1	0.3	**0.5**	-6.9	**-5.9**	-5.5	0.9	**1.5**
medium 6	5	-6.2	**-5.5**	-5.1	0.5	**1.1**	-8.0	**-6.7**	-5.3	1.3	**2.7**
small 1	2	-4.6	**-4.1**	-3.6	0.7	**1.0**	-6.6	**-5.3**	-3.9	1.9	**2.7**
small 2	2	0.0	**0.1**	0.2	0.1	**0.2**	-3.5	**-2.4**	-1.3	1.6	**2.2**
small 3	3	-4.1	**-3.7**	-2.9	0.7	**1.2**	-4.5	**-4.0**	-3.4	0.6	**1.1**
small 4	3	-5.2	**-4.9**	-4.5	0.4	**0.7**	-6.6	**-6.5**	-6.3	0.1	**0.3**
small 5	3	-2.9	**-2.6**	-2.4	0.3	**0.5**	-3.9	**-3.8**	-3.7	0.1	**0.2**
small 6	2	-3.6	**-3.3**	-3.1	0.3	**0.5**	-6.3	**-4.9**	-3.4	2.1	**2.9**
small 7	3	-3.0	**-2.8**	-2.6	0.2	**0.4**	-5.8	**-4.6**	-3.4	1.2	**2.4**

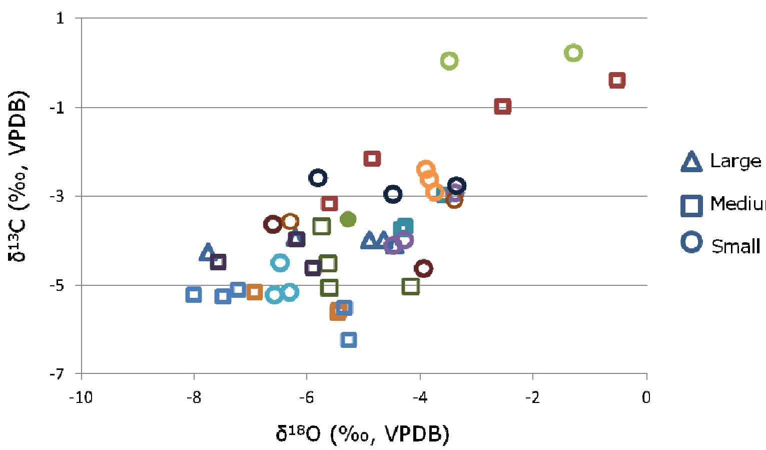

Figure 6.3 Carbon and oxygen stable isotope values for *Melanoides tuberculata* shell carbonates from deep area, trench 7/74 of Bir Tarfawi ca. 125 ka remnant.

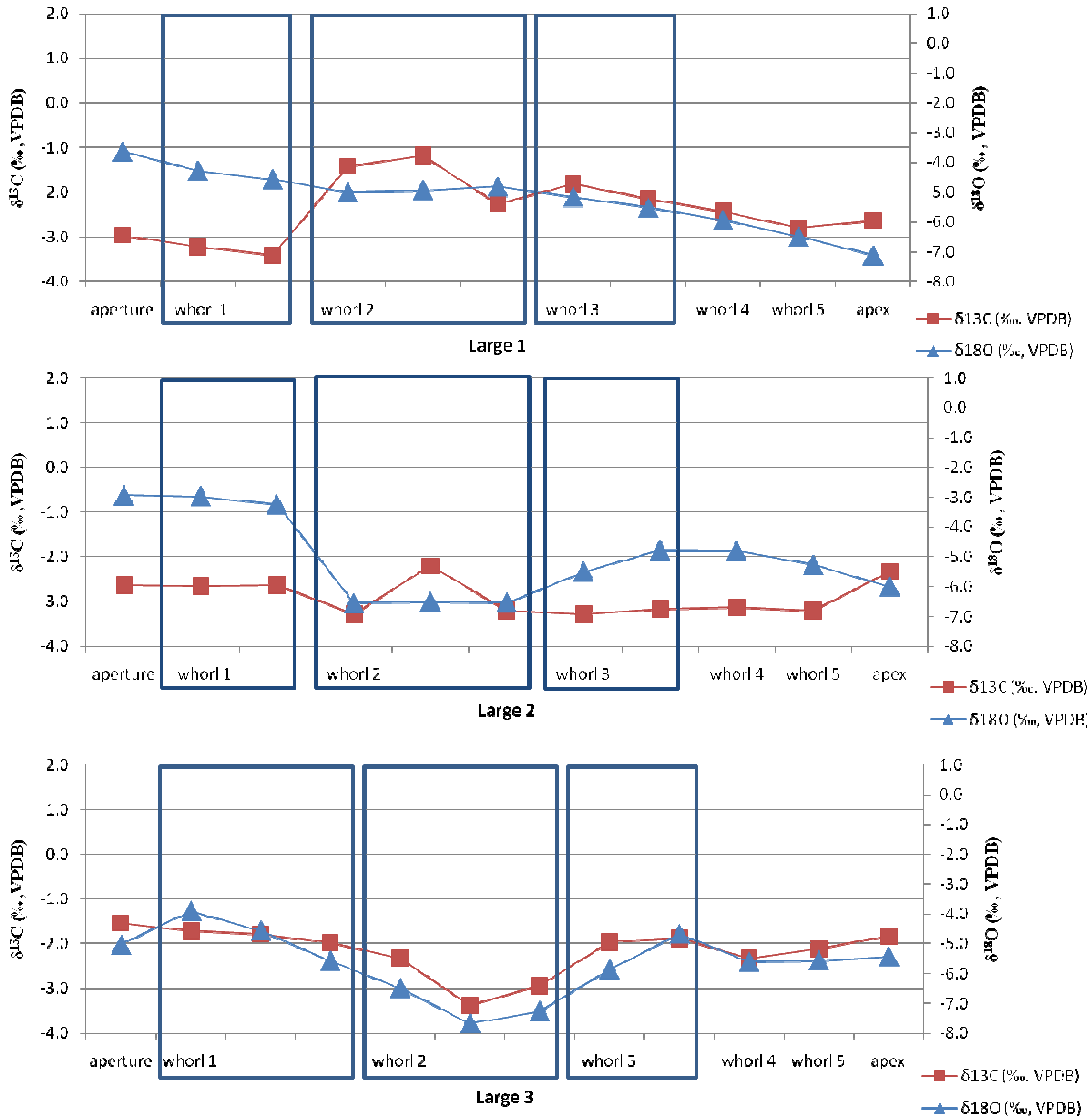

Figure 6.4 $\delta^{13}C$ and $\delta^{18}O$ values along the length of large (1.9 and 2.8 cm) ***Melanoides tuberculata*** **shells from White Lake remnant.**

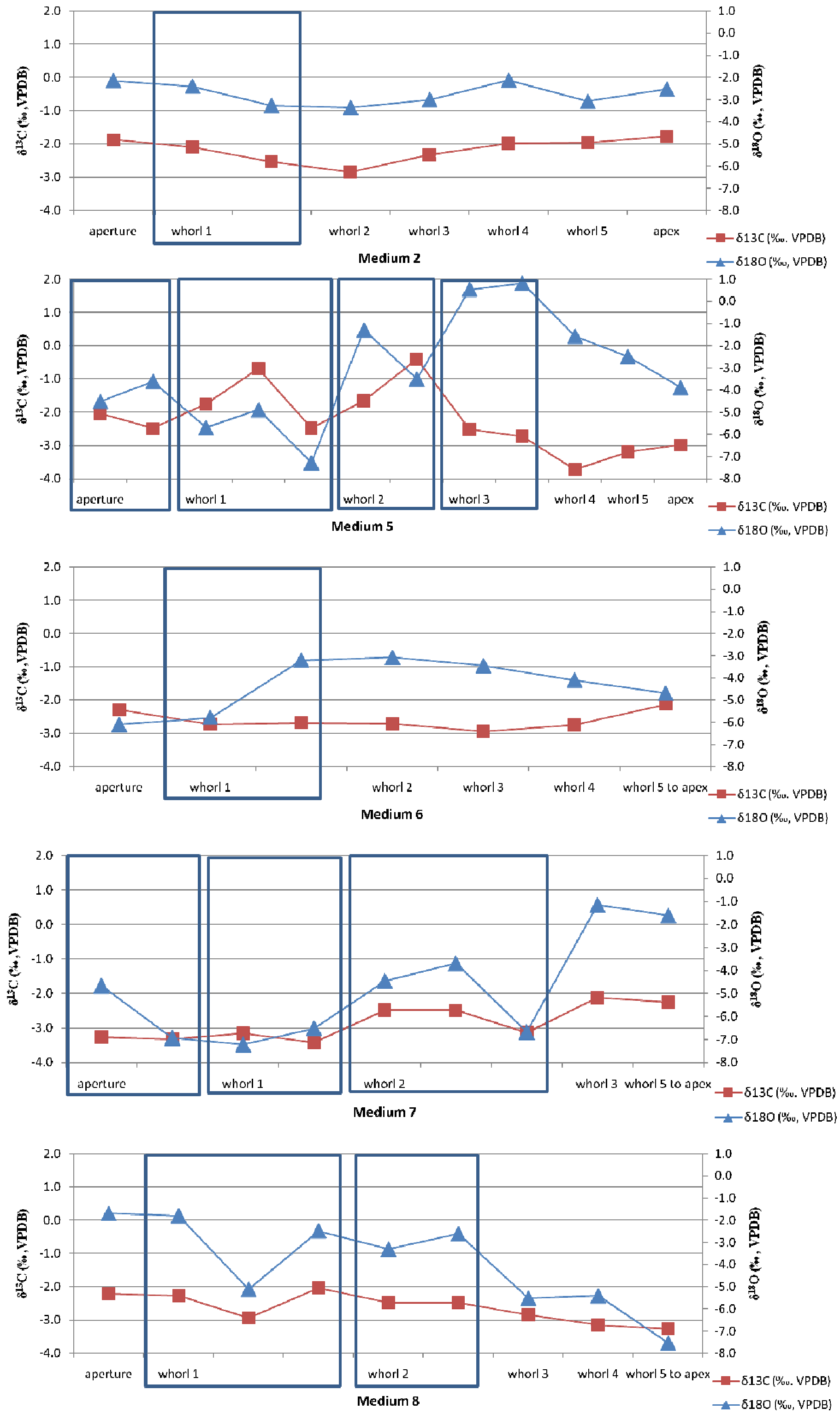

Figure 6.5 $\delta^{13}C$ and $\delta^{18}O$ values along the length of medium (1.5 to 1.8 cm) ***Melanoides tuberculata*** **shells from White Lake remnant.**

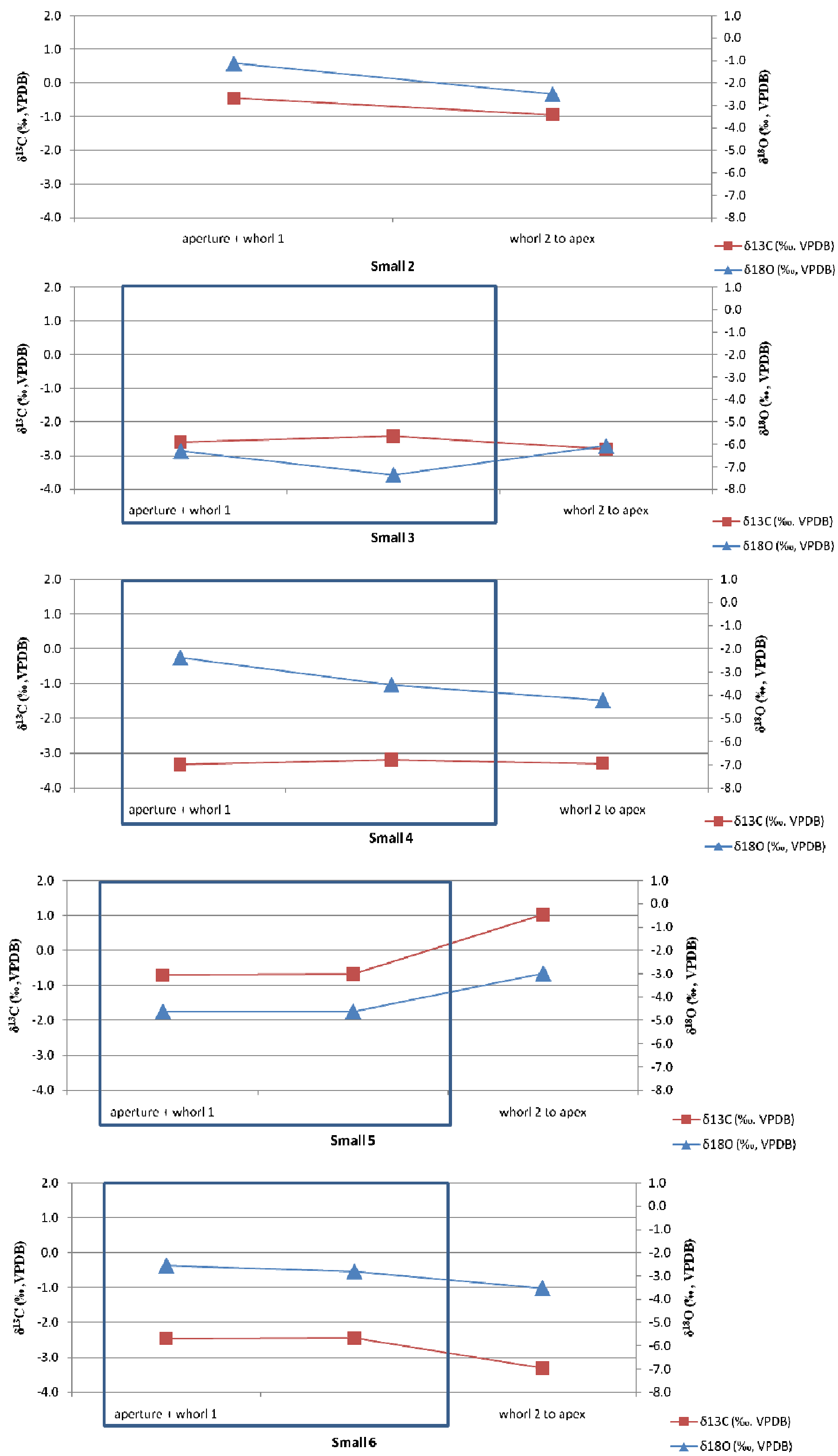

Figure 6.6 $\delta^{13}C$ and $\delta^{18}O$ values along the length of small (1.0 to 1.2 cm) ***Melanoides tuberculata*** **shells from White Lake remnant.**

for $\delta^{18}O$ values from shallow and deep *Melanoides* shell carbonates were -6.0‰ to 0.5‰ and -8.0‰ to -0.5‰, respectively.

Shanahan et al., (2005) suggests the *Melanoides tuberculata* shell carbonate $\delta^{18}O$ value ranges below 0.5‰ are likely due to vital effects. Most White Lake and Bir Tarfawi ca. 125 ka remnants gastropod shell carbonate $\delta^{18}O$ values had standard deviations and ranges well above 0.5‰ and therefore the variations can be interpreted as environmental changes. A similar specific range for $\delta^{13}C$ vital effect values was not determined by Shanahan et al., (2005). However, the gill breathing gastropods such as *M. tuberculata* were determined to precipitate carbonate with $\delta^{13}C$ values within expected equilibrium values (Shanahan et al., 2005).

Hypothesis: Seasonality

The null hypothesis for seasonality is that the $\delta^{18}O$ values of the White Lake remnant *Melanoides tuberculata* shell carbonate did not vary along the length of the shell. Large gastropods (>1.9 mm long) had oxygen isotope ranges of more than 3.5‰ (Table 6.1, Figure 6.1 and Appendix A). The large 1 gastropod had monotonically increasing $\delta^{18}O$ values between the apex and the aperture (Figure 6.4). Large 2 gastropod had low oxygen isotope values at two points along the shell, the apex and the penultimate whorl. The lowest $\delta^{18}O$ value for large 3 gastropod was in whorl 2. The medium subsampled gastropods had $\delta^{18}O$ ranges from 1.1‰ to 8.1‰ and $\delta^{13}C$ ranges from 0.8‰ to 3.3‰ (Table 6.1, Figures 6.1 and 6.5). Small shells had only two or three samples along the length and had $\delta^{13}C$ and $\delta^{18}O$ ranges 0.1‰ to 1.7‰ and 1.0‰ to 1.8‰, respectively (Table 6.1, Figures 6.1 and 6.6). Three small shells had $\delta^{13}C$ ranges of less than 0.5‰; it is likely these values were vital effects rather than indicative of environmental changes.

Table 6.4 Oxygen isotope ranges and carbon isotope means from Quaternary African lake gastropod shell carbonate (adapted from Kieniewicz, 2007).

Reference	Location		Age	Environment	range $\delta^{18}O$	mean $\delta^{13}C$	Species
Abell, 1985	Lake Malawi		modern	lake	1.2	-0.8	*Melanoides tuberculata*
Abell & Hoelzmann, 2000	Dry Selima P267	Sudan	5.57-8.82 ka	shallow lake	5.3	-8.2	*Biomphalaria pfeifferi*
					1	-6	*Melanoides tuberculata*
					0.6	-3.7	
					1.7	-3.8	
					1.3	-5.2	
					0.2	-8.3	
Abell & Hoelzmann, 2000	West Nubian Paleolake P195	Sudan	9.4-7.390 ka	shallow lake	4	-2.3	*Melanoides tuberculata*
					3	-1.4	
					1.8	-0.5	
					2.1	-1.8	
					1.8	-0.1	
					2.2	-4.6	
					2.5	-3.9	
Abell & Hoelzmann, 2000	Wadi Howar	Sudan	9.4-5.5 ka	pools in wadi	9.4	0.9	*Melanoides tuberculata*
					7.7	-6.8	
Abell & Nyamweru, 1988	Chalbi Basin	Kenya	11 ka	lake	1	-6	*Melanoides tuberculata*
Abell & Nyamweru, 1988	Olturot	Kenya	10.4 ka	small lake	3		*Melanoides tuberculata*
					2.8		
					1.7		
					2.3		
Abell & Nyamweru, 1988	Buffalo Springs	Kenya	modern	springs	0.75		*Melanoides tuberculata*
Abell & Williams, 1989	Lake Besaka	Ethiopia	15-20 ka	lake	1	-3.3	*Cleopatra bulmoides*
			11-9 ka		1.5	-3.3	*Melanoides tuberculata*
			11.2 ka		3.1	-4.7	
Abell & Williams, 1989	Aladi Springs	Ethiopia	11.07 ka	springs	1	-4.5	*Melanoides tuberculata*
Abell & Williams, 1989	Erer Springs	Ethiopia	6.67 ka	springs	1	-4.3	*Melanoides tuberculata*
Abell & Williams, 1989	Lake Lyadu	Ethiopia	modern	lake	3.7	-3.6	*Bulinus sp.*
					4	1.9	*Cleopatra bulmoides*
					6.1	-2.3	*Melanoides tuberculata*
Leng et al., 1999	Lake Awasa	Ethiopia	modern	river fed lake	0.3	-4.8	*Melanoides tuberculata*
Leng et al., 1999	Lake Tilo	Ethiopia	modern	crater lake	1.9	-6.5	*Melanoides tuberculata*
Smith et al., 2004	Kharga Oasis	Egypt	~125 ka	lake	0.6		*Melanoides tuberculata*
Kieniewicz, 2007					0.6	-3.91	

The $\delta^{13}C$ and $\delta^{18}O$ along shell values did not have consistent variation. That is, the aperture of each individual did not always have $\delta^{18}O$ enriched values. It was not possible to associate *Melanoides* beginning or end-of-life conditions with a specific season. *Melanoides* secreted shell throughout life (Abell, 1985; Pointier et al., 1992). Young are produced every two weeks after sexual maturity and not constrained to a specific season as long as conditions for active growth are present (Livshits and Fishelson, 1983; Pointier et al., 1992). While there is uncertainty about the life span of *Melanoides*, the largest animals could have survived through at least one year and likely two (Abell, 1985; Pointier et al., 1992; Livshits and Fishelson, 1983). Therefore, it was not possible to associate the beginning-of-life conditions within a specific season.

Gastropod shell carbonates from Quaternary stable African lakes (Table 6.4) had oxygen isotope composition ranges of less than 2‰. This low "along shell" range is thought indicate short-term stability in these lakes (Abell, 1985; Abell and Nyamweru, 1988; Abell and Williams, 1989; Leng et al., 1999; Kieniewicz and Smith, 2007; Smith et al., 2004). Large changes in the oxygen isotope composition of gastropod shell carbonates are common for seasonal or ephemeral water bodies (Abell and Nyamweru, 1988). Gastropod populations in Lake Lyadu, Ethiopia, a small modern lake that undergoes highly evaporative seasonal conditions, exhibited shell carbonate oxygen isotope ranges above 3‰ (Table 6.4 and Abell and Williams, 1989). Other lakes, including P267 Dry Selima (Sudan), West Nubian Paleolake (P195) (Sudan), and Olturot (Kenya), had broad along-shell oxygen isotope ranges interpreted as a seasonally variable environment based on temperature and/or evaporative conditions (Abell and Hoelzmann, 2000; Abell and Nyamweru, 1988). An extreme condition in the northern Sudan is Wadi Howar with gastropod shell $\delta^{18}O$ ranges greater than 7‰, the result of severe evaporation since water in the wadi may not have been sustained from season to season (Abell and Hoelzmann, 2000).

Medium and large *Melanoides* shell carbonates from the shallow White Lake remnant had $\delta^{18}O$ ranges above 3‰ (with one exception) (Table 6.1). These ranges were similar to $\delta^{18}O$ shell carbonate ranges for Lake Lyadu with its highly evaporative seasonal conditions. White Lake remnant small shell $\delta^{18}O$ ranges of 1.8‰ or less indicated stable lake conditions during the shorter life span of these gastropods (Table 6.1).

Carbon isotopic composition was controlled by photosynthesis-respiration processes, input water, and exchange with atmospheric CO_2 (McKenzie, 1993). The photosynthesis-respiration processes are maximized when temperature and light are optimized. Variations in the $\delta^{13}C$ values for *Melanoides* shell carbonates reflect these potentially cyclic processes.

Ostracods were also collected from White Lake remnant trench 10/86. Ostracod trace element analyses suggested that the lake water in the area of trench 10/86 was shallow and subject to temperature fluctuations (Deckker and Williams, 1993). The large $\delta^{18}O$ carbonate variations along the shell of *Melanoides tuberculata* from 10/86 White Lake remnant is consistent with shallow water sensitive to enriched oxygen composition moisture additions, temperature changes, and/or evaporation. These large $\delta^{18}O$ carbonate ranges imply that the seasonality hypothesis should be rejected. Rejection of the seasonality hypothesis supports the idea that variation along the *Melanoides* shell can serve as a proxy for seasonal changes at the White Lake.

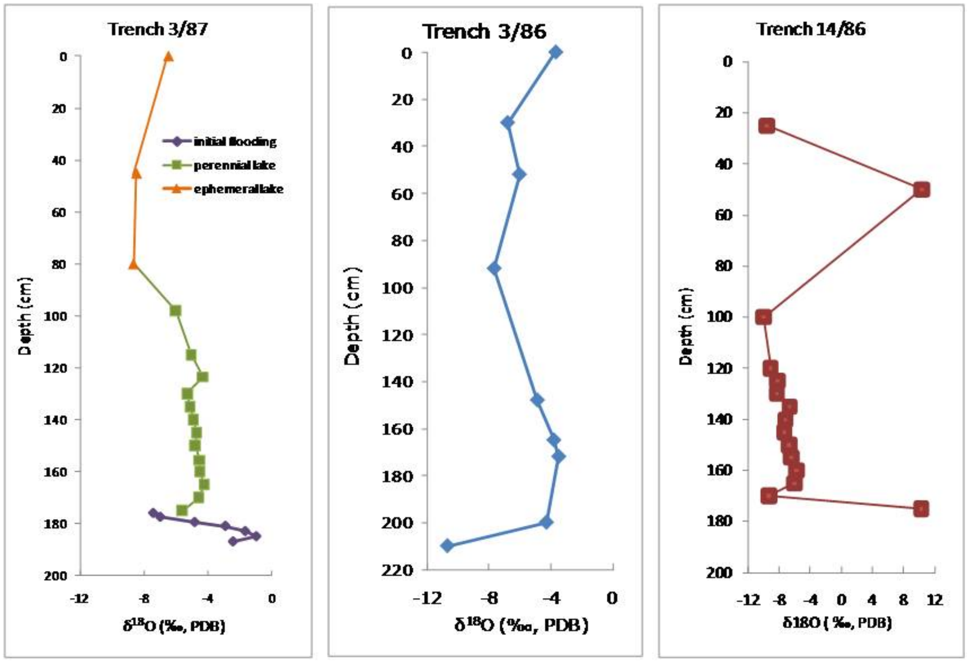

Figure 6.7 Depth versus sediment carbonate oxygen isotope compositions for Bir Tarfawi Pleistocene stratigraphic sequences Bir Tarfawi ca. 125 ka trenches 3/87 and 3/86, and White Lake trench 14/86 (McKenzie, 1993).

Hypothesis: Source Moisture

The null hypothesis for source moisture was that the stable oxygen isotope composition in *Melanoides tuberculata* shell carbonate from the White Lake remnant shallow area was isotopically consistent with compositions from source waters. The potential source waters included Atlantic Ocean monsoon, Indian Ocean monsoon, Mediterranean precipitation, Nubian Aquifer, and Nile flooding.

White Lake Sediment Carbonates

McKenzie (1993) described three lake stages (initial flooding, perennial lake, and ephemeral lake) based on sediment carbonate oxygen isotope composition, within the Bir Tarfawi pluvial sequence of trench 3/87 northwest of the BT-14 main trench (Figure 6.7). $\delta^{18}O$ values from trench 3/87 dating to ca. 125 ka were considered to be consistent with

Atlantic-sourced precipitation (-11‰ VSMOW) and subsequent evaporation resulting in a mean perennial lake sediment carbonate value of -4.9‰ (Figure 6.8 and McKenzie, 1993; Schwarz and Morawska, 1993). A similar pattern is noted in trench 3/86 sediment carbonate (McKenzie, 1993). White Lake remnant sediment carbonate (<130 ka) $\delta^{18}O$ values also exhibit a three-stage lake sequence (Figure 6.7), suggesting similar processes may have occurred during both wet periods (McKenzie, 1993; Schwarz and Morawska, 1993). However, the mean $\delta^{18}O$ value for the White Lake remnant sediment carbonate was more depleted (-7.0‰) than the ca. 125 ka remnant trenches 3/87 and 3/86 values (-4.9 ‰, -5.0‰). The difference in $\delta^{18}O$ values was perhaps due to warmer temperatures during the later pluvial phase, causing additional evaporation resulting in more enriched sediments carbonates (Figure 6.7 and Kim and O'Neil, 1997). The Bir Tarfawi remnants (Figure 6.7) had low oxygen isotope values during initial flooding and ephemeral phases, which is consistent with depleted Atlantic Ocean or deep aquifer source moisture (McKenzie, 1993). During these initial and final phases, the water in the lake basin may be more sensitive to temperature changes and source precipitation isotopic compositions due to lower water volumes and short residence times (Leng and Marshall, 2004).

With Saharan wet phases ascribed to the northward movement of the ITCZ and accompanying Atlantic monsoonal moisture, the Sudan received increased moisture at least in some pluvial cycles (Crombie et al., 1997; Ghoneim and El-Baz, 2007; Gossel et al., 2004; Hoelzmann et al., 2000; McKenzie, 1993; Sultan et al., 1997; Tisserand et al., 2009). In the Holocene, a number of lakes existed in northern Sudan including Dry Selima (P267) and the West Nubian Paleolake (P195) (Figure 2.1 and Abell et al., 1996). During this time, no large lakes developed at Bir Tarfawi, possibly due to a less intense wet phase during the Holocene than in the Pleistocene. Figure 6.8 shows that the mean $\delta^{18}O$ sediment carbonate values

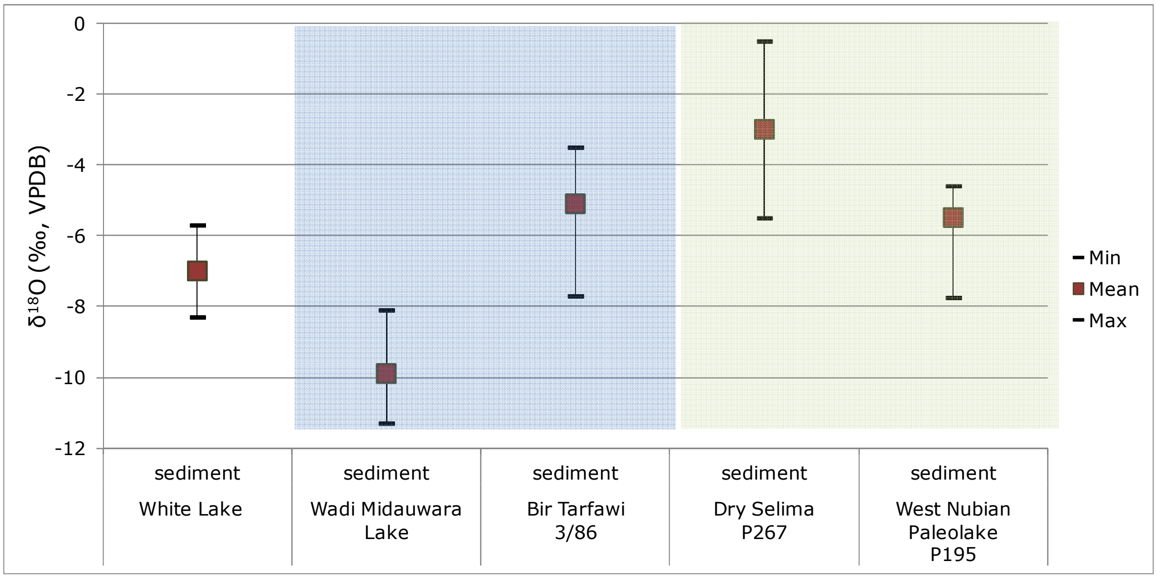

Figure 6.8 Sediment carbonate values from three Saharan wet phases >130 ka (White Lake), ca. 125 ka (Lake Midauwara, Bir Tarfawi) and Holocene (Dry Selima and West Nubian Paleolake) (Kieniewicz, 2007; Bradbury and Hill, 2008; Abell et al., 1996; McKenzie, 1993).

for remnants of the ca. 125 ka Bir Tarfawi perennial lake and the Holocene P195 Sudanese lake remnants were very similar. This similarity may be due to moisture delivery from the same source(s) resulting from the proximity of the ITCZ to each of the lake remnant regions during the Pleistocene and Holocene wet phases respectively. The hypothesized northward extent of the ITCZ during the Pleistocene was sufficient to bring rain bands locally over Bir Tarfawi and to create perennial lakes. During the Holocene, the ITCZ and associated rain bands likely did not extend as far north, creating perennial lakes in northern Sudan and seasonal playa lakes at Nabta, Egypt (de Noblet et al., 1996; Rossignol-Strick and Paterne, 1999; Wendorf and Schild, 1998). The large range of $\delta^{18}O$ values for Dry Selima may be due to the lake's position at the edge of the precipitation band during the Holocene wet phase (Abell et al., 1996).

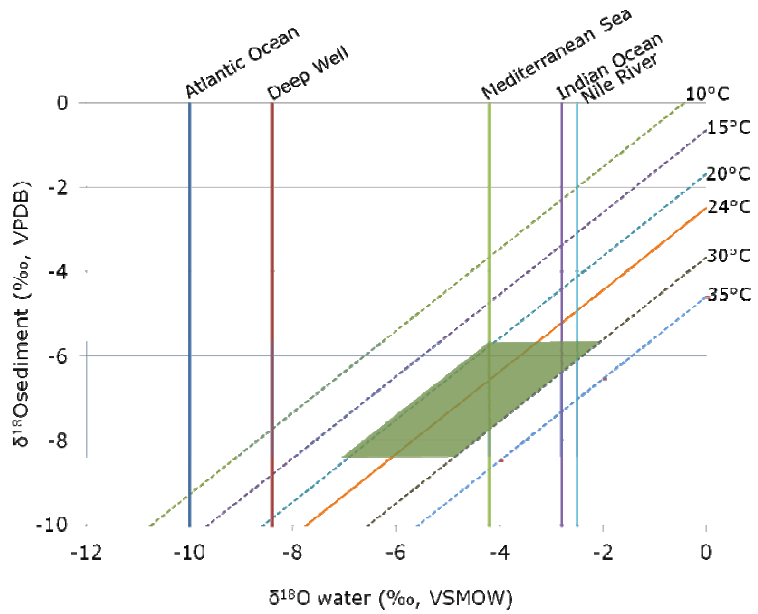

Figure 6.9 Equilibrium shell carbonates precipitate oxygen isotope values from potential source waters over a range of temperatures (Kim and O'Neil, 1997). Modern Nile River oxygen isotope values from Khartoum (Kebede et al., 2005). The red line current average temperature of 24°C (Vose et al., 1992). The blue box range of $\delta^{18}O$ values for White Lake remnant sediment carbonate. The green area glacial and interglacial temperatures proposed by Global Climate Models (de Noblet et al., 1996).

The earlier White Lake remnant with perennial sediment carbonate average $\delta^{18}O$ values ~2‰ lower than later Bir Tarfawi and P195 Sudan lake remnants suggested slightly different conditions for lake development such as more isotopically depleted source waters, lower temperatures, or more intense rainfalls. Atlantic-sourced precipitation represents the most depleted water source (Figure 6.9) for precipitation in the eastern Sahara, since the distance between the source and local precipitation was greatest. This continental effect resulted in depleted oxygen isotope values with increasing distance from the source (Dansgaard, 1964; Joseph et al., 1992). However, the White Lake and Bir Tarfawi ca. 125 ka remnants were in essentially the same location; therefore, isotopic compositions of precipitation based on continental effect should be similar for both remnants, assuming the source waters had the same temperatures.

A 2‰ difference in $\delta^{18}O$ values between the ca. 125 ka Bir Tarfawi, Holocene Sudanese and White Lake remnants sediment carbonate indicated an approximately 8°C temperature difference in the equilibrium carbonate precipitation from Atlantic-sourced moisture (Kim and O'Neil, 1997). Global Climate Models suggest surface air temperatures 3° to 7° warmer during the last interglacial (de Noblet et al., 1996), so an 8° temperature difference between wet phases may be possible. An additional source of depleted source waters were rapid, high volume rainfall that reduced the opportunity for evaporation during rain events, maintaining an oxygen isotope composition close to condensation values and, therefore, very highly depleted (amount effect, Dansgaard, 1964). Initial heavy monsoonal precipitation at White Lake could account for the lower $\delta^{18}O$ remnant sediment carbonate values during this wet period. Likely, then, temperature and/or amount effect contributed to the differences in mean $\delta^{18}O$ values between the wet phases represented by the White Lake, Bir Tarfawi ca. 125 ka and P195 lake remnants.

The Wadi Midauwara Lake remnant near Kharga (ca. 125 ka, Figure 6.8) has the most depleted sediment carbonate $\delta^{18}O$ values compared with stratigraphic sequences at Bir Tarfawi and northern Sudan (Kieniewicz and Smith, 2007). Kieniewicz and Smith (2007) considered the Wadi Midauwara remnant $\delta^{18}O$ values inconsistent with groundwater discharge or Atlantic-sourced moisture since an equilibrium carbonate-water precipitation temperature of 9°C would be required to precipitate carbonates with oxygen isotope values from those sources. This temperature was outside the predicted decrease of 4±2°C for glacial periods from Global Climate Models for the region (Kieniewicz, 2007; Liu et al., 2009; Bonnefille et al., 1990). However, the Global Network of Isotopes in Precipitation

(IAEA, 2006) database has documented daily temperatures as low as 14°C at Kharga, 12°C at Aswan, and 11.5°C on the Mediterranean coast at Sidi Barrani, but Khartoum has no temperature recorded below 20.5°C.

It is not clear from the GNIP database how sustained these temperatures were (IAEA, 2006). It would take sustained low air temperatures to produce lake temperatures of 9°C. Kieniewicz and Smith (2007) suggest that $\delta^{18}O$ values at Wadi Midauwara Lake would require an additional precipitation source enriched in ^{18}O to produce the measured sediment carbonate values.

Djara Cave located northwest of Lake Midauwara contains speleothems as well as plant and animal remains that indicate Pleistocene moisture reached the region from two directions (Brook et al., 2002; Kindermann et al., 2006). Speleothems have dates similar to those of the Bir Tarfawi 125 ka and White Lake remnants. Djara Cave speleothem $\delta^{18}O$ values, ranging from -14.2 to -10.4‰, were interpreted to result from westerly winter and summer monsoon precipitation (Brook et al., 2002).

Evaluating sustained low oxygen isotope values from *Etheria elliptica* shell from the eastern edge of Wadi Howar, Rodrigues et al. (2000) also considered an additional precipitation source. In this analysis of $\delta^{18}O$ from monthly shell growth, there appeared to be two rain periods. It was suggested that the pattern was similar to that of modern East Africa where there are multiple rain seasons, including Atlantic Ocean and Indian Ocean input (Rodrigues et al., 2000). If this is the case, Indian Ocean-sourced precipitation reached the Khartoum area during the early Holocene. Modern precipitation at Khartoum is confined to May through October with oxygen isotope values ranging from -8.3‰ to +6.8‰ (VSMOW) with no precipitation source noted (IAEA, 2006). The modern maximum northward range of the ITCZ is about 20°N just north of Khartoum (15°N), allowing

Atlantic-sourced monsoonal precipitation into the area between June and September (Kebede et al., 2005). For the *Etheria elliptica* early Holocene data, the secondary moisture source explains an extension of the rainy season beyond that expected for the Atlantic monsoon. Wadi Howar is a water channel rather than a lake, thereby requiring a continuous water source to maintain flow, although pools have been documented (Abell and Hoelzmann, 2000). Depleted $\delta^{18}O$ values evident throughout the life span of the analyzed *E. elliptica* appear to eliminate output from one of the lakes as a source, since water moving from the lake down the wadi would be expected to demonstrate evaporative enrichment in $\delta^{18}O$.

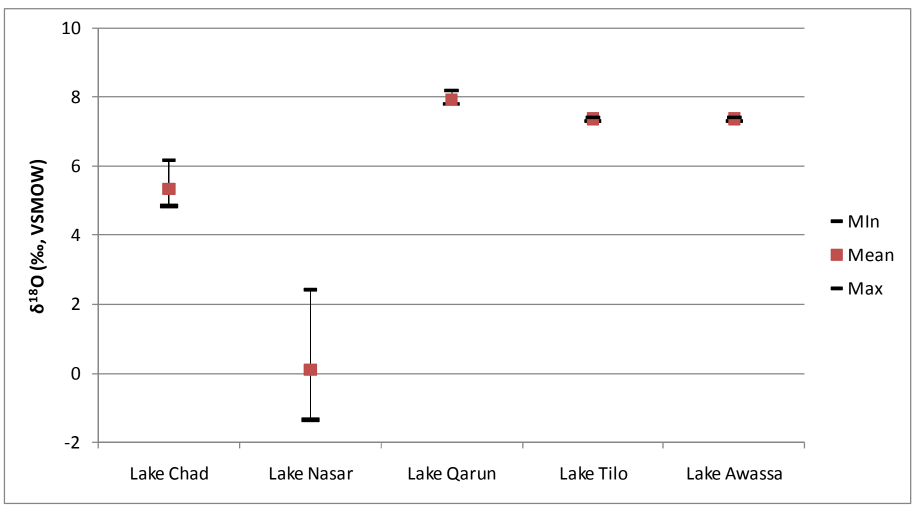

Figure 6.10 Surface water $\delta^{18}O$ values from multiple areas on modern lakes in Chad, Egypt (Nasser and Qarun) and Ethiopia (Tilo and Awassa) (Goni, 2006; Keatings et al., 2007; Aly et al., 1993; Leng et al., 1999).

Another potential water source in the region is the Nile River, suggested as the source of a paleolake covering the entire Kharga to Bir Tarfawi basin (Maxwell et al., 2010). The modern oxygen isotope value of the main Nile River at Khartoum is +2.51‰ (VSMOW), which would result in an equilibrium carbonate precipitate of -0.1‰ VPDB (at 24°C; Kim and O'Neil, 1997). The main water sources for the Nile River are the White and

Blue Niles, which originate in Lake Victoria and the Ethiopian Highlands respectively. Currently, the White Nile maintains the Nile throughout the year with high seasonal flow due to Atlantic-sourced monsoonal moisture. The White Nile watershed is located generally east of the Congo Air Boundary (CAB; Figure 2.1), which provides an opportunity for Indian Ocean moisture to penetrate the region during March through May (Levin et al., 2009). Speculation about the isotopic composition of the paleo-Nile limited by when the White and Blue Nile became source waters, the movement of the ITCZ and CAB with associated rain bands, as well as differences in climatic conditions, including temperature. Estimates of potential oxygen isotope values are also complicated by disagreement on the configuration and timing of the connection for Nile source waters (Maxwell et al., 2010; Rossignol-Strick and Paterne, 1999). It should be noted that Lake Midauwara and Bir Tarfawi ca. 125 ka lake would likely be segments of this large Nile-sourced lake. The difference in mean $\delta^{18}O$ values (-10‰ and -5‰ respectively) was significantly greater than noted in modern regional lakes (Figure 6.10).

Equilibrium sediment carbonate: oxygen isotope values for each of the proposed moisture sources at temperatures between 10 °C and 35 °C were shown in Figure 6.9. Global Climate Models predict a range of temperatures for the last interglacial suggesting equilibrium precipitated carbonates from Mediterranean and Indian Ocean sources were likely contributors to the Bir Tarfawi pluvial events neglecting evaporation (de Noblet et al., 1996; Prell and Kutzbach, 1987). Surface water at modern Lake Chad has a mean oxygen isotope value of 5.3‰ (VSMOW), from average precipitation of -4‰ (VSMOW) with a mean temperature ~ 30°C indicating a contribution from evaporation (Goni 2006; IAEA, 2006). Predicted interglacial temperatures were at or greater than modern conditions. Evaporation of lake water would then be likely during the Bir Tarfawi pluvial periods' maintaining the Atlantic monsoon as a potential moisture source (Bonnefille et al., 1990; de Noblet et al., 1996).

Perennial lake sediment carbonate $\delta^{18}O$ values do not rule out either Atlantic or additional sources for the Bir Tarfawi pluvial phases. Instead, oxygen stable isotope compositions from Djara Cave in north central Egypt and the eastern edge of Wadi Howar in Sudan suggest these areas experienced moisture sources from the westerlies and Indian Ocean respectively (Brook et al., 2002; Rodrigues et al., 2000). Kieniewicz and Smith (2007) argued that oxygen isotope values from Wadi Midauwara Lake (between Bir Tarfawi and Djara) are inconsistent with a single Atlantic-sourced monsoon. However, sediment carbonates represent relatively long-time period proxies. A short-term proxy consistent with seasonal changes, such as *Melanoides tuberculata,* may provide additional constraints to determine moisture sources.

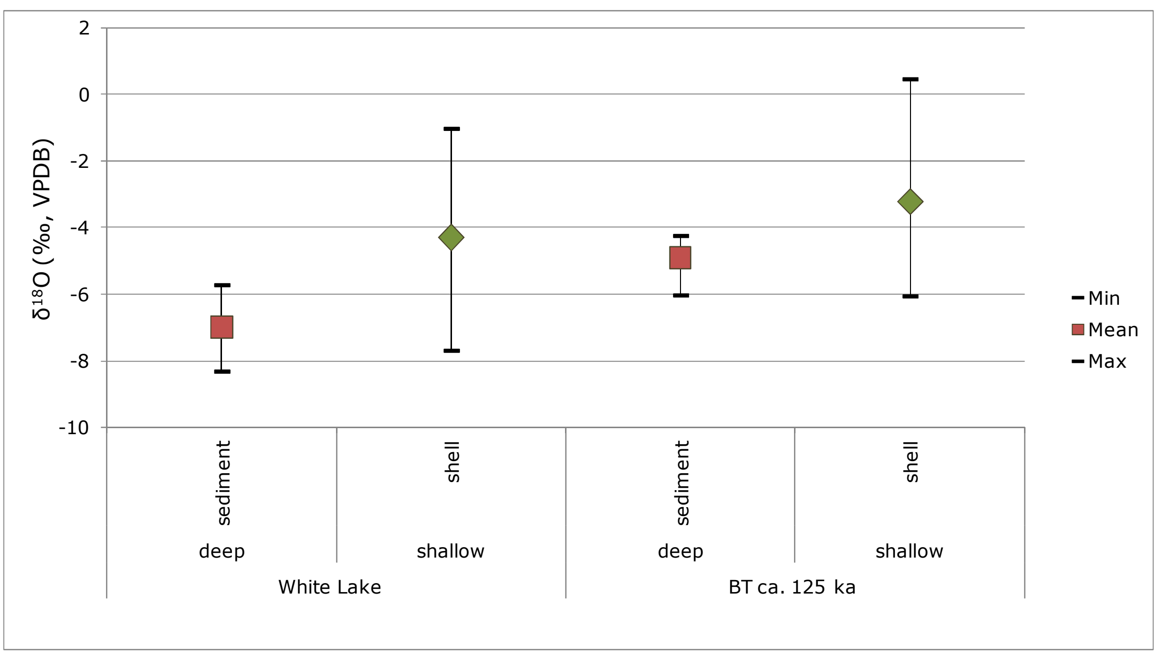

Figure 6.11 White Lake and Bir Tarfawi ca. 125 ka remnant deep water sediment carbonate and shallow water *Melanoides tuberculata* shell carbonate $\delta^{18}O$ values (McKenzie, 1993).

Melanoides tuberculata

Melanoides tuberculata shells reflect short-term environmental conditions since the shell was secreted continuously in equilibrium with surrounding waters over its life span of up to 2 years (Abell and Hoelzmann, 2000; Livshits and Fishelson, 1983; Pointier et al., 1992). The most depleted $\delta^{18}O$ isotopic compositions of the White Lake remnant large gastropod shells (-7.7‰) were well within the range of sediment carbonate $\delta^{18}O$ values (-8.3 to -5.7‰) (Figure 6.11). No White Lake remnant gastropod shell carbonates had $\delta^{18}O$ values more depleted than sediment carbonate values. 78% of the *Melanoides* shell carbonate $\delta^{18}O$ values were more enriched than the values for sediment carbonates. It should be noted that the *Melanoides* and the sediment are not from the same location in the lake remnant. The sediment was from trench 14/86 and represents a deeper water facies whereas the gastropods came from the more shallow area of trench 10/86.

Melanoides tuberculata shell carbonate $\delta^{18}O$ values from shallow area of Bir Tarfawi ca. 125 ka remnant (Figure 6.11, Appendix B) were measured to compare with deep lake sediment carbonate values obtained by McKenzie (1993). The ca. 125 ka remnant shallow area shell carbonate oxygen isotope values were not more depleted than the deep water sediment carbonate values (Figure 6.11). Seventy three percent of the ca. 125 ka shallow remnant *Melanoides* shell carbonate $\delta^{18}O$ values were more enriched than the deep water remnant sediment carbonate values. Shallow water *Melanoides* from both Bir Tarfawi pluvial phases had a significant number of $\delta^{18}O$ values, indicating enriched surrounding waters. High $\delta^{18}O$ lake water values were likely due to input of enriched moisture sources or evaporation.

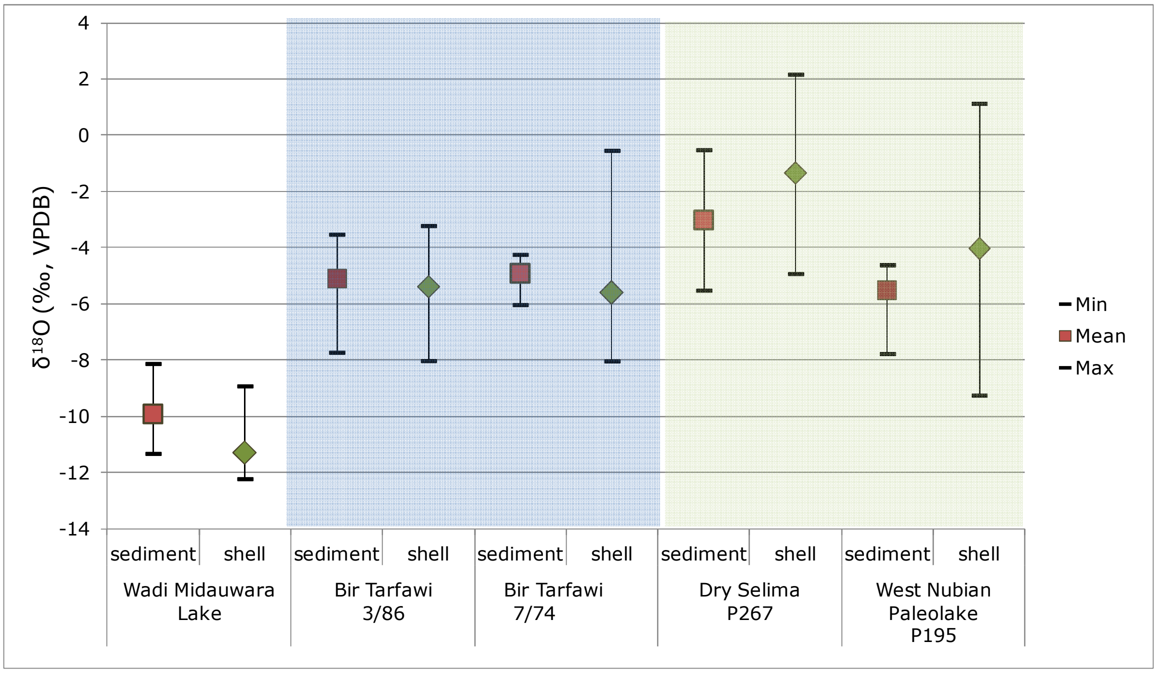

Figure 6.12 Deep water sediment and shell carbonate $\delta^{18}O$ values from Pleistocene and Holocene lake remnants of Bir Tarfawi and Sudan (Abell et al., 1996; Bradbury and Hill, 2008; Kieniewicz, 2007; McKenzie, 1993).

Deep water *Melanoides* shell carbonates (Figure 6.12) had 20% (P267), 27% (Bir Tarfawi ca. 125 ka remnant trench 7/74), 20% (trench 3/86 northwest of BT-14), and 55% (P195) of the $\delta^{18}O$ values more enriched than the coincident sediment carbonate values. None of the Wadi Midauwara shell carbonates were more enriched than those in the surrounding sediment. The conditions for Bir Tarfawi ca. 125 ka, P195, and P267 were considered stable freshwater bodies and the range in $\delta^{18}O$ values interpreted as a response to significant seasonality (Abell et al., 1996; McKenzie, 1993). Seasonality would result from precipitation constrained to certain months of the year and evaporation during intervening months.

The higher percentage of enriched values for shallow remnant *Melanoides* suggested the lake area represented by these proxies may have been more sensitive to the two key enrichment factors: enriched moisture sources and evaporation. These differences suggested a discontinuity between the lake water and the water surrounding the gastropods. Since perennial lake waters were well-mixed with respect to oxygen isotope compositions (Goni, 2006; Leng et al., 1999), the disparity between lake and gastropod oxygen isotope values suggested that the gastropods spent only part of their life span in water with oxygen isotopic composition similar to that of the main lake.

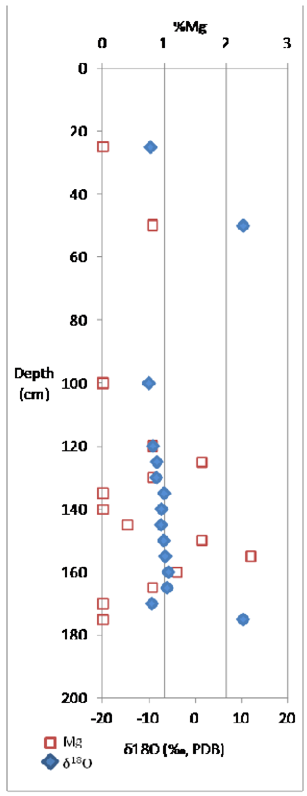

Figure 6.13 $\delta^{18}O$ values and %Mg from White Lake remnant (14/86) sediment carbonate (McKenzie, 1993).

The results of trace element analysis of in situ ostracods (Deckker and Williams, 1993) provides further evidence of significant environmental variation within the trench 10/86 stratigraphy. Temperature fluctuations that are likely the result of shallow water were noted in trench 10/86 ostracod analysis. This provided support for the presence of possible evaporative conditions (Deckker and Williams, 1993). Alternately, small pools would more readily reflect local precipitation from sources with high $\delta^{18}O$ values such as the Mediterranean (-4.2‰ VSMOW), Indian Ocean (-2.8‰ VSMOW) and potentially the Nile River discharge (> -3.3‰ VSMOW, based on modern values).

The lack of abrasion on the White Lake remnant *Melanoides* shell surfaces suggests that there was not a regular ebb and flow to the lake in the area where the coquina was deposited. Therefore, a single transgressive lake event may have occurred with the subsequent regression stranding the gastropods in a shallow pool, which ultimately was subject to evaporation. This could also explain the taphonomy of the *Melanoides* coquina (death assemblage).

The %Mg content of White Lake sediment carbonate is low (mean ~1) (Figure 6.13). Magnesium tends to substitute for calcium in carbonate as the salinity of the water increases (McKenzie, 1993; Eugster and Hardie, 1978). Therefore, magnesium can be considered a proxy for evaporation. The low levels of magnesium in the remnant sediment carbonates reaffirm White Lake as a fresh water body and implied that evaporation was not a key process within the deep lake. There were two excursions to slightly higher values. The first occurs at greater than 150 cm, corresponding to unit 2 in trench 14/86, which is situated below the unit associated with the coquina (Figure 5.1 and Hill, 1993). A second high value indicated fresh to brackish water at 139 cm or within unit 3 (McKenzie, 1993). Both of these values may indicate a regression of the lake. This excursion suggested that conditions may have existed to cause a lake regression resulting in evaporation and potential stranding of gastropods in a separate pool.

Either a separate pool or shallow waters would be more sensitive to the addition of source water with varying oxygen isotopic composition. Potential input sources were likely highly constrained within specific periods, resulting in seasonal signals noted within gastropod shells. Modern Mediterranean moisture affects northern Egypt from October through March (IAEA, 2006). Indian Ocean-sourced precipitation impinges on Ethiopian Highlands in April and March (Kebede et al., 2005). The modern rainy season across the Sahel encompasses June through September, and has both an Atlantic and Indian Ocean component (Joseph et al., 1992; Nicholson, 2008, 2009; Nicholson and Flohn, 1980).

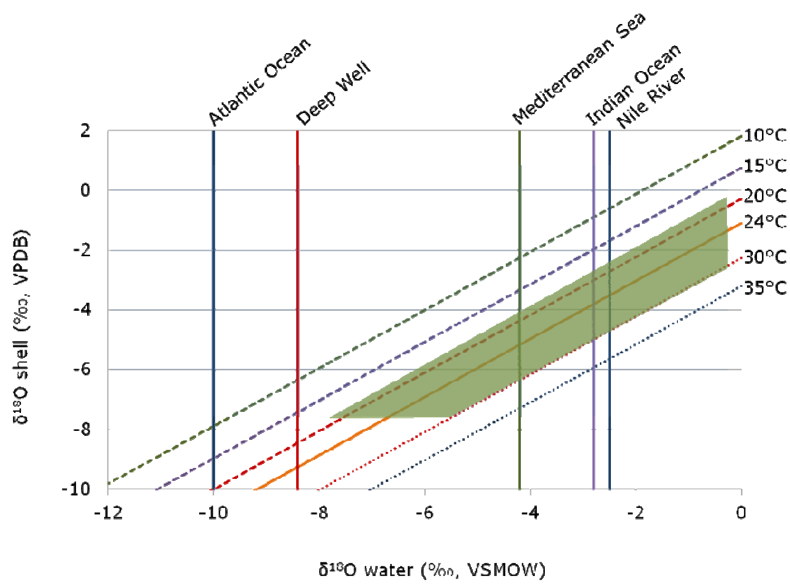

Figure 6.14 Equilibrium shell carbonates (aragonite) precipitate oxygen isotope values from potential source waters over a range of temperatures (Grossman and Ku, 1986). Modern Nile River oxygen isotope values from Khartoum (Kebede et al., 2005). The orange line current average temperature of 24°C (Vose et al., 1992). The blue box range of δ^{18}O values for the White Lake remnant shell carbonates. The green area glacial and interglacial temperatures proposed by Global Climate Models (de Noblet et al., 1996).

If there were insignificant evaporative effects, then both the Mediterranean (-4.2‰ VSMOW) and Indian Ocean (-2.85‰ VSMOW) sources are consistent with enriched White Lake sediment carbonate oxygen isotope values (Figure 6.9 and Kim and O'Neil, 1997).

However, precipitation at modern Lake Chad averages -4‰ (VMSOW), and the surface lake water mean is 5.3‰ (VSMOW) (Goni, 2006). Therefore, Atlantic-sourced moisture of -11‰ (VSMOW) could evaporate to form perennial White Lake water having values consistent with carbonate precipitation. White Lake sediment carbonates range from -8.3‰ to -5.7‰, and corresponding lake water at 24°C would need to be -5.9‰ (VMSOW) and -3.3‰ (VSMOW) respectively (Figure 6.9 and Kim and O'Neil, 1997). Both of these water values are within the difference range measured at modern Lake Chad. Also, a combination of Atlantic monsoon and enriched water sources (Indian Ocean or Mediterranean Sea) could produce the White Lake remnant sediment and shell carbonate $\delta^{18}O$ values.

Groundwater has been postulated as a continuous source for lakes within the Nubian aquifer (Gossel et al., 2004; Patterson et al., 2005; Robinson et al., 2007; Sonntag et al., 1978; Thorweihe, 1990). Deep aquifer water has substantially more depleted oxygen isotope compositions than modern precipitation, likely due to recharge during previous wet phases from Atlantic-sourced precipitation (Gossel et al., 2004; Patterson et al., 2005; Robinson et al., 2007; Sonntag et al., 1978; Thorweihe, 1990). Deep well water near Bir Tarfawi has a $\delta^{18}O$ value of -8.4‰ (VSMOW) (McKenzie, 1993). Equilibrium precipitation of sediment carbonate from this moisture source would need to occur at much cooler temperatures than expected during the interglacial (Figure 6.9 and Kim and O'Neil, 1997). Mixing of deep well water and precipitation, along with evaporation, would make aquifer water a viable source for the Bir Tarfawi remnants.

The high percentage of enriched $\delta^{18}O$ values for the White Lake remnant *Melanoides* shell carbonates with respect to the deep water sediment carbonate suggests the surrounding waters were sensitive to source water oxygen isotopic input or evaporation. Sensitivity to these enrichment conditions implied shallow water or a separate pool for the gastropods. No

White Lake shallow water remnant shell carbonates had $\delta^{18}O$ values more depleted than -7.7‰. For these values of shell carbonates to originate from Atlantic-sourced moisture, water temperatures would have been 10°C (Figure 6.14 and Grossman and Ku, 1986). This was unlikely, based on current understanding of temperatures during interglacials (de Noblet et al., 1996; Prell and Kutzbach, 1987). Deep well water at Bir Tarfawi has a $\delta^{18}O$ value of -8.4‰ which, based on equilibrium theory, is consistent with White Lake remnant sediment and shell carbonate values. An advantage of groundwater as a depleted source is that significant evaporation does not have to be invoked to explain sediment oxygen isotope values. Mediterranean Sea and Indian Ocean sources alone were also unlikely to produce the most depleted shell values, based on equilibrium precipitation (Figure 6.14 and Grossman and Ku, 1986). However, 77% of values above the sediment carbonate maximum are consistent with Mediterranean Sea and Indian Ocean sources. The source moisture hypothesis can be rejected since $\delta^{18}O$ values are not consistent with a single moisture source. The most parsimonious explanation of oxygen isotopic compositions in the White Lake remnant for both sediment and shell carbonates seems to be multiple moisture sources.

Summary

White Lake remnant (>130 ka) deep water sediment carbonate $\delta^{18}O$ values had the same three-phase pattern noted in the Bir Tarfawi (ca. 125 ka) remnant sediment carbonate (Figure 6.7 and McKenzie, 1993). Average $\delta^{18}O$ values from remnants of different ages imply variation in conditions potentially the result of difference in temperature or source moisture combinations. *Melanoides* shell carbonates from shallow waters of both ca. 125 ka and White Lake remnants are 70% or more enriched $\delta^{18}O$ values than deep water carbonates. These shell carbonate values are consistent with additional moisture sources other than the significantly depleted Atlantic monsoon. Oxygen isotopic composition changes along the gastropod shell support seasonal variation, either evaporation or additional enriched source moisture. It is likely that conditions during the White Lake and Bir Tarfawi ca. 125 ka wet periods were similar enough to enable utilizing fossil evidence from Bir Tarfawi ca. 125 ka remnant to extrapolate environmental conditions to the earlier pluvial period.

CHAPTER SEVEN: LANDSCAPE IMPLICATIONS

A freshwater lake remnant older than 130 ka with Middle Paleolithic lithics, exists at Bir Tarfawi in the Egyptian Western Desert (Schild and Wendorf, 1993; Schwarcz and Morawska, 1993). Pluvial sediments of the same age are present in the Sudan at Wadi Hussein (Szabo et al., 1995). North and east of Bir Tarfawi, spring-fed tufas develop near Kharga and Wadi Kukur respectively (Crombie et al., 1997; Sultan et al., 1997). Djara Cave, north of Kharga, contains speleothems of a similar date. All of these examples represent increased moisture in the usually arid Saharan desert. Speleothems in Djara Cave may have formed as the result of increased intensity in westerly and summer monsoonal moisture (Brook et al., 2002). Tufas may be the result of recharge of the Nubian aquifer, possibly from multiple sources (Crombie et al., 1997; Kieniewicz and Smith, 2007; Sultan et al., 1997; Smith et al., 2004). Lake formation in the Sudan and the Egyptian Western Desert are interpreted as indicating the northward movement of the ITCZ, allowing the Atlantic monsoon and other moisture sources to bring local precipitation into the region. The same moisture sources responsible for lake formation could have recharged the aquifer (Crombie et al., 1997; Ghoneim and El-Baz, 2007; Gossel et al., 2004; Hoelzmann et al., 2000; McKenzie, 1993; Sultan et al., 1997; Tisserand et al., 2009). These events were typical of a Saharan wet phase. The exact northern position of the ITCZ and the intensity of local precipitation may vary between phases, causing differences in the location of perennial lakes, but there is an overall consistent pattern.

Besides mollusks and ostracods no fossils were recovered at the White Lake remnant to provide additional information concerning the surrounding habitat. Similarities in the relationship between the three-phase pattern noted in sediment carbonate $\delta^{18}O$ values and the enriched shallow water shell carbonates suggest that White Lake and ca. 125 ka remnants had similar conditions. These similarities allow extrapolation of the fossil record from younger Bir Tarfawi sediments to older White Lake deposits. Bir Tarfawi ca. 125 ka vertebrate fossils indicate the immediate area abutting the younger lake was a reeded shallows surrounded by rich vegetation (Bocheński, 1993; Gautier, 1993; Kowalski, 1993). The main avian fossil component was comprised of cormorants who were ground nesters within reeds and partially submerged trees and shrubs. In addition, occupying this rich vegetative area near shore were numerous rodents such as shrew. The modern corollary to the rodent population at BT-14 is in Lower Omo, Ethiopia. Lower Omo is a combination of riverine trees and shrubs adjacent to acacia open shrub land (Kowalski, 1993). A similar mixture of high moisture vegetation with nearby low moisture grasses could therefore be expected in the vicinity of White Lake. In modern eastern and northern Africa, a water body supports significant vegetation in the immediate area, but there is a rapid transition to arid landscapes (Ashley et al., 2009). Sand grouse and snake (*Eryx*) fossils present at Bir Tarfawi ca. 125 ka attest to the arid landscape within the vicinity of the lake (Kowalski, 1993). The greater area surrounding the Bir Tarfawi ca. 125 ka lake was considered Sahelian or grassland with scattered trees and shrubs (Gautier, 1993; Wendorf et al., 1993b). Fossil evidence included birds (houbana, namaqua doves, and vultures) as well as large and small mammals (gazelles, giraffe, rhinoceros, and antelope), all adapted to the savanna (Bocheński, 1993; Gautier, 1993).

Scattered lake environments alone may not have provided sufficient resources to support Middle Paleolithic hominin movement since Bir Tarfawi and Wadi Hussein, Sudan were located 500 km apart. Additional resources were likely required for hominins to traverse this distance with no known means of carrying water (Basell, 2008). Even though these lakes represent a wet period in the eastern Sahara, the region overall was likely semi-arid. The question remains, what was the nature of the environment between Wadi Hussein and the White Lake?

Melanoides tuberculata from trench 10/86 at White Lake remnant representing shallow waters had $\delta^{18}O$ values indicating seasonality and multiple sources for local precipitation. A seasonal signal along the gastropod shell is likely the result of either differences in precipitation sources or significant evaporation or both. The magnesium content of sediment carbonates indicated that evaporation did not likely cause enriched shell carbonate oxygen isotope values. Multiple moisture sources, however, could provide seasonal variation. Modern moisture sources in the region are constrained to specific months. Mediterranean moisture occurs from October through March (IAEA, 2006). Indian Ocean-sourced precipitation falls between April and March (Kebede et al., 2005). The modern rainy season occurs across the Sahel from June through September with an Atlantic and Indian Ocean components (Joseph et al., 1992; Nicholson, 2008, 2009; Nicholson and Flohn, 1980). Multiple moisture sources improve the probability of resource availability across the savanna connecting wet phase lake environments.

Evidence from the Holocene wet phase indicated the type of landscape expected between the White Lake and Wadi Hussein. Small Holocene sites consisting of hearths and a few lithics were located on the sand sheet in areas such as small basins, depressions in the giant ripples or near rock outcrops (Close, 1990). Similar sites were scattered throughout the

Western Desert and the Selima Sand Sheet and were interpreted as rest stops for Neolithic cattle herders (Close, 1990). Because cattle require regular watering, sites within depressions and basins must have preserved water from recent rains, allowing successful crossing of the sand sheet (Close, 1990). Since the Saharan wet phases in the Pleistocene were more intense than those of the Holocene, as evidenced by the significantly higher number of medium to large mammals at Bir Tarfawi (ca. 125 ka) compared with Kiseiba (Holocene), similar surface water should have been available in the Pleistocene (Gautier, 1993). Pleistocene wet phase lithics found on the sand pan support this assumption (Olszewski et al., 2010).

While excavations at White Lake did not provide direct evidence of fauna other than *Melanoides tuberculata* and a few ostracods, similarity to the younger Bir Tarfawi deposits suggests that the near lake area would have been rich in birds and gazelle. Analogs for aggregation of plant and animal life surrounding water can be seen in modern Africa, including areas around the Nile River and lakes in Egypt, Chad, Ethiopia, and Kenya. The likely availability of surface water within depressions and basins across the grassland would ensure that hominins could traverse the Sand Sheet and reach the perennial lakes of the Sudan and the Western Desert of Egypt.

East African sites contemporaneous with White Lake were sited near water sources (Clark et al., 2003; Basell, 2008; Grun, 2006). These sites show hominins were adapted to multiple vegetative zones but that water was a constant nearby. Mobility of modern hunter-gatherers is based on moving to resources without strategy changes (Kelly, 1992). If the same motivation can be applied to Pleistocene hominins, then expansion of ranges would require water sources within reasonable distances to accommodate the hominin adaptive strategy displayed in East Africa (Marlowe, 2004). The environment of the Saharan wet phases produces scattered perennial lakes with intervening surface water within depressions suitable for the needs of gazelle and other savanna-adapted fauna. These same wet periods could also provide the resources needed for hominin range expansion into the Western Desert.

CHAPTER EIGHT: CONCLUSIONS

Melanoides tuberculata from trench 10/86 of White Lake remnant had $\delta^{18}O$ along shell ranges up to 8.1‰ representing seasonal variation in the isotopic composition of the surrounding waters. None of the $\delta^{18}O$ values were more depleted than those of the sediment carbonate. Further, 77% of the $\delta^{18}O$ gastropod carbonate values from the White Lake remnant shallow water area are more enriched than those of the deeper water lake sediment carbonate. The more enriched shell carbonate values are likely due to a combination of moisture sources. *Melanoides* from trench 10/86 likely lived in shallow water that was more sensitive to isotopic composition changes due to multiple water source input. Shallow water would also be more sensitive to seasonal temperature changes that would affect the isotopic composition of shell carbonates secreted. The White Lake remnant represents a wet phase lake that was subject to seasonal differences involving temperature and water input from multiple sources.

The range of $\delta^{18}O$ White Lake remnant shell carbonate values (-7.7‰ to 0.8‰) does not point to a single moisture source. Futhermore, the lack of a significant magnesium content implies evaporation was not a factor in the isotopic composition of the surrounding water (McKenzie, 1993). Based on the enriched shell carbonate $\delta^{18}O$ values and insignificant evaporation, waters surrounding *Melanoides* were likely the result of multiple water sources. Enriched shell carbonate values point to the Indian Ocean and/or the Mediterranean Sea as source components for the White Lake waters. However, more depleted values of both sediment (McKenzie, 1993) and shell carbonate require additional depleted isotopic compositions such as Atlantic monsoon or groundwater. An increase in the percent magnesium composition may point to lake regression that created both the shallow water environment and coquina.

Finally, similarities between the White Lake and Bir Tarfawi ca. 125 ka remnants and comparisons with modern northern African lake landscapes suggest a gradation of vegetation surrounding the lake and extending out to an arid landscape. Shallow edge waters with reeds adjacent to shrubs and trees morphing into savanna would be a likely landscape. These types of vegetation were supportive for a number of species. The presence of lithics in the White Lake remnant suggests that hominins may have been one of those taxa.

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APPENDIX A

White Lake Remnant Shallow Water *Melanoides tuberculata* Shell Carbon and Oxygen Isotopic Composition

Table A.1 White Lake Remnant Shallow Water *Melanoides tuberculata* Shell Carbon and Oxygen Isotopic Composition

Shell ID	Shell Location (aperture, whorl, apex)	Sample ID	$\delta^{13}C$ (‰. VPDB)		$\delta^{18}O$ (‰, VPDB)	
			Mean	STD	Mean	STD
large 1	aperture	L011C	-2.96	0.04	-3.62	0.07
		L011b	-3.22	0.07	-4.28	0.07
		L011a	-3.40	0.06	-4.56	0.06
	2	L012c	-1.42	0.08	-4.99	0.09
		L012b	-1.18	0.04	-4.93	0.04
		L012a	-2.26	0.07	-4.81	0.10
	3	L013b	-1.81	0.04	-5.15	0.06
		L013a	-2.16	0.06	-5.52	0.06
	4	L014	-2.44	0.04	-5.94	0.05
	5	L015	-2.80	0.03	-6.49	0.06
	apex	L016	-2.64	0.05	-7.12	0.05
large 2	aperture	L17C	-2.63	0.09	-2.94	0.16
		L17B	-2.65	0.08	-2.98	0.13
		L17A	-2.63	0.07	-3.25	0.17
	2	L18C	-3.29	0.08	-6.57	0.22
		L18B	-2.20	0.09	-6.53	0.12
		L18A	-3.20	0.11	-6.54	0.21
	3	L19B	-3.29	0.04	-5.52	0.13
		L19A	-3.17	0.07	-4.78	0.12
	4	L20	-3.13	0.07	-4.80	0.13
	5	L21	-3.21	0.08	-5.27	0.14
	apex	L22	-2.33	0.05	-6.00	0.22
large 3	aperture	L301D	-1.53	0.05	-5.04	0.06
		L301C	-1.71	0.04	-3.91	0.06
		L301B	-1.79	0.03	-4.57	0.04
		L301A	-1.98	0.02	-5.58	0.03
	2	L302C	-2.33	0.02	-6.50	0.01
		L302B	-3.38	0.05	-7.67	0.03
		L302A	-2.94	0.03	-7.28	0.03
	3	L303B	-1.97	0.02	-5.86	0.03
		L303A	-1.89	0.07	-4.68	0.06
	4	L304	-2.33	0.05	-5.60	0.03
	5	L305	-2.11	0.04	-5.57	0.04
	apex	L306	-1.82	0.03	-5.43	0.03

Shell ID	Shell Location (aperture, whorl, apex)	Sample ID	$\delta^{13}C$ (‰. VPDB)		$\delta^{18}O$ (‰, VPDB)	
			Mean	STD	Mean	STD
medium 1	whole	M01	-3.39	0.06	-4.88	0.06
medium 2	aperture	L23C	-1.87	0.08	-2.14	0.16
		L23B	-2.10	0.10	-2.40	0.15
		L23A	-2.53	0.06	-3.26	0.16
	2	L24	-2.85	0.08	-3.36	0.13
	3	L25	-2.33	0.08	-3.00	0.19
	4	L26	-1.98	0.07	-2.13	0.12
	5	L27	-1.96	0.06	-3.08	0.14
	6	L28	-1.78	0.05	-2.53	0.24
medium 3	whole	M02	-2.22	0.05	-3.42	0.07
medium 4	whole	M03	-2.75	0.07	-3.60	0.08
medium 5	aperture	M101E	-2.04	0.05	-4.52	0.06
		M101D	-2.49	0.03	-3.61	0.04
		M101C	-1.76	0.03	-5.69	0.04
		M101B	-0.70	0.03	-4.89	0.02
		M101A	-2.49	0.02	-7.26	0.03
	2	M102C	-1.66	0.02	-1.30	0.04
		M102A	-0.41	0.04	-3.50	0.03
	3	M103B	-2.52	0.02	0.54	0.03
		M103A	-2.73	0.04	0.81	0.11
	4	M104	-3.74	0.02	-1.57	0.03
	5	M105	-3.19	0.06	-2.50	0.07
	apex	M106	-2.99	0.03	-3.89	0.03
medium 6	aperture	M201D	-2.30	0.03	-6.11	0.07
		M201C	-2.72	0.06	-5.80	0.07
		M201A	-2.68	0.03	-3.20	0.03
	2	M202	-2.71	0.03	-3.08	0.06
	3	M203	-2.95	0.03	-3.45	0.05
	4	M204	-2.75	0.03	-4.11	0.05
	5	M205	-2.13	0.05	-4.70	0.05
medium 7	aperture	M301D	-3.27	0.04	-4.67	0.02
		M301C	-3.32	0.04	-6.94	0.03
		M301BD	-2.49	0.03	-4.46	0.03
		M301B	-3.16	0.03	-7.23	0.02
		M301A	-3.43	0.03	-6.51	0.05
	2	M302B	-2.49	0.03	-3.69	0.05
		M302A	-3.13	0.04	-6.68	0.02
	3	M304	-2.12	0.03	-1.13	0.04
	apex	M305	-2.26	0.03	-1.61	0.03

Shell ID	Shell Location (aperture, whorl, apex)	Sample ID	$\delta^{13}C$ (‰. VPDB)		$\delta^{18}O$ (‰, VPDB)	
			Mean	STD	Mean	STD
medium 8	aperture	M401D	-2.22	0.02	-1.70	0.02
		M401C	-2.27	0.03	-1.80	0.02
		M401B	-2.94	0.03	-5.11	0.03
		M401A	-2.04	0.05	-2.48	0.06
	2	M402B	-2.48	0.03	-3.30	0.04
		M402A	-2.48	0.02	-2.62	0.03
	3	M403	-2.85	0.02	-5.51	0.05
	4	M404	-3.15	0.03	-5.41	0.04
	apex	M405	-3.26	0.04	-7.53	0.05
small 1	whole	S01	-3.48	0.06	-5.53	0.06
small 2	aperture	S141	-0.45	0.04	-1.13	0.03
	apex	S142	-0.94	0.04	-2.51	0.05
small 3	aperture	S131B	-2.60	0.02	-6.31	0.04
		S131A	-2.43	0.05	-7.35	0.04
	apex	S132	-2.81	0.04	-6.08	0.03
small 4	aperture	S101B	-3.32	0.05	-2.38	0.03
		S101A	-3.19	0.04	-3.55	0.05
	apex	S103	-3.30	0.05	-4.23	0.05
small 5	aperture	S111B	-0.70	0.03	-4.63	0.04
		S111A	-0.68	0.03	-4.62	0.03
	apex	S112	1.03	0.04	-2.98	0.06
small 6	aperture	S121B	-2.45	0.02	-2.56	0.04
		S121A	-2.44	0.03	-2.81	0.04
	apex	S122	-3.31	0.02	-3.52	0.04
small 7	whole	S02	-2.56	0.04	-4.18	0.05
small 8	whole	S03	-2.16	0.05	-1.00	0.04
Overall	Mean		-2.42	0.05	-4.30	0.07
	STD		0.79		1.82	
	Min		-3.74	0.02	-7.67	0.01
	Max		1.03	0.11	0.81	0.24

APPENDIX B

Bir Tarfawi ca. 125 ka Remnant Shallow Water *Melanoides tuberculata* Shell Carbon and Oxygen Stable Isotope Composition

Table B.1 Bir Tarfawi ca. 125 ka Remnant Shallow Water *Melanoides tuberculata* Shell Carbon and Oxygen Stable Isotope Composition

Shell ID	Shell Location (aperture, whorl, apex)	Sample ID	$\delta^{13}C$ (‰. VPDB)		$\delta^{18}O$ (‰, VPDB)	
			Mean	STD	Mean	STD
large	aperture	L101D	-2.79	0.02	-1.63	0.02
		L101C	-2.49	0.03	-2.76	0.05
		L101B	-2.94	0.04	-2.45	0.03
		L101A	-2.86	0.03	-2.75	0.02
	2	L102B	-2.54	0.04	-3.54	0.02
		L102A	-2.98	0.04	-3.30	0.04
	3	L103	-3.65	0.02	-3.55	0.01
	apex	L104	-3.76	0.03	-0.47	0.02
medium 1	aperture	M101C	-2.16	0.02	-4.03	0.04
		M101B	-2.11	0.03	-4.65	0.02
		M101A	-2.76	0.03	-6.04	0.03
	2	M102	-3.14	0.03	-5.91	0.04
	3	M103	-3.25	0.02	-4.34	0.02
	apex	M104	-3.00	0.04	-3.54	0.05
medium 2	aperture	M201E	-3.23	0.05	-1.45	0.08
		M201D	-3.10	0.02	-3.61	0.05
		M201C	-2.94	0.02	-3.76	0.03
		M201AB	-3.34	0.05	-4.99	0.08
	2	M202	-3.63	0.02	-5.38	0.03
	3	M203	-2.93	0.04	-1.81	0.07
	apex	M204	-2.69	0.03	0.49	0.04
small 1	aperture	S101C	-5.24	0.05	-2.86	0.08
		S101B	-5.59	0.03	-3.95	0.03
		S101A	-5.25	0.03	-4.14	0.03
	2	S102	-4.66	0.03	-2.88	0.05
	apex	S103	-4.11	0.05	-1.43	0.04
small 2	aperture	S201B	-2.46	0.06	-3.60	0.08
		S201A	-2.77	0.03	-3.44	0.03
	2	S202	-3.08	0.03	-1.98	0.04
	apex	S203	-3.28	0.03	-2.46	0.06
Overall	Mean		-3.29	0.03	-3.21	0.04
	STD		0.88		1.50	
	Min		-5.59	0.02	-6.04	0.01
	Max		-2.11	0.06	0.49	0.08

APPENDIX C

Bir Tarfawi ca. 125 ka Remnant Deep Water *Melanoides tuberculata* Shell Carbon and Oxygen Stable Isotope Composition

Table C.1 Bir Tarfawi ca. 125 ka Remnant Deep Water *Melanoides tuberculata* Shell Carbon and Oxygen Stable Isotope Composition

Shell ID	Shell Location (aperture, whorl, apex)	Sample ID	δ13C (‰. VPDB)		δ18O (‰, VPDB)	
			Mean	STD	Mean	STD
large	aperture	L101B	-3.91	0.02	-6.22	0.03
		L101A	-4.24	0.02	-7.75	0.02
	2	L102	-3.98	0.02	-4.91	0.02
	3	L103	-4.09	0.04	-4.46	0.05
	apex	L104	-3.98	0.03	-4.65	0.04
medium 1	aperture	M101B	-3.16	0.02	-5.60	0.02
		M101A	-2.14	0.02	-4.85	0.03
	2	M102	-0.37	0.01	-0.52	0.02
	apex	M103	-0.98	0.02	-2.54	0.03
medium 2	aperture	M101B	-5.05	0.03	-5.60	0.03
		M101A	-5.03	0.03	-4.17	0.03
	2	M102	-4.51	0.03	-5.64	0.02
	apex	M103	-3.66	0.02	-5.74	0.02
medium 3	aperture	M101	-4.60	0.03	-5.89	0.06
	2	M102	-4.47	0.03	-7.57	0.03
	apex	M103	-3.96	0.03	-6.19	0.06
medium 4	aperture	M201	-2.52	0.03	-3.57	0.04
	2	M202	-2.51	0.03	-4.27	0.03
	apex	M203	-3.18	0.05	-4.33	0.06
medium 5	aperture	M301	-5.55	0.03	-5.45	0.04
	2	M302	-5.14	0.03	-6.93	0.05
	apex	M303	-5.61	0.03	-5.46	0.03
medium 6	aperture	M101B	-5.21	0.03	-8.01	0.02
		M101A	-5.25	0.04	-7.48	0.05
	2	M102	-5.10	0.03	-7.22	0.02
	3	M103	-5.50	0.03	-5.35	0.03
	apex	M104	-6.22	0.05	-5.27	0.08
small 1	aperture	S101	-3.62	0.01	-6.60	0.01
	apex	S102	-4.62	0.02	-3.94	0.03

Shell ID	Shell Location (aperture, whorl, apex)	Sample ID	$\delta^{13}C$ (‰. VPDB)		$\delta^{18}O$ (‰, VPDB)	
			Mean	STD	Mean	STD
small 2	aperture	S201	0.05	0.01	-3.49	0.02
	apex	S202	0.23	0.03	-1.29	0.03
small 3	aperture	S101	-2.92	0.02	-3.39	0.02
	2	S102	-3.97	0.02	-4.28	0.03
	apex	S103	-4.11	0.02	-4.48	0.04
small 4	aperture	S201	-4.49	0.01	-6.48	0.01
	2	S202	-5.15	0.03	-6.32	0.05
	apex	S203	-5.21	0.02	-6.58	0.02
small 5	aperture	S301	-2.38	0.01	-3.91	0.02
	2	S302	-2.61	0.03	-3.84	0.03
	apex	S303	-2.90	0.01	-3.74	0.02
small 6	aperture	S101	-3.56	0.04	-6.30	0.04
	apex	S102	-3.08	0.02	-3.40	0.04
small 7	aperture	S201	-2.57	0.04	-5.82	0.07
	2	S202	-2.95	0.03	-4.48	0.05
	apex	S203	-2.75	0.03	-3.37	0.04
small whole	aperture	S101	-2.78	0.04	-4.75	0.04
	apex	S201	-4.24	0.02	-5.82	0.03