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LATE QUATERNARY LAKES IN THE ZIWAY-SHALA BASIN, SOUTHERN ETHIOPIA.

UNIVERSITY OF CAMBRIDGE (GREAT BRITAIN), PH.D., 1979

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LATE QUATERNARY LAKES IN THE ZIWAY-SHALA BASIN, SOUTHERN ETHIOPIA

by

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Dissertation submitted to the University of Cambridge for the Degree of Doctor of Philosophy

April 1979

UNIVERSITY OF CAMBRIDGE

Board of Graduate Studies

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Late Quaternary lakes in the Ziway-Shala Basin, Southern Ethiopia

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Date 30th. April. 1979 Signed F. Alayne Street.

Abstract

The sequence of environmental changes in the Main Ethiopian Rift during Late Quaternary time has been reconstructed from fluctuations in lake level in the Ziway-Shala Basin, a closed drainage situated at 7°-8°30'N. The four present-day residual lakes (1558-1636 m.a.s.l.) are fed primarily by runoff from the Ethiopian Highlands. The highest summits on the eastern divide of the basin ($\geqslant$3850 m) were glaciated during the Late Quaternary. During wetter phases the lakes united to form a single waterbody which overflowed into the Awash River to the north via the Meki-Dubeta col (ca. 1670 m). Past water-level fluctuations have been established using three main approaches: 1) a detailed geomorphic and bathymetric investigation of the former lake bed and shorelines, based on field and air-photo studies; 2) stratigraphic description and ^{14}C dating of the Rift-floor sediments and 3) a limited investigation of depositional environments. Changes in lacustrine conditions have also been related to variations in ice extent, vegetation and soils within the catchment area. Four major lacustrine highstands and six minor ones have occurred since 30,000 BP. These can be grouped into a complex Late Pleistocene lacustral interval (ca. 30,000 to ca. 21,000 BP) and a complex postglacial lacustral interval (ca. 11,500 to ca. 4800 BP); separated by a prolonged period of aridity. Since 4800 BP, lake levels have remained low and fluctuating, apart from a brief late Holocene maximum after 2500 BP. Overflow into the Awash occurred intermittently during the intervals ca. 24,000 - ca. 21,000 BP, 9400 - 8500 BP and 6500 - 4800 BP. Because the topography of the basin has remained relatively stable over the last 30,000 years, it is possible to interpret these fluctuations in climatic terms, although this is not valid for earlier periods. A simple water-balance model has been used to compute the changes in rainfall and runoff required to sustain the lakes at different times. This indicates that precipitation has varied by at least $\pm$ 25% since 20,000 BP. The reconstructed sequence of climatic and environmental changes closely parallels the history of other lakes fed from the Ethiopian Highlands, particularly Abhé.

Acknowledgments

This dissertation is dedicated to the memory of Dr. Bill Morton, Acting Head of the Department of Geology, Addis Ababa University, who provided the logistical support and encouragement which made my fieldwork possible. I would also like to thank Frances Dakin (Mrs. M.A.J. Williams) and Dr. Chuck Wood for their generous hospitality and for introducing me to the geology of Ethiopia. I am deeply indebted to Bekele-Furu Desissu, Jinty Milne, Clare Blinch, Mikael Tesfaye, Liz Oswald, David Lauder, Berhanu and Abbe for their cheerful and competent assistance in the field.

Permission to work in Ethiopia was granted by the Institute of Ethiopian Studies, Addis Ababa University, and by the Ethiopian Ministry of Mines. I would also like to acknowledge the help I received from the Ethiopian Wildlife Conservation Organization, the Ethiopian Ministry of Land Reform and Administration, the National Water Resources Commission, the Awash Valley Authority, the Chilalo Agricultural Development Unit, and AIDBANK. I am particularly grateful to the UNDP Geothermal Project for their cooperation and for allowing me to use their O-itu camp facilities during my 1975-76 field season. Professor Pierre Gouin, of the Geophysical Observatory, kindly loaned me various items of surveying equipment, and his colleagues Dr. David Emilia and Dr. Brian Last were responsible for the geophysical soundings of the Meki-Dubeta overflow channel. Ato Getachew Aweke, of the Herbarium, Addis Ababa University, carried out a botanical survey of the Mt. Badda bog sites. Dr. Peter Goll of the Department of Parasitology provided me with transport to the island of Tulu Gudu.

I have benefited from many stimulating discussions with Professor Bill Bishop; Ron Bordley (D.O.S.); Dr. Jim Bowler; Professor Desmond Clark; Dr. Gerry Dekker (N.W.R.C.); Dr. Mario Di Paola; M. Frédéric Gèze; Professor Glyn Isaac; Dr. Celia Kamau; Ted Lloyd (UNDP); John

Makin, Tim Kingham, and Chris Birchall (L.R.D.); Professor Paul Mohr; Neil Munro (Huntings); Dr. Maurice Taieb; Dr. Mike Talbot; Tom Tischler (E.W.C.O.); and Dr. Martin Williams. Professor Karl Butzer has provoked me into re-evaluating some of my most cherished ideas. Professor Mohr, Mr. Lloyd, Dr. Williams and Dr. A. Baumann have all supplied me with unpublished geological data.

I would like to thank the following experts who have identified specimens collected in the Ziway-Shala Basin and advised me on problems of taxonomy: Dr. D.S. Brown, Dr. P. Goll, Dr. J.G.J. Kuiper, Mr. B.W. Sparks and Dr. B. Verdcourt (mollusca); Dr. K.E. Banister and Dr. P.H. Greenwood (fish remains); Dr. R.H. Bate (ostracods); Mr. T.R. Williams (crab remains); Dr. G.E. Wickens and the Jodrell Laboratory, Kew (charcoal and wood); Ato Getachew Aweke and Dr. B. Verdcourt (plant specimens); Dr. D. Geraads and Mr. R. Ciochon (mammalian bones); and Professor J.D. Clark, Mr. H. Kurashina and Mr. S. Brandt (artifacts). Dr. Alan Hamilton kindly carried out pollen analyses on cores collected from Mt. Badda and the Balé Mountains. All the diatom analyses from the Ziway-Shala Basin are the work of Dr. Françoise Gasse and her student Mlle. Claudine Descourtieux, to whom I am indebted for permission to quote their unpublished results. I would also like to thank Dr. G. Delibrias, Dr. R. Gillespie, Dr. R.L. Otlet and Dr. V.R. Switsur for the radiocarbon dates which form the backbone of this investigation.

Facilities for sedimentological analyses were provided by the Department of Geography, Cambridge; the Laboratoire de Géologie du Quaternaire, C.N.R.S., Bellevue; and the School of Geography, Oxford. I am very grateful to M. Lucien Casta (C.N.R.S.) and Dr. R. Perrin (Dept. of Applied Biology, Cambridge) for their advice on analytical techniques and to Mike Diver, Sophie Bieda and Chris Jackson for their help with laboratory analyses. I would also like to thank the Department of Mineralogy and Petrology, Cambridge, the Department of Geology and Mineralogy, Oxford, and Oxford Polytechnic for the use of their equipment. Computer time was provided by the Oxford University Computing Laboratory. The figures in this dissertation were drafted by Mr. A. Shelley, Miss M. Loveless, Mrs. J. Hill and Miss M. Potter, and photographed by Mr. J. Landon, Mr. P. Masters and Mr. R. Coe. Table 8.1 was set up by Mr. R.B. Toomey of Computerset, Ltd.

Financial support for this investigation was provided by a NERC postgraduate studentship and subsequently by a grant from C.N.R.S. (A.T.P. internationale no. 2043, under Dr. R. Bonnefille). I would also like to acknowledge the generous contributions made by the Royal Society and the Department of Geography, Cambridge, towards my fieldwork expenses.

Finally, I must extend my warmest thanks to my supervisor, Mr. A.T. Grove, to my friend and colleague Dr. Françoise Gasse, and most of all to my parents, for their unfailing support and encouragement.

F. Alayne Street
April 1979.

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CHAPTER ONE

Introduction

Since the middle of the nineteenth century, geologists have recognized that the global climatic changes which brought about Quaternary glaciations also induced modifications in the hydrological cycle in low latitudes (Lartet, 1865; Russell, 1885; Gilbert, 1890). The clearest evidence for fluctuations in water balance is provided by the varying extent of closed-basin lakes, which are deep and dilute during periods of increased runoff or reduced evaporation, but become shrunken and often saline during episodes of greater aridity. The environmental and archaeological impact of changes in lacustrine conditions has been particularly marked in eastern Africa. This was noted by many early geologists and prehistorians such as Gregory (1921), Leakey (1931), Wayland (1934) and Nilsson (1931, 1935, 1940). Following the view already adopted in the United States, most of these workers assumed a correlation between the "pluvial" periods, during which the lakes were more extensive, and northern hemisphere glaciations. According to Nilsson (1940, p. 2):

> The existence of these great volumes of water where we now find only small drying lakes must be due to former moister climates during so called Pluvial epochs. The much debated "Desiccation of Africa", already recorded by the first exploring white men in East Africa, is the present link in the sequence of events which register the changes in climate.

In East Africa, a sequence of four pluvial periods, thought to be equivalent to the four classical glaciations of the Alps, was erected on the basis of geological investigations at many scattered and poorly dated sites (Washbourn, 1967a; Bishop. 1971). This pluvial/glacial scheme relied for its theoretical basis on the climatological theories of Brooks (1914) and Simpson (1934). Alternative climatic models for eastern Africa (Hume and Craig, 1911) were conveniently ignored. Following the First Pan African Congress on Prehistory in Nairobi in 1947, the stratigraphic-climatic /

terminology acquired a veneer of official respectability which encouraged its application across the whole of the African continent, in areas far removed from the original type localities. It still occasionally crops up in papers written as recently as 1975 (Pouclet, 1975).

From 1939 onwards, however, the "Pluvial Hypothesis" was strongly challenged on both stratigraphic and climatological grounds (Solomon, 1939; Balout, 1952; Tricart, 1956; Cooke, 1958; Flint, 1959). The French geographer Tricart wrote in a prophetic paper in 1956:

> We should ... distinguish two geographical types of Pluvials: those of mid-latitudes, which coincide with glaciations (North Africa, western United States, Iran etc.) and those of low latitudes which, on the contrary, fit into interglacial periods. (p. 167, transl.)

Within a few years his predictions were verified by the first radiocarbon dates on lake sediments from the tropical Sahara (Faure et al., 1963). In 1965 the Wenner-Gren Symposium officially recommended the abandonment of the pluvial nomenclature and a return to normal stratigraphic practice (Bishop and Clark, 1967).

Since this turning point in 1965, work on the ancient lakes of eastern Africa and the Sahara has revealed an increasingly complex picture of lake-level response to Late Quaternary climatic changes. Two main lines of research have emerged. One group of workers, such as Celia Washbourn-Kamau (Washbourn, 1967a,b; Washbourn-Kamau, 1971, 1972; Kamau, 1977), Michel Servant (1973), Professor Karl Butzer (Butzer et al., 1969, 1972; Butzer, 1976) and Martin Williams (Williams et al., 1977) have concentrated their attention on shoreline features and/or sedimentary exposures, while others, notably the students of Professor Dan Livingstone at Duke University, have studied the microflora and sediment chemistry of lacustrine cores (Kendall, 1969; Richardson and Richardson, 1972; Hecky and Degens, 1973; Harvey, 1976; Holdship, 1976). The first approach has the advantage that it yields fairly precise data on lake depth and area at different times. In basins which have been tectonically and hydrographically stable, these figures can be converted into estimates of past precipitation using simple water-balance methods (Washbourn, 1967b). The second approach does not readily provide quantitative data on lake depth, but has furnished a wealth of information on lake-water chemistry, productivity, and surface temperatures, as well as on sediment inputs from the catchments. Since salinity serves as a useful independent check on reconstructed water levels, the two methods are essentially complementary. They were successfully combined in a comprehensive study of the palaeolakes in the Central and Northern Afar by Francoise Gasse (Gasse, 1975, 1977a; Gasse and Delibrias, 1977; Gasse and Street, 1978a).

Dr. Gasse argues convincingly that high lake levels in the Horn of Africa occurred during the period 40,000 - 23,000 BP, which is regarded as an inter~~glacial~~stadial in higher latitudes (Dreimanis and Raukas, 1975), as well as during the Holocene interglacial. Her work has clearly demonstrated the sensitivity of the Rift Valley lakes to variations in runoff from the Ethiopian Highlands.

This thesis is an investigation of the fluctuations in lake level during Late Quaternary time in the Ziway-Shala Basin, an internal drainage basin in South-central Ethiopia (Figure 1.1). At this latitude the Main Ethiopian Rift, which forms the northern extension of the Eastern Rift System of Africa, bisects the Ethiopian Highlands, creating a series of closed lake basins bounded to NW and SE by massive, faulted escarpments and to NE and SW by volcanic complexes located at intervals along the Rift floor.

The Ziway-Shala Basin is in many ways an ideal "field laboratory" for studies of Late Quaternary environmental change; and in particular for testing the relationship between lake-level fluctuations and glaciation (Nilsson, 1940; Grove et al., 1975). The four present-day lakes, which lie at 1558 - 1636 m.a.s.l., are the shrunken remnants of a much larger waterbody which once overflowed into the Awash catchment to the north. Its shorelines are clearly visible up to altitudes of 112 m above the lowest lake (Lake Shala), and show remarkably little tectonic or isostatic deformation. Because of this long-term stability, Late Quaternary fluctuations in water level can be confidently interpreted in climatic terms.

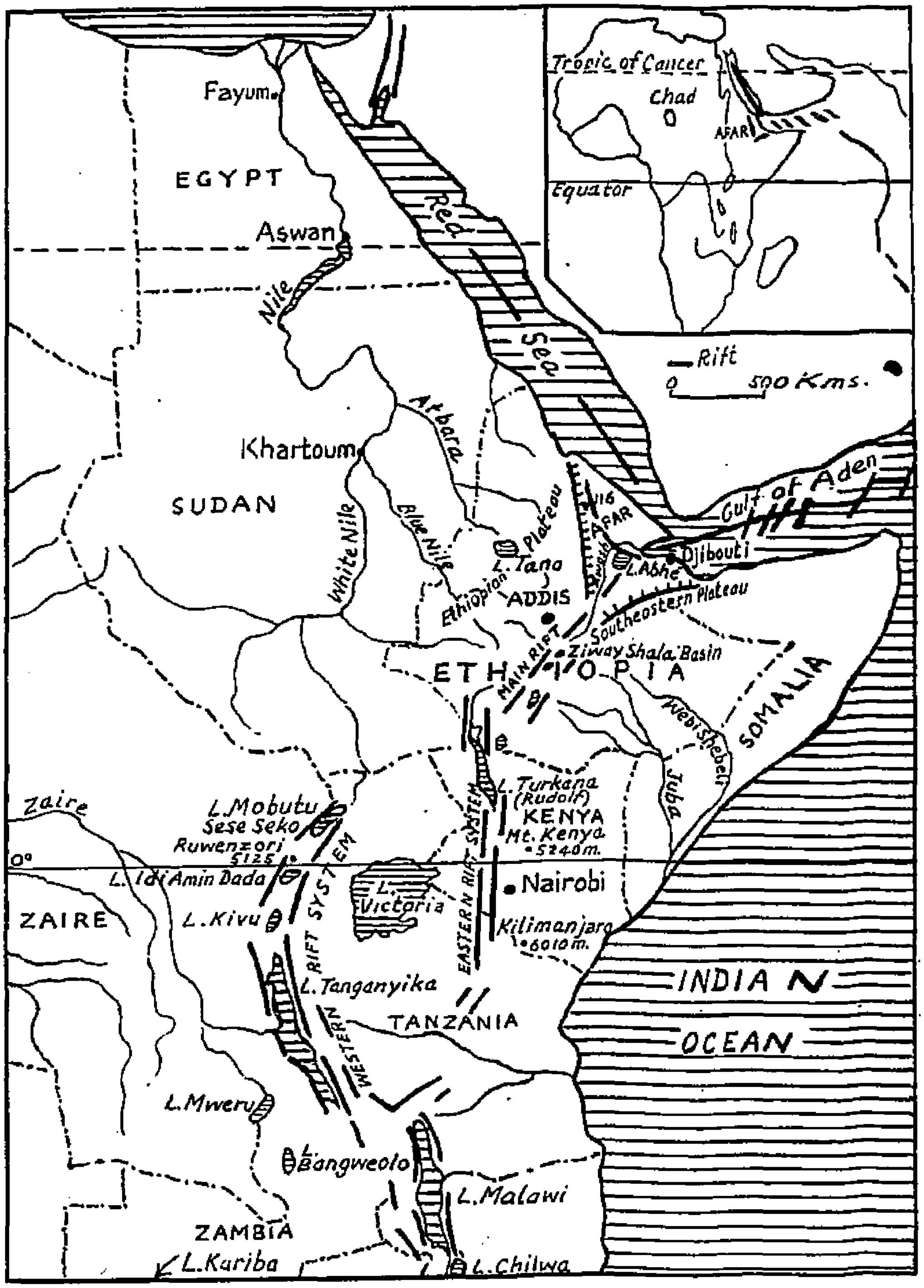


Fig. 1.1: General location map of eastern Africa, showing the Rift system.

The information obtained from geomorphic studies of the lacustrine strandlines has been confirmed and extended through stratigraphic investigations of the Rift-floor sediments. Deep exposures have been created by surface streams as the lakes receded after their last major highstand. The sections along the Bulbula gorge are up to 50 m deep and probably represent some of the finest exposures of Late Quaternary lacustrine sediments in Africa. In addition, the surface sequences can now be compared with a long, continuous core (162 m) from the northern shore of Lake Abiyata. The diatom assemblages in this core, which are being studied by Mlle. Claudine Descourtieux, a research student working under Dr. Francoise Gasse, provide an independent source of evidence on the variations in water depth and salinity at the centre of the basin.

A further advantage is that a wide range of materials suitable for radiocarbon dating are present in the Rift-floor sediments. These include charcoal, carbonized wood, organic mud, freshwater shells, ostracods, fish bones, lake marl, algal limestones and soil humus. For this reason it has not been necessary to rely exclusively on carbonates, which are particularly susceptible to contamination, in order to establish a lake-level chronology; although the dominance of soda-rich volcanics over sedimentary rocks within the catchment provides exceptionally favourable conditions for dating shell and other lacustrine carbonates.

On the eastern boundary of the Ziway-Shala Basin, the land rises to over 4000 m. The highest summits bear glacial moraines, indicating that small ice caps and cirque glaciers developed on them during the Late Quaternary. This situation not only provides an opportunity almost unique in Africa to establish the relationship between glaciation and lake-level fluctuations within a single catchment, but also makes it possible to estimate the temperature lowering at the last glacial maximum, based on the altitudinal depression of the snowline. Pollen studies by Dr. Alan Hamilton on a peat core from the Afroalpine area provide a valuable indication of vegetation changes in the mountains since deglacierization.

Several features make the Ziway-Shala Basin particularly suitable for a quantitative study of past changes in precipitation using water-balance methods. They include the relatively slow rate of crustal extension in this part of the Main Rift; the large horizontal and vertical amplitude of the water-level fluctuations; the availability of supporting information on variations in lake-water salinity, based on the diatom studies by Dr. Gasse and her students; and the existence of unusually detailed modern topographic, hydrological and limnological surveys. Past evaporation rates from open water can be calculated using palaeotemperature estimates derived from local snowline and diatom data, and a tentative attempt can also be made to assess the effects of the vegetation changes indicated by pollen analysis on surface runoff from the catchment.

The main aims of this study are therefore as follows:

1) To assemble and evaluate the geomorphic and stratigraphic evidence for variations in lake level and salinity in the Ziway-Shala Basin, in order to establish the sequence of fluctuations through time.

2) To calculate the changes in precipitation required to sustain the lakes at selected times in the past, using a simple water-balance approach.

3) To compare the sequence of climatic and environmental changes reconstructed in this way with the evidence from other lakes in the Horn of Africa (Taieb, 1974; Grove et al., 1975; Gasse, 1975, 1977a; Gasse and Delibrias, 1977; Gasse and Street, 1978a; Butzer et al., 1969, 1972; Butzer, 1976; Williams et al., 1977). More general comparisons with the rest of Africa and other low-latitude areas can be found in Street and Grove (1976, and in press).

This thesis is organized in the following way: Chapters 2 and 3 review the existing information on modern conditions in the Ziway-Shala Basin and summarize the findings of previous research on Late Quaternary environments. Chapter 4 is an outline of the methods used in the present investigation. Chapters 5 and 6 describe the sequence of lake shorelines and the stratigraphy of the most important sections, while Chapter 7 briefly outlines the sedimentological criteria which were used to interpret the characteristics of the lake deposits in environmental terms. All this information is brought together in Chapter 8 to provide a picture of the variations in lacustrine conditions through time. This is compared with the evidence for changes in ice extent, vegetation cover and soil development in the catchment. Chapter 9 describes the water-balance model which was used to convert the fluctuations in lake area into estimates of past precipitation. Finally, in Chapter 10, the environmental and climatic history of the Ziway-Shala Basin is compared with other lakes in Ethiopia, Northern Kenya and Djibouti which are also fed directly or indirectly from the Ethiopian Highlands.

CHAPTER TWO

Present-Day Environments of the Ziway-Shala Basin

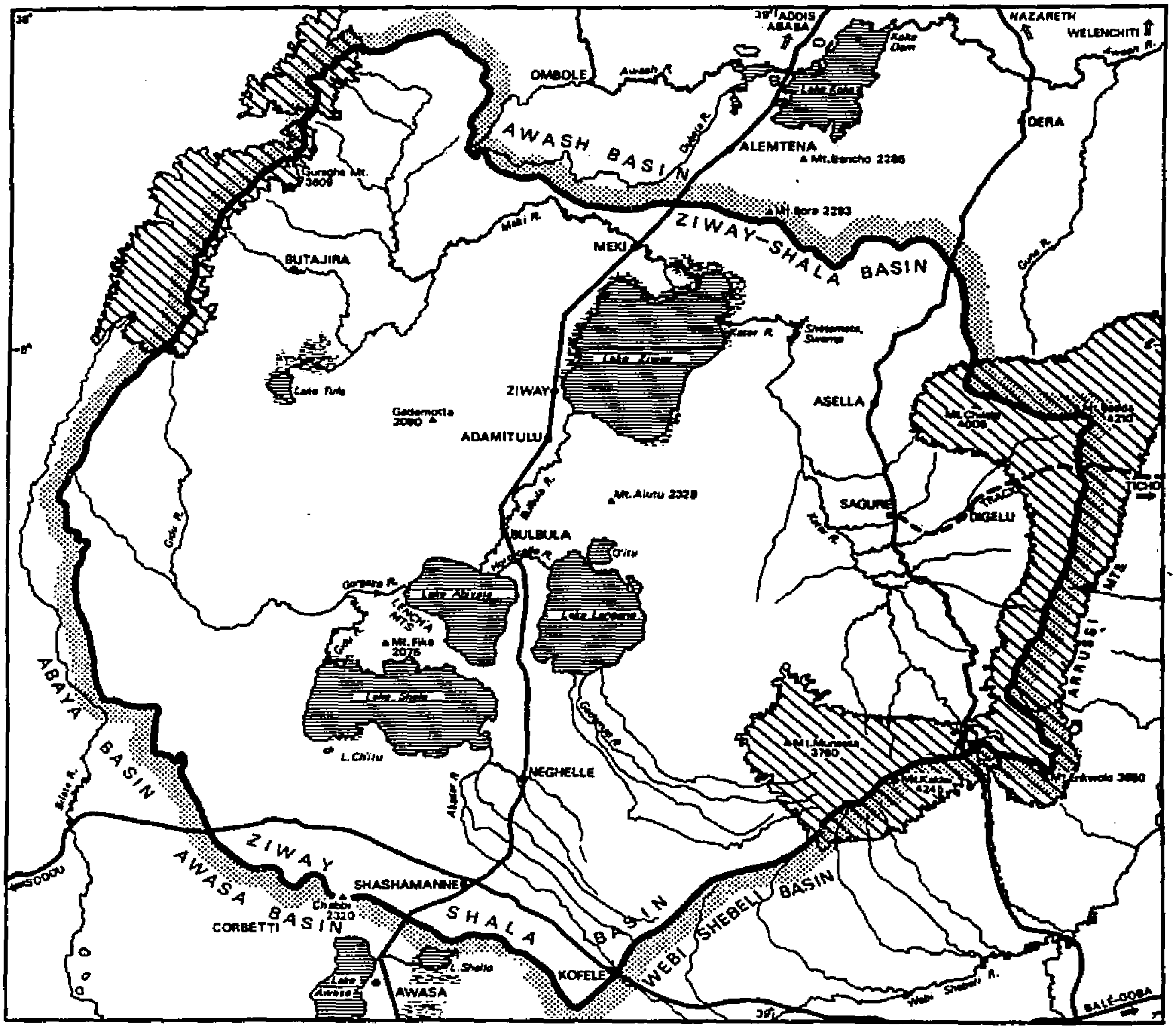
2A GEOLOGICAL SETTING

I. Introduction

The Main Ethiopian Rift is a NNE-SSW trending trough bordered by the Ethiopian and Southeastern Plateaus, which exceed 2500 m in elevation over wide areas (Fig. 2.1). In the region of the Ziway-Shala Basin, the Rift floor rises to 1550-1700 m above sea level. It is bounded by strongly faulted escarpments, 70-80 kms apart (Figs. 2.2, 2.3).

In marked contrast to the adjacent Afar sector, where new basaltic crust is being generated along an accreting plate margin, the Main Rift is a continental graben characterized by slow crustal attenuation and predominantly silicic volcanism (Barberi et al., 1975). Useful summaries of the regional geology and tectonics can be found in Baker et al. (1972), Di Paola (1972a) and Meyer et al. (1975).

The continuing development of the rift system raises several important problems which confront any attempt to interpret past fluctuations of the lakes. The most serious concerns the topographic stability of the lake basins and their catchment areas. One of the main reasons for the abandonment of the original "pluvial hypothesis" was the dawning realization that climatic effects could not easily be disentangled from those of earth movements (Bishop, 1971). Although Celia Washbourn-Kamau, in her painstaking study of the Kenyan lakes Nakuru and Elmenteita (1967b) was able to show that their Holocene strandlines were essentially undisturbed, measurements of shorelines in more active sectors such as the Afar have revealed significant mid- to late Holocene tilting of the graben floors (Rognon, 1975b; Gasse, 1975). Where these displacements are greatest, around Lake Asal, it becomes very difficult to reconstruct the geometry of former lakes with any confidence (Gasse and Stieltjes, 1973; Gasse, 1975).



Land over 3000 metres

Modern lakes

Marsh

Boundary of Ziway Shala Basin

Mountain peaks (heights in metres)

mls 0 5 10 15

0 5 10 15 20 kms

Fig. 2.1: General location map of the Ziway-Shala Basin (topographic base: U.S. Defense Mapping Agency 1:250,000 maps).

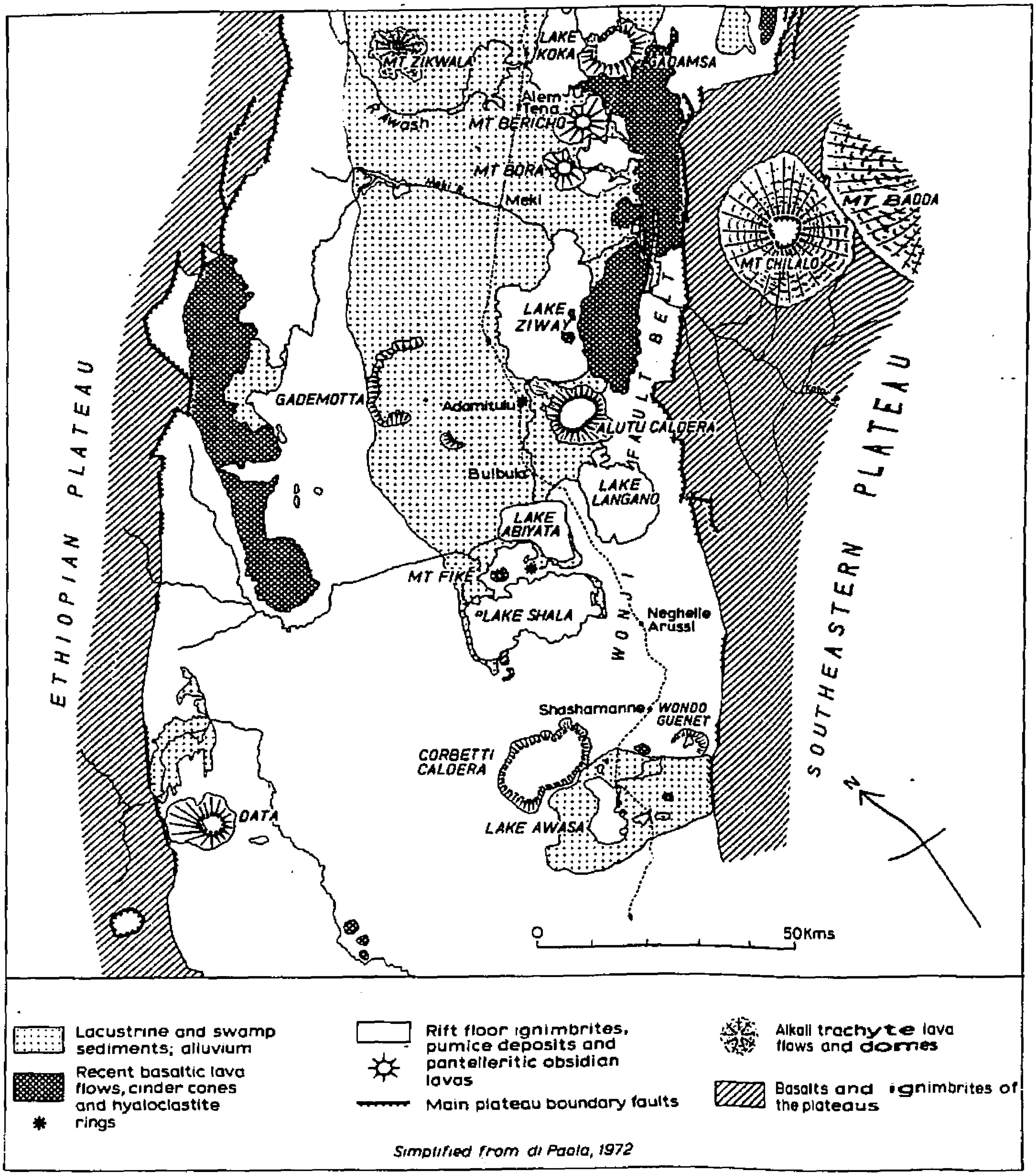


Fig. 2.2: Simplified geological map of the Main Rift between 7^{o} and $8^{o}40'$ N.

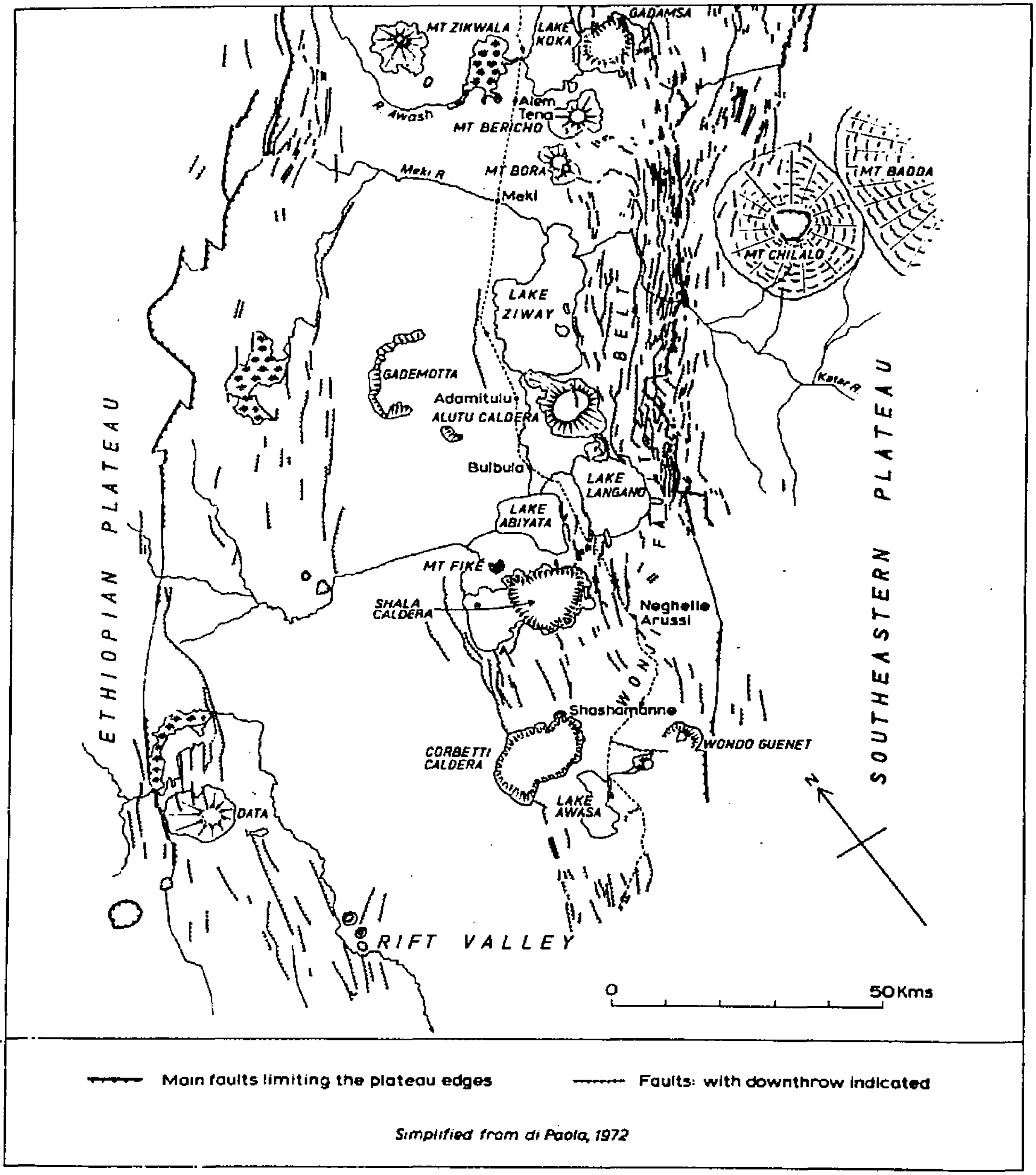


Fig. 2.3: **Simplified tectonic map of the Main Rift between 7° and 8°40' N.**

In such cases, palaeohydrologic calculations could be seriously misleading.

In section IV (below), I have tried to consider whether or not the drainage divides and shoreline levels of the Ziway-Shala Basin could have been disturbed by Late Quaternary faulting and volcanicity. Side effects of this activity could also be important. For example, in Lake Kivu, long-term variations in hot-spring activity are thought to have had an important influence on water levels (Hecky and Degens, 1973). In addition, since the distinctive character of the Ziway-Shala lake sediments is largely due to the explosive silicic volcanism typical of the Main Rift, we need to know something of the time-distribution of the eruptions in order to interpret past depositional environments.

Some attempt to answer these questions can be made from the available data on the evolution of the lake basin and the record of its shorelines and sediments.

II. Evolution of the Main Rift

Both plateaus are largely underlain by '*Trap Series*' flood basalts and intercalated ignimbrites of Oligocene to Miocene age (28 to 19 m.y. and upwards), which rest on an upwarped Mesozoic basement. The updoming reaches a maximum in the Ziway-Awasa area (Baker *et al.*, 1972; Morbidelli *et al.*, 1975; Jones, 1976). However most of these basalts are now concealed by a cover of silicic stratoid volcanics, the *Nazareth Series* of Meyer *et al.* (1975), which is dated from 9.5 to ca. 2 m.y. and also underlies the present Rift floor (Di Paola, 1972a). The alkaline and peralkaline ignimbrites and related lavas of the Nazareth Series seem to have issued mainly from fissures bordering an embryonic rift trough.

An abortive phase of crustal extension on the Southeastern Plateau, to the east of the proto-rift, gave rise to the extinct Pliocene trachytic shield volcanoes of the Arussi Mountains; Kakka, Badda, Chilalo and Enkwolo, which rest on the Trap basalts. Dikes and lavas from these centres have been dated between 3.6 and 2.1 m.y. (Mohr and Potter, 1976).

The main phase of rifting seems to have occurred about 1.8 to 1.6 m.y. ago (Meyer et al., 1975). Some large and denuded rift-floor calderas, such as the Gademotta Ridge west of Ziway (dated at 1.05 m.y. by Laury and Albritton, 1975) may also have come into being at this time. Subsequent volcanic activity has been largely confined to the Wonji Fault Belt (Fig. 2.3), a 5-15 km wide, NNE to SSW-trending zone of intense Quaternary faulting which tends to be axial to the Rift floor (Mohr, 1967; Baker et al., 1972; Mohr et al., 1975; Meyer and Stets, 1976). The faults are normal, short and sinuous, and in places are associated with open tensional fissures (Mohr, 1967; Gibson, 1967; Gouin and Mohr, 1967). They are disposed en échelon along the rift, with zones of offset being characterized by young silicic caldera volcanoes (Fig. 2.2). Several important clusters of hot springs are centred on the fault belt and their role in maintaining the lakes is discussed in chapter 2C.

The Quaternary Wonji Series[1] volcanics are of several different types. Large volcanic complexes (Bora-Bericho) and reactivated calderas (Boseti Gudda, Gadamsa, Alutu, Corbetti) are spaced at intervals along the Wonji Fault Belt (Plate 1). Their latest volcanic products are alkali rhyolites (pantellerites) - mainly air-fall and ash-flow deposits, but also minor quantities of highly viscous obsidian lava (Meyer et al., 1975; Di Paola, 1972a). Lake Shala occupies a very large caldera (Plate 2) which has recently been dated at 0.24 - 0.20 m.y. (P.A. Mohr, in litt., 9.1.78). The foundering of the main basin was accompanied by the explosive eruption of extensive sheets of ignimbrite and pumice breccia (United Nations, 1973; E.F. Lloyd, 1977).

Basaltic lavas and cinder-cones of Quaternary age occur in two main fields (Fig. 2.2). One is located near the main western boundary faults of the Rift and the other east and northeast of Lake Ziway. Both clearly originated from fissure eruptions. Another small field of recent basalts lies southwest of Shala on the axis of the Rift. Where these eruptions took place

1. The Aden Series of Mohr (1971).

underwater, the highly explosive activity built up low rings or cones of fragmented basaltic ejecta (hyaloclastites) (Rittmann, 1962, pp. 72-3). Notable examples are the islands of Tulu Gudu and Galila in Lake Ziway, the Adamitulu tuff cone, and Lake Chitu Haro south of Shala.

Occasional subaquatic eruptions of rhyolitic magma have also occurred. Mt. Fiké (2075 m), on the isthmus between Shala and Abiyata, is a double cone of stratified pumice resulting from eruptions of base-surge type (Moore, 1967) (Plate 27). Similar subaqueous volcanics outcrop in the denuded cones of the Lencha Mountains on the south shore of Abiyata (E.F. Lloyd, 1977).

In Figure 2.2, the major rock types of the region are grouped to emphasize their distribution. Despite the occurrence of the Rift-floor basalts and minor exposures of older basalts in the Arussi Mountains and escarpment areas, the catchment is principally underlain by highly alkaline silicic volcanics. This accounts for the richness of the drainage waters in dissolved silica, Na^{+} and K^{+}. By contrast, Ca^{++} and Mg^{++} are relatively depleted. The copious production of pumice, notably by Alutu, has been a dominant influence on Late Quaternary basin sediments and soils. The implications of these factors for lake-water chemistry and deposits are elaborated in chapter 2E.

III. Late Quaternary volcanism

When a better chronological framework becomes available, it may prove possible to use some of the Middle and Late Pleistocene subaqueous volcanics to date the lake-level fluctuations. At present, however, the rift-floor sediments provide the most detailed evidence on the time span of Late Quaternary explosive activity.

Mts. Bora (2293 m) and Bericho represent one of the most recent volcanic complexes of the Rift Valley, according to Di Paola (1972a). They are probably responsible for a large proportion of the pumice deposits on the Meki-Awash divide, including a widespread surface ash in the mid-Meki Valley, but the perfect preservation of the mid-Holocene strandlines flanking Mt. Bora indicates that ejection of pumice has been minimal since 5000 BP.

Mt. Alutu (2328 m) (Plate 1) which is at present in a dormant fumarolic state, appears to have erupted at frequent intervals during the later Quaternary. The first dated records of its activity are probably the air-fall tuffs found at archaeological sites on the Gademotta Ridge (Albritton, 1974; Laury and Albritton, 1975). They were deposited around 0.181 ± 0.006 m.y. and 0.149 ± 0.013 m.y. ago. Both the Late Pleistocene and Holocene lake deposits of the Bulbula Plain contain air-fall and fluvially-reworked Alutu pumices, which are described in chapter 6C. Intermittent late Holocene activity is evident from two ash layers in the Macho area to the west (Fig. 5.9), which were deposited shortly before 1540 ± 60 and shortly after 230 ± 50 yr BP respectively (Haynes and Haas, 1974).

It is not yet possible to date the eruptions of the Corbetti complex in this way, but it is highly likely that some of the pumice layers in the Shala lake beds are derived from the dormant Chabbi centre. Even if the reported wartime ashfall from Chabbi is discounted (Mohr, 1971, p.228), it is clear from the very fresh obsidian flows that Corbetti has been active in the recent past.

In contrast, there is remarkably little evidence of basaltic volcanism within the last 10,000 years. Chitu Haro (Plate 3) and Mechefera, the freshest of the tuff rings southwest of Shala, clearly antedate the early/mid-Holocene highstand of the lakes, as do many of the basaltic lavaflows in the same area (Fig. 5.8). However some small spatter cones just north of Chitu Haro overlie Holocene deep-water silts. Tulu Bila, a fresh-looking cinder cone on the isthmus between Abiyata and Shala, must date from a Late Quaternary lowstand of the lakes: hyaloclastites would have been produced had the water level been high. Lavas from the crater outcrop below undated diatomites in the Mudi gully (0738-B14). Similarly, the line of hyaloclastite centres between Adamitulu and the island of Galila all show signs of having been breached or planed off by the Holocene lake. Like the basalt field east of Ziway, they are almost certainly of Late Pleistocene age.

Mt. Fiké, although its cone and twin craters are very well-preserved (Plate 27), is probably quite old. An eroded calcrete yielding MSA (Middle Stone Age) artifacts has developed on the volcanic ash overlying the rhyolite plug inside the main crater. The mountain can therefore be discounted as a source of recent volcanics.

IV. Topographic stability

We can now try to assess the stability of the lake basin during the time-period of interest for palaeoclimatic reconstructions. The picture of the Late Quaternary which emerges is one of comparative tectonic quiescence, despite locally active volcanism (P. Gouin, pers. comm. 1975; Mohr, 1977): relations between the Holocene strandlines and the Wonji Fault Belt only serve to confirm this impression.

It is now established that the summit volcanoes of the Arussi Mountains, which supported icecaps during the last glaciation, have existed since the Pliocene. The main features of the Rift floor date back to the Late Pliocene and Early Quaternary. The 162 m-long core taken from Lake Abiyata confirms that lacustrine conditions have existed for most of the last few hundred thousand years, although they must have been greatly modified by the collapse of the Shala caldera. The most critical factors are therefore the stability of the Ziway-Awash and Shala-Awasa watersheds, and the extent to which Late Quaternary lake levels could have been disturbed by either faulting or regional warping.

Lake Awasa lies in a down-faulted basin within the crest of an uplifted region near the eastern margin of the Rift (Mohr, 1967) (Fig. 2.2). The watershed to the north (lowest point ca. 1747 m) is largely underlain by ignimbrites, veneered in places by more recent pumices (Di Paola, 1972a,b). According to Mohr (1967), the formation of the basin predates the latest Chabbi faulting and obsidian lavas. Since no traces of a surface connection can be found, it appears that Shala and Awasa have remained separate during the Holocene and probably throughout the Late Quaternary.

The upper limit on lake levels has been set instead by the Ziway-Awash watershed. This is underlain by a thick sequence of bedded pumice units, attributable to the Bora-Bericho complex, resting on the ignimbrites and lavas of the Nazareth Series (Venzo, 1971a; Italconsult, 1970). The divide is evidently of partly depositional origin, although transverse arching of the Rift floor has also been invoked (Mohr, 1966; Venzo, 1971b).

Signs of a recent overflow across this divide are limited to the Meki-Dubeta channel (see chapter 5). The very close correspondence between the floor of this spillway and the uppermost early to mid-Holocene shoreline level eliminates the possibility of significant tilting or warping of the Rift floor during the last 10,000 years. Late Pleistocene highstands would seem to have been controlled by the same threshold. But too little is known about their shoreline features for the absence of tilting to be established. The work of Laury and Albritton (1975) has established that the watershed elevation was radically different as recently as the later Middle Quaternary (chapter 3C).

With one or two exceptions, the Wonji Fault Belt does not appear to have displaced Late Quaternary sediments. Along the east side of Lake Ziway, it overlaps the eastern boundary faults of the Rift (Fig. 2.3), and the greater part of the Rift floor is unfaulted. The fault zone is offset south of Mt. Alutu, and runs between Lakes Langano and Abiyata. Intersecting the eastern rim of Lake Shala, it then peters out southwards in a field of open fissures. The main axis is subsequently displaced several kilometres to the west between Shala and Corbetti (Mohr et al., 1975; Catlin et al., 1973, Gèze, 1974).

Although major displacements have occurred along the fault belt north-east of Langano since the Middle Pleistocene, the Holocene shorelines are unaffected. They faithfully follow the indentations of the Wonji faultblocks. Subsidence of the grabens to the east postdates the deposition of an outlier of deeply weathered, westward-dipping plateau gravels (0738-B24) which is now perched high up on Mt. Ilala, part of the eastern rim of O-itu (Fig. 5.6, Plate 11). The unrolled assemblage of evolved Acheulian bifaces discovered in situ by Frederic Gèze (Plate 4) suggests an age of the order of 0.2 to 0.3 m.y. for the deposit (Clark, 1975; Wendorf et al., 1975).

The only areas where Late Quaternary sediments are demonstrably faulted are west of Ziway, northeast of Lake Shala (Fig. 6.13), and east of

Fiké. The maximum displacements observed on the southern faults are ⩽ 5-10 m. Gèze (1974, p.175) has suggested that the West Ziway faultscarp is of Holocene age, but it is almost certainly much older (chapter 5C). There is correspondingly little sign of recent development of tensional fissures related to the fault belt. The only examples recorded within the confines of the Holocene or Late Pleistocene lakes lie west and north of Ziway (Albritton, 1974; Di Paola, 1976), and south of Chitu Haro.

We can therefore place some confidence in Late Quaternary lake-level fluctuations as a basis for water-balance calculations.

2B CLIMATE, VEGETATION AND SOILS

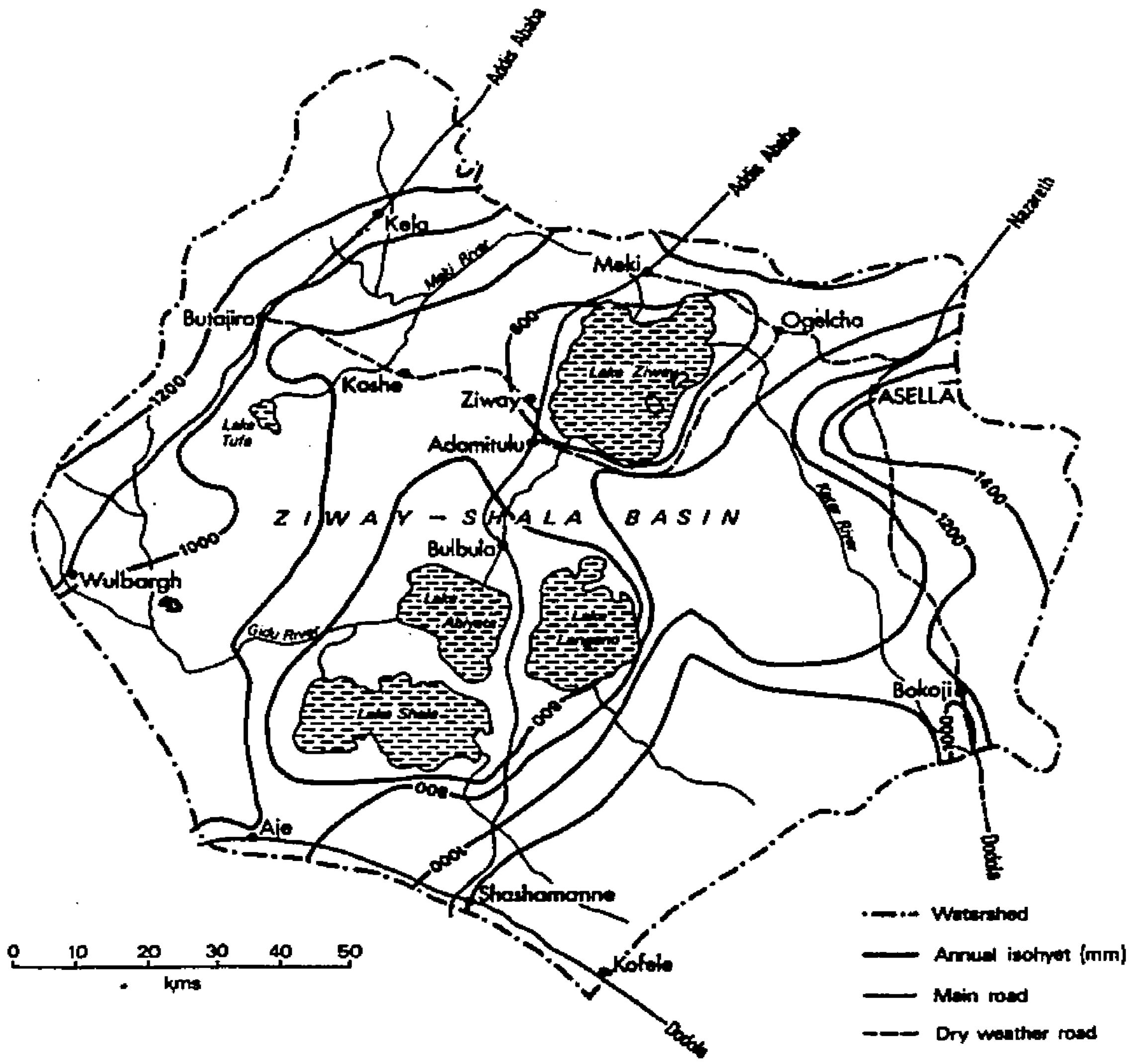
I. Climate

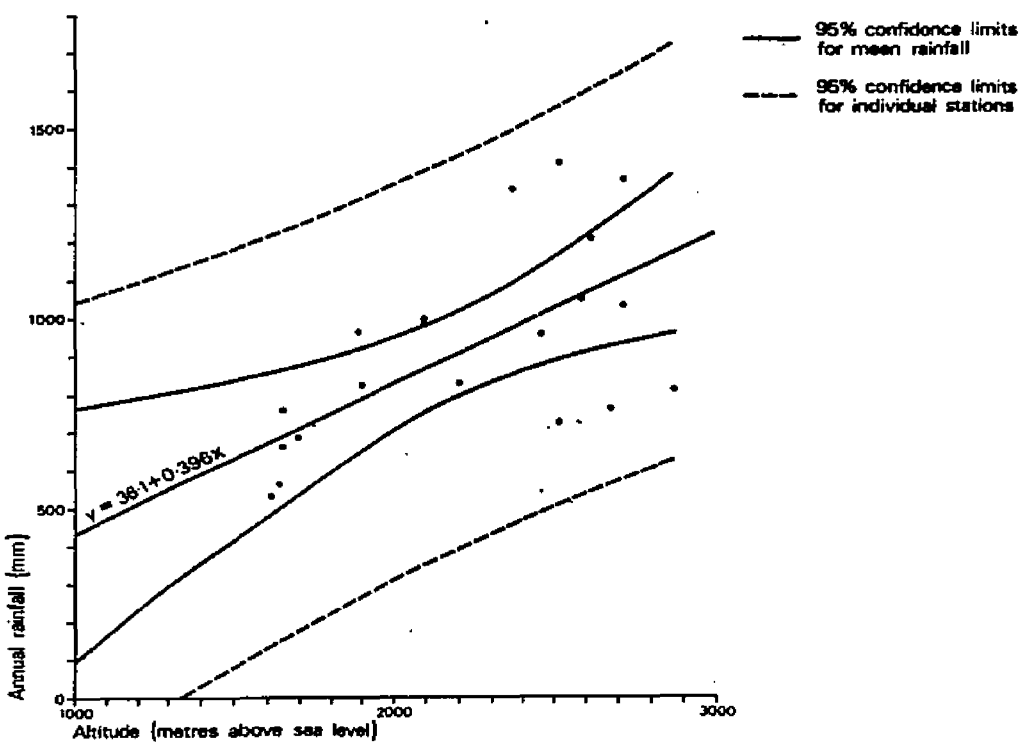
The present climate of the Rift floor is semi-arid, with pronounced wet and dry seasons, although altitude has a moderating influence on the temperatures experienced. A markedly bimodal rainfall pattern results from seasonal shifts in the convergence zones between the NE Trades and airflows from the Atlantic and Indian Oceans (Suzuki, 1967; Butzer, 1971, p. 33; Griffiths, 1972). From March to May, southeasterly winds from the Indian Ocean bring the unpredictable "small rains". After a short break in May and June come the main rains, when the Northern ITCZ lies well to the north, and humid airstreams from the Atlantic and Indian Oceans converge over the highlands. The months October to March are dry and rather windy, with winds blowing dominantly from the northeast.

Mean annual rainfall amounts recorded around the lakes range from ca. 465 mm at Langano (1600 m) to more than 800 mm on the Meki-Awash and Shala-Awasa divides (Fig. 2.4). But it is important to note that annual totals in southern Ethiopia have varied in a quasi-cyclical fashion since at least 1898, when official records began in Addis Ababa (Wood and Lovett, 1974; Kingham, 1975) (Fig. 5.14). Because of these fluctuations, the short and sometimes non-overlapping records from local stations should be regarded with caution.

Fig. 2.4a (above): Mean annual rainfall over the Ziway-Shala Basin

Fig. 2.4b (below): Rainfall/altitude regression for the Ziway-Shala Basin. (dated from Kingham, 1975)


Addis Ababa
Addis Ababa
Nazareth
Kela
Meki
Butajira
Ogelcha
Koshe
Ziway
ASELLA
Lake Tufe
Lake Ziway
Adamitulu
1200
1000
800
1400
1200
ZIWAY – SHALA BASIN
Bulbula
Wulbargh
Lake Abiyata
Lake Langano
Lake Shala
Gidu River
Meki River
Katar River
Bokoji
1000
Aje
Shashamanne
Kofele
Dodola
Dodola
0 10 20 30 40 50
kms
Watershed
Annual isohyet (mm)
Main road
Dry weather road


95% confidence limits for mean rainfall
95% confidence limits for individual stations
y = 38·1+0·396X
Annual rainfall (mm)
Altitude (metres above sea level)
0
500
1000
1500
1000
2000
3000

Nevertheless, a steady increase in rainfall with altitude is evident, reaching a maximum of 1600 mm around the 3000 m contour (Kingham, 1975). For stations up to 2900 m, the mean gradient approximates 400 mm/1000 m, and is significant at the 1% level (Fig. 2.4). On the Arussi summits, the rainy seasons are prolonged, and much of the precipitation falls as hail or snow. This simple altitudinal pattern is somewhat complicated by variations in exposure: for example the wide grassy plains of the Upper Katar ($\leqslant$ 1000 mm) lie in the rain shadow of the high shield volcanoes.

Air temperatures and open water evaporation also show strong altitudinal trends, although available data from within the basin are scanty. At Ziway (1640 m) the mean daily screen temperature is 19.3°C, and at Langano, around 20-21°C (Makin et al., 1976; T. Kingham, pers. comm. 1976). Seasonal variations are small, the lowest temperatures occurring during the main rains. The altitudinal lapse rate calculated for the Rift and the Afar is 6.75°C/ 1000 m ($r^2 = 0.972$, $p < 0.001$). This figure, which is based on data for 25 stations below 2500 m (United Nations, 1973, fig. 9) agrees reasonably well with previous estimates, which range from 5.5 to 7.0°C/ 1000 m (Fantoli, 1966; Griffiths, 1972; Bauduin and Dubreuil, 1973).

Evaporation, which is closely correlated with temperature, is referred to in more detail in chapter 9. Estimated losses from open water are 1.6 m/yr at Awasa and 2.0 m/yr at Ziway (Makin et al., 1975, 1976). These figures are quite low compared with lowland lakes such as Abhe or Turkana (Gasse, 1975; Butzer, 1971) and emphasize the influence of high altitude on the climate and hydrology of the basin.

II. Vegetation

On the Rift floor itself, the present vegetation, where still untouched by farming or fuel-gathering activities, ranges from *Acacia* woodland (Plates 14, 15) to bushed grassland.[1] Ribbons of gallery forest line the permanent rivers. The lake-margin vegetation types are described in more detail in chapter 2 E III.

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1. Terminology follows Lind and Morrison (1974).

Climbing up the mountainous flanks of the Rift, one successively encounters Combretum bushed grassland, remnants of dry, montane, Podocarpus/ Juniperus/Olea forest and finally, beginning at about 3200 - 3500 m, ericaceous scrub and Afroalpine moorland (Plate 79). Extensive montane grasslands occupy poorly drained cracking clays between 2000 and 2800 m, particularly in the Katar Basin (Makin et al., 1975).

III. Soils

The distribution of soils within the basin is known only in outline. The map shown in Fig. 2.5, which is based on a reconnaissance survey by L.R.D. (Makin et al., 1975; King and Birchall, 1975), reveals several aspects of present surface soils which are relevant to hydrology and to palaeo-environmental interpretations.

The major soil types in the Rift Valley clearly show the influence of soil parent material and extent of weathering. They tend to be aligned in broad belts parallel to the Rift axis. The central belt of skeletal or weakly developed soils (Tv) has formed from lacustrine sediments, river alluvium and pumice. Infiltration tests show that many profiles on unconsolidated deposits are highly permeable, and likely to generate little surface runoff (Makin et al., 1976).

Rising ground bordering the Rift floor typically bears heavier, more strongly developed soils overlying weathered ignimbrite, pumice, or basalt. Although King and Birchall (1975) mapped these as eutric or chromic cambisols (Be and Bc) profiles measured west of Ziway and Meki (chapter 5C) showed clear evidence of clay translocation, and layers or nodules of carbonate at depth (Plate 5). Such soils can be tentatively classified as orthic luvisols (FAO, 1974) or brown calcimorphic soils (Young, 1976). Whatever their classification, the presence of clay-rich B horizons, and often calcrete as well, may considerably restrict permeability (Makin et al., 1976).

At higher altitudes, the distribution of deep, reddish latosols (Ne) closely mirrors the former extent of montane forest (Tato, 1970; Riché and

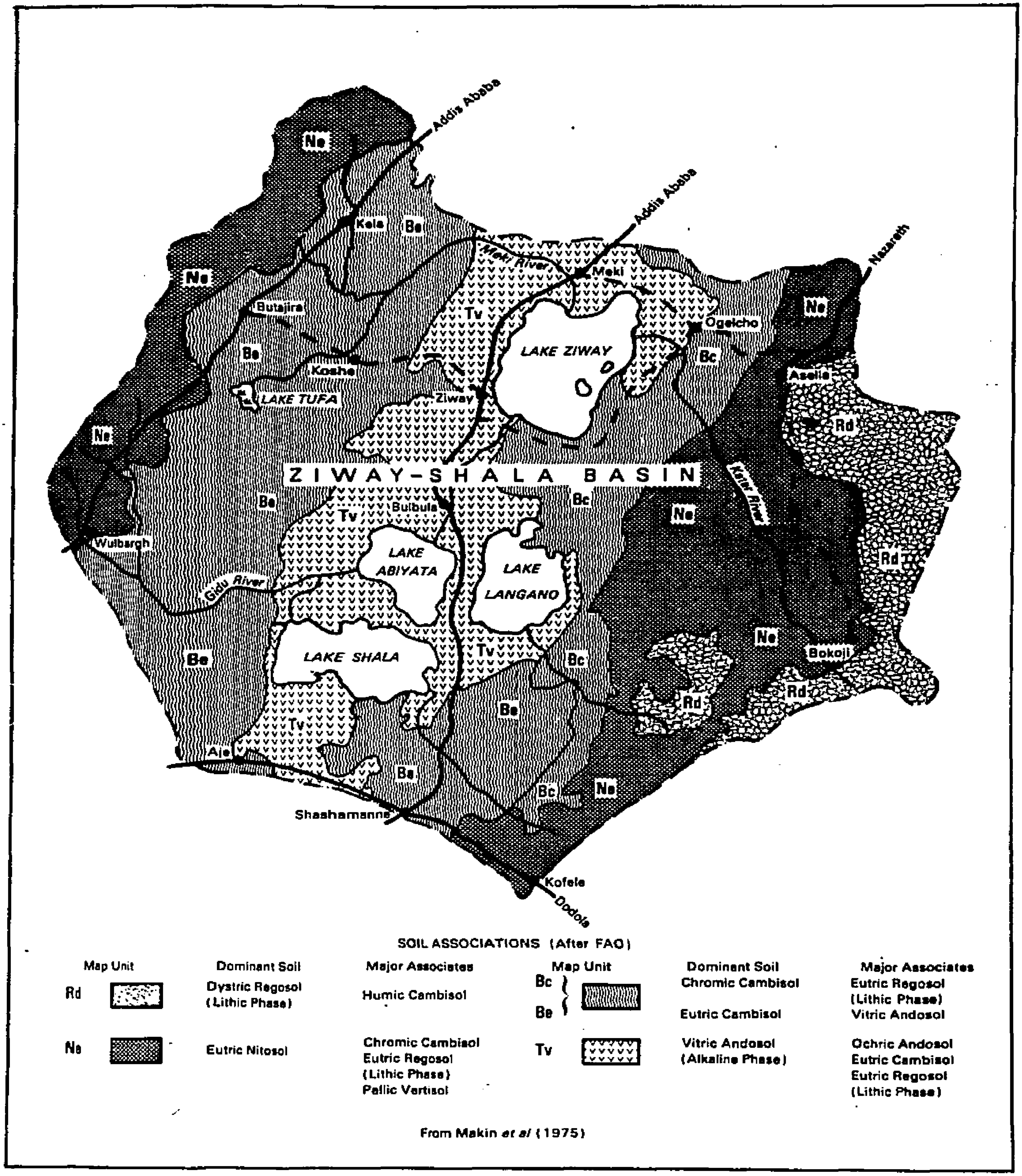


Fig. 2.5: Soil associations in the Ziway-Shala Basin.

Ségalen, 1969). They occur widely on both escarpments and on the foothills of the Arussi Mountains, but are best developed on rolling terrain underlain by basaltic parent materials. Clay contents are high (up to 70% in the subsoil) and the presence of clay pans at depth may further limit infiltration. In poorly drained areas, such as the central Katar plains, the red soils give way to dark cracking clays (pellic vertisols) or gleyed soils, whereas on the Arussi volcanoes, they pass upwards into well-drained montane soils. Unfortunately, the high mountain regions remain an almost total blank on the map. Riché and Ségalen classified the soils on the Galama Ridge (near the upper limit of the Ericaceous Belt) as andosols, whereas the FAO Soil Map of Africa (1970) portrays them as dystric regosols (Rd). Whatever their exact status, these soils are thin, and occupy quite steep slopes under high rainfall.

This rather striking banded soil distribution reinforces the other altitudinal controls on runoff generation (chapter 2C). What is perhaps less obvious is that the main soil types also show a trend in age from the youngest soils on lacustrine sediments or recent volcanics to the deep, red soils of the escarpments. Hence there is a more than coincidental resemblance between Fig. 2.5 and Fig. 5.1, which shows the extent of the Holocene 1670 m lakes. Surface soils are often useful as a check on shoreline limits derived by other methods (chapter 5C). The latest period of slope stability and soil formation on the Rift flanks (areas Be and Bc) appears to correspond closely to the early/mid-Holocene lacustral interval (chapter 8).

2C HYDROLOGY

The modern-day Rift lakes are fed mainly by perennial rivers from the highlands. Although direct rainfall on the lake surfaces may account for 20 to 25% of the inputs, subsurface inflows are probably of small overall importance, and net gains or losses of groundwater to other basins can be neglected for the purposes of this study.

The largest catchments are those of the Katar (3400 km^2), which rises above 4000 m on the slopes of the Arussi Mountains, and the Meki, which drains 2300 km^2 of the western escarpment and foothills. Surface runoff is highly seasonal. Both rivers reach maximum discharges during the main rains, with a secondary, unreliable peak in March to May. On average 78% of the Katar flow into Lake Ziway occurs between July and October (Table 2.1). Baseflow during the dry seasons is maintained either by high-altitude springs (Wenner, 1967) or by permanent swamps like Lake Tufa (Fig. 2.1).

Less information is available on the southern rivers. Lake Langano is maintained largely by five smaller streams draining the forested escarpment to the southeast. The runoff pattern resembles the Katar but the streams may dry up in the winter months. Unfortunately, the discharge data collected by CADU at 10 stations from April 1970 to April 1972 (Wenner, 1973b) appear to be too low by a factor of 10 or more when compared with the A.V.A. published data for the Gedemso River or for the outflow from Lake Langano (Table 2.1) (T. Kingham, pers. comm., 1976), and so must be viewed with suspicion.

The situation with regard to Lake Shala is even less satisfactory. Its surface inflows come from two sources: the perennial Adabar River, which enters from the eastern escarpment, and the main branch of the Gidu River, whose basin adjoins the Meki to the south. Neither has been gauged. The other distributary of the Gidu, the Gorgeza, is merely a rainy-season spillway into Lake Abiyata (Fig. 2.1), but happens to have been gauged (Table 2.1).

The longterm average contribution of all the rivers to the lakes is probably about 1500 mcm/yr, which is greater than the discharge of the Awash 50 km above Lake Abhé (Gasse and Street, 1978a). It is important to note that the figures given in Table 2.1 refer to 1969-73, when runoff was unusually high (chapter 5 B III).

Only a small proportion of the total surface runoff is currently derived from the Rift floor, where periods of water deficit are longest (Brown and Cochemé, 1973) and infiltration capacities generally high. The UNDP hydrogeologists have suggested that the critical altitude for runoff generation is about 1700 m, based on a comparison of annual rainfall data with Thornthwaite potential evapotranspiration estimates (United Nations, 1973). The latter are

Table 2.1: Mean monthly and annual river discharges in the Ziway-Shala Basin during the period 1969 to 1973 (mcm)

River	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total
Katar[1] at Ogelcho	6.69	7.32	19.63	15.76	16.35	12.31	62.17	145.33	90.17	30.07	8.49	6.07	420.36
Meki[1] at Meki Town	6.93	12.93	31.61	31.91	21.61	21.91	78.92	124.52	68.95	23.81	8.12	5.38	436.60
Gedemso[2] near Lake Langano	0.71	1.07	4.47	5.24	4.17	4.49	17.00	20.67	15.06	7.03	1.76	0.78	82.45
Gorgeza[1]	0.69	0.68	1.40	0.49	0.72	0.75	2.15	3.41	2.50	0.81	1.17	0.81	15.58
Bulbula[1]	14.98	10.12	10.16	8.16	7.10	5.40	8.08	24.03	45.84	43.63	20.42	12.93	210.85
Horocallo[1]	3.86	2.53	1.82	1.42	1.23	0.91	1.06	4.80	12.74	15.54	7.39	4.36	57.66

Sources:

1. Makin et al. (1976).
2. Kingham, pers. comm. 1976 (A.V.A. data, for 1969-72 only).

mcm = million cubic metres.

not very reliable in East Africa (Dagg and Blackie, 1970). A check for Addis Ababa confirms that the more rigorous Penman formula yields much higher results (Brown and Cochemé, 1973), implying that the critical altitude lies rather higher, at about 2000 m.

Runoff coefficients and their possible variations through space are considered in more detail in chapter 9.

The importance of groundwater inflows into the lakes is harder to assess. They originate in two main ways: as meteoric waters which descend from the highlands through the faults and fissures of the escarpment zones; or as geothermal waters. The cold, escarpment-foot springs appear to contribute only a small proportion of the discharge into the lakes, and the residence time of their waters underground is quite short (Wenner, 1970, 1973a; Lloyd, 1977).

Geothermal waters well up to the surface in three main areas: on Tulu Gudu Island, Lake Ziway; in O-itu Bay, North Langano; and surrounding Lake Shala (Fig. 2.6). No springs occur near Abiyata. Their output and chemistry have been extensively investigated by UNDP and the Preussag Company (United Nations, 1973; Baumann et al., 1975). Discharges for all measured outlets are listed in Table 2.2. Baumann and his co-workers

Table 2.2: Discharges of measured hot springs in the Ziway-Shala Basin

Area	Measured flow (l/s)	No. springs measured	Estimated annual flow (mcm/yr)
Shala	108.67	18	3.43
Chitu Haro	6.0	9	0.19
N. Langano	38.87	15	1.23
Tulu Gudu	1.8	3	0.06
		Total	4.90

From United Nations (1973).

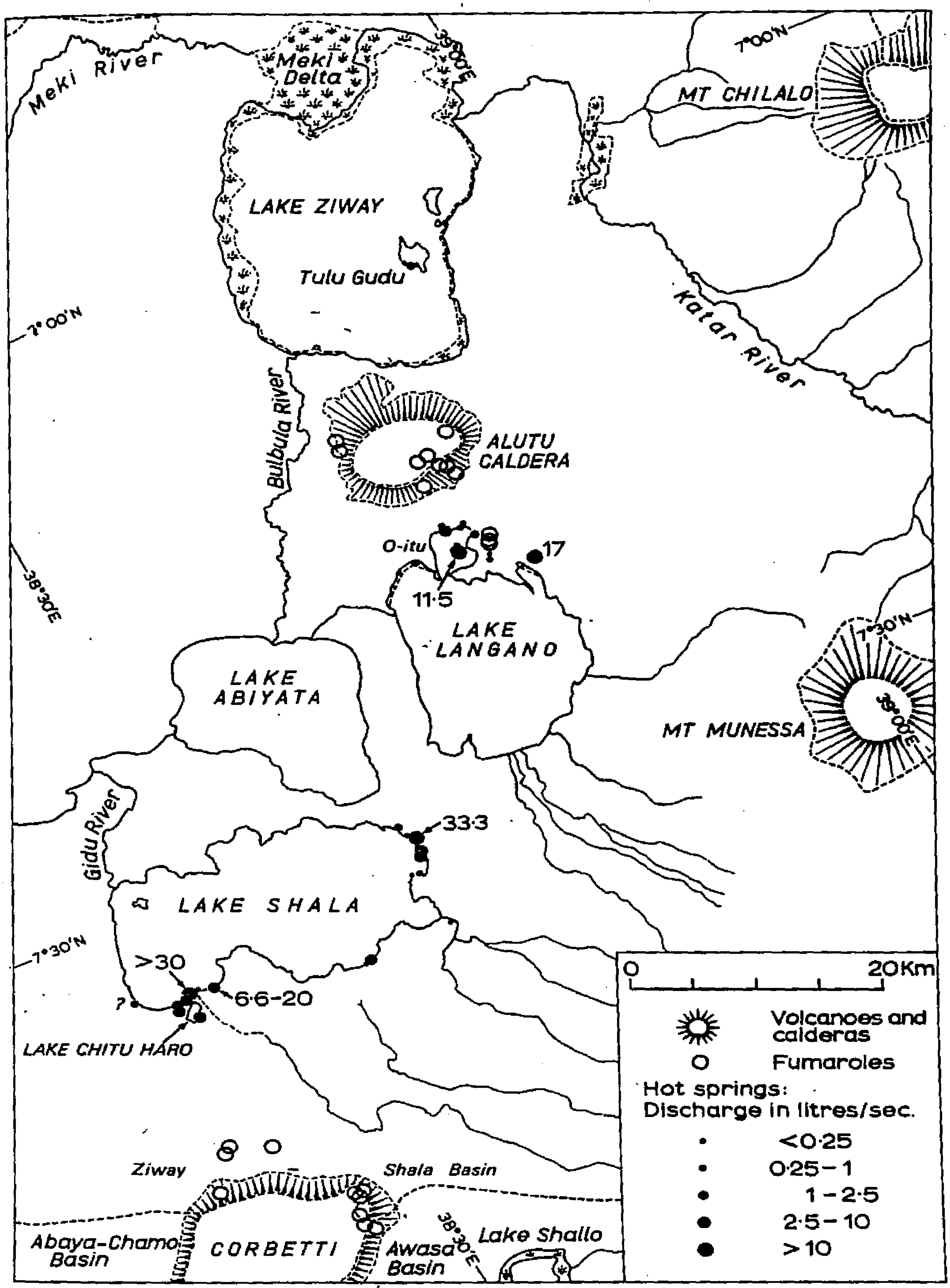


Fig. 2.6: Distribution of hot springs and fumaroles in the Ziway-Shala Basin (data from United Nations 1973).

concluded from their chemical analyses that the hot springs originate through varying degrees of mixing of meteoric waters with a high-temperature groundwater reservoir, and in fact stronger flows have often been observed after the main rains (Mohr, 1971, p.229).

Despite the care with which the subaerial springs have been recorded, a major question mark hangs over the importance of groundwater inflows into Shala and Langano, and their contribution to the water budget of Lake Shala in particular. For Ziway and Abiyata, rainfall and surface inflows balance losses by evaporation and outflow (McKerchar and Douglas, 1974) so closely that groundwater flows can safely be neglected. But it seems certain that Shala is partly dependent on groundwater, because the inflowing streams are small, and the lake level relatively stable. The same conclusion has also been reached on the grounds of the similarity between the lake-water chemistry and the springs (United Nations, 1973): this is not a major obstacle to palaeohydrologic calculations using C. Washbourn-Kamau's method (1967b) provided that all inflows originate within the basin and maintain approximately constant proportions through time.

At this point we need to ask whether the surface and subsurface catchments are coincident. The possibility of deep seepage from the higher Awasa Basin northwards towards Lake Shala, and from Lake Ziway into the Awash, cannot be ruled out, although positive evidence is slight. In the case of Ziway, the strongest argument against it is the high predictive success of the surface water-balance model by McKerchar and Douglas (1974). Lake Awasa, however, is surprisingly fresh for a terminal lake, which could be explained by seepage through the lake bed, a process also believed to operate in Lakes Naivasha and Baringo (United Nations, 1973; Richardson and Richardson, 1972; Beadle, 1974). Fortunately for our purposes, both the L.R.D. team who were considering the water balance of the lake (Makin et al., 1975), and Baumann et al. (1975) who studied the chemistry of the geo-thermal waters, have rejected the idea of large subsurface losses from Awasa. As a first approximation, therefore, the Ziway-Shala catchment will be assumed to be hydrologically, as well as topographically, closed.

2D PHYSIOGRAPHIC SETTING

I. Introduction

In the present relatively arid period, the Galla Lakes form two disjunct hydrological systems (Fig. 2.1). The general physiographic setting of the lakes is shown in Fig. 2.7 and Plates 6-7. Ziway and Langano overflow into Abiyata whereas the lowest lake, Shala (1558 m), lies in a separate terminal basin. A rise in the level of Lake Abiyata to 1592.0 m, only 13.8 m above its 1972 highstand, would unite the two systems.

As a result of their different tectonic and hydrological settings, the four modern lakes vary considerably in their depth, turbidity and salinity, providing a range of modern analogues for past conditions. Fortunately, too, their bathymetry (Fig. 2.8) and water chemistry (Table 2.3) are comparatively well known, as a result of a preliminary survey during the Italian Occupation (Brunelli et al., 1941) and subsequent work by Addis Ababa University (Baxter et al., 1965), Italconsult (1970), UNDP (United Nations, 1973), the Preussag Company (Baumann et al., 1975) and L.R.D. (Makin et al., 1976). The available map coverage has recently improved dramatically with the publication of the U.S. Defense Mapping Agency 1:250,000 series, the D.O.S. 1:50,000 toposheets, and a partial series of 1:20,000 basemaps prepared for the Ethiopian Ministry of Mines.

We shall first consider the modern lake basins and then the area inundated by the Holocene highstand.

II. Lake Ziway

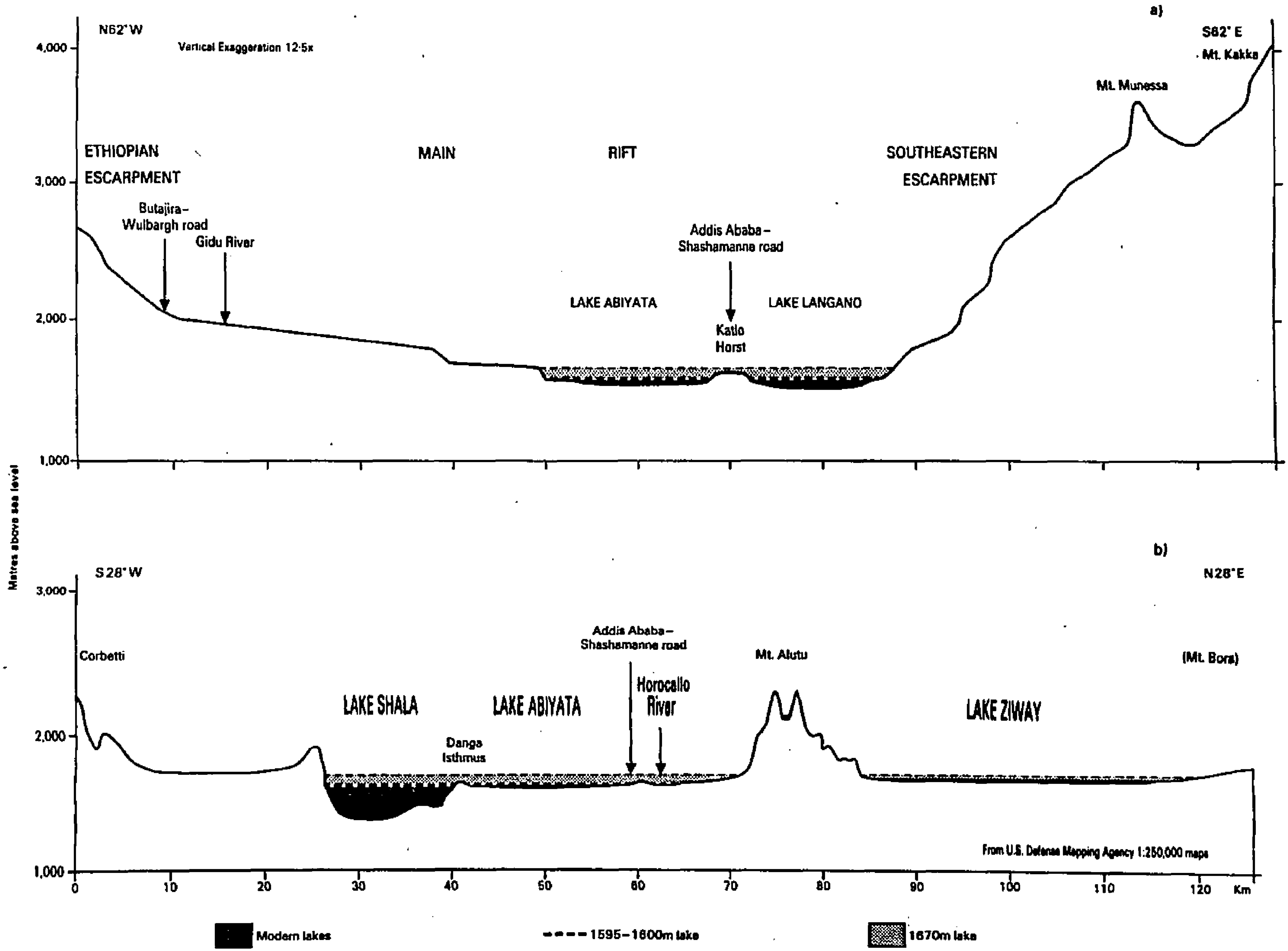
Lake Ziway (1636 m), the highest and most extensive of the lakes (Table 2.3), lies in a shallow downfaulted basin flanking a large basalt field (Fig. 2.2). Eleven islands of varying size, formed of scoria or hyaloclastite, rise above lake level, the maximum depth so far encountered being only 9.55 m (Fig. 2.8, Plate 8).

The Meki and Katar Rivers, which enter from the northwest and eastern sides, have both deposited late Holocene deltas in the lake. The Meki Delta

Fig. 2.7: Topographic cross sections of the Ziway-Shala Basin

a) WNW/ESE

b) SSW/NNE


a)
N62°W
S62° E
Vertical Exaggeration 12·5x
Mt. Kakka
Mt. Munessa
ETHIOPIAN
ESCARPMENT
MAIN
RIFT
SOUTHEASTERN
ESCARPMENT
Butajira–
Wulbargh road
Gidu River
Addis Ababa–
Shashamanne road
LAKE ABIYATA
LAKE LANGANO
Katlo
Horst
4,000
3,000
2,000
1,000
Metres above sea level
b)
S 28°W
N28°E
3,000
Corbetti
Addis Ababa–
Shashamanne road
Mt. Alutu
(Mt. Bora)
Horocallo
River
LAKE SHALA
LAKE ABIYATA
LAKE ZIWAY
Danga
Isthmus
2,000
1,000
From U.S. Defense Mapping Agency 1:250,000 maps
0
10
20
30
40
50
60
70
80
90
100
110
120
Km
Modern lakes
1595–1600m lake
1670m lake

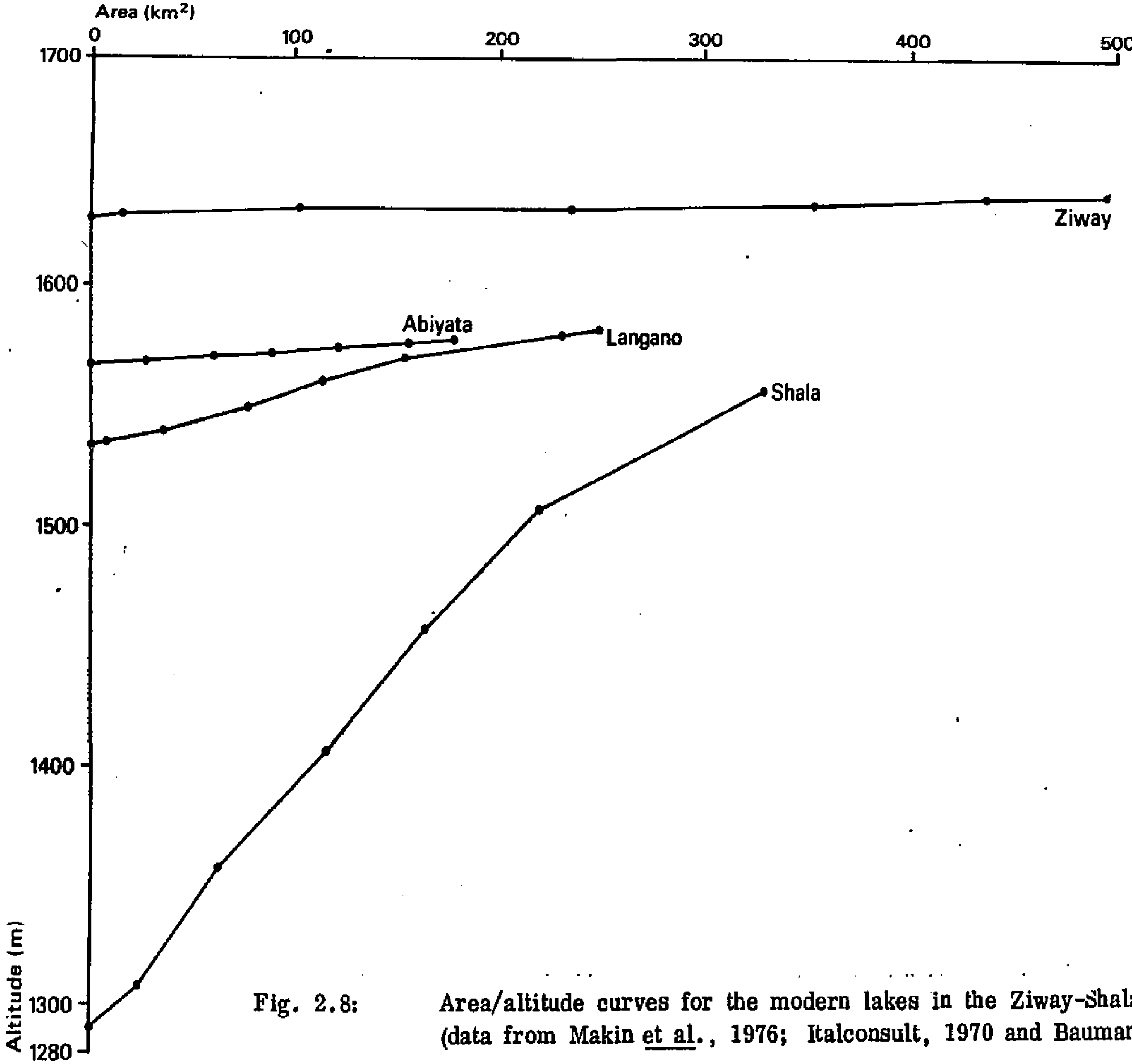


Fig. 2.8: Area/altitude curves for the modern lakes in the Ziway-Shala Basin (data from Makin et al., 1976; Italconsult, 1970 and Baumann et al., 1975).

Table 2.3: Physical and Chemical Characteristics of the Southern Ethiopian Rift Lakes (surface samples)

Lake	Surface Altitude (m.a.s.l.)	Maximum Recorded Depth (m)	Area (Km^2)	Volume (Km^3)	Estimated Turnover Time (y)[1]	Total Dissolved Solids (‰)	pH.	Dominant Ions[2]	Carbonate Alkalinity (meq/l)	Number of Samples	Salinity Classification: Talling & Talling (1965)[3]	Salinity Classification: Gasse (1975)
Ziway *	1636	9.55	442	1.6	1.6	0.29	7.9	Na>Mg, Ca:HCO_3	3.7	15	Class I	Oligohaline
Langano *	1582	47.9	241	5.3	3	1.27	9.1	Na:HCO_3 > Cl	14.6	10	Class II	Oligohaline
Abiyata *	1578	14.2	175	1.1	9.5	8.09	9.4	Na:HCO_3 > Cl	102.3	6	Class III	Mesohaline
Shala *	1558	266.0	329	36.7	56	16.7	9.7	Na:HCO_3 > Cl	162.2	11	Class III	Mesohaline
Chitu Haro **	1558	n.d.	0.8	n.d.	n.d.	32.1	10.2	Na:HCO_3 > Cl	401.7	1	Class III	Polyhaline
Awasa +	1680	21.6	92	1.3	n.d.	0.61	8.7	Na:HCO_3	8.4	8	Class II	Oligohaline
Abaya +	1285	13.1	1070	8.2	n.d.	0.61	8.0	Na:HCO_3	9.2	2	Class II	Oligohaline
Chamo +	1233	12.7	350	n.d.	n.d.	0.65	8.6	Na:HCO_3	10.0	1	Class II	Oligohaline
Chew Bahir +	520	Ephemeral				8.10	9.2	Na:Cl > HCO_3	51.6	1	Class II/III	Mesohaline

n.d. no data

1. Given by mean depth/depth of water lost annually (by evaporation and outflow).
2. 'HCO_3' includes $CO_3^=$ and HCO_3^-
3. Class I Low salinity, conductivity < 600 µmhos/cm
   Class II Moderate salinity, conductivity 600 - 6000 µmhos/cm
   Class III Very saline, > 6000 µmhos/cm

Sources:

* Gasse and Street (1978a) and Wenner (1967).

** Baumann et al. (1975).

\+ Grove, Street and Goudie (1975) and Vatova (1941), updated from United Nations (1973); Makin et al. (1975); Bentor (1967).

($60\ km^2$) has been mapped in detail by Makin et al. (1976) (Fig. 2.9). Gentle levées (formed of sandy clay loams) and basin clays at the apex give way to sandy flood channels bordered by seasonally waterlogged, fluviolacustrine clays on the distal margin. The latter grade lakewards into permanent swamps. The Katar Delta ($10\ km^2$) is less complex and does not show a pattern of old distributaries. Both are deltas of the "fluvial-wave interaction" type (Collinson et al., 1978).

Since Ziway receives the bulk of the freshwater discharge, its waters are dilute, with a mean dissolved solid content of 0.29 g/l (Table 2.3). Lake level is maintained by a lava threshold at 1635.6 m just north of Adamitulu, and the outflowing stream, the Bulbula, descends some 57 m over a distance of 35 km between Lakes Ziway and Abiyata. In the central part of its course, it has incised a gorge more than 50 m deep into unconsolidated sediments (Plate 9).

III. Lake Langano

Like Ziway, Langano occupies a largely faultbounded depression. The main body of the lake is 47.9 m deep (fig. 2.8) and is underlain by at least 88 m of sediments (Wenner, 1967). Its northeast and southwestern shores, which lap against the horsts and grabens of the Wonji fault belt, are steep and indented, whereas at the north end a small circular caldera, O-itu (North Bay), has been flooded by the lake to a depth of 8.85 m (Plates 10-11).

The lake waters are coloured red-brown by fine suspended sediment brought in by streams from the escarpment (Plate 6). More concentrated than Ziway, the lake contains 1.27 g/l of dissolved solids (Table 2.3), although it overflows seasonally into Abiyata via the Horocallo River. However the annual outflow is only a quarter of the volume carried by the Bulbula (Table 2.1) and the channel of the Horocallo is less incised ($\leqslant$ 11 m).

IV. Lake Abiyata

Abiyata, which is the smallest of the lakes, lies in a saucer-shaped hollow (Fig. 2.8) within a deep, downfaulted trough almost infilled by an

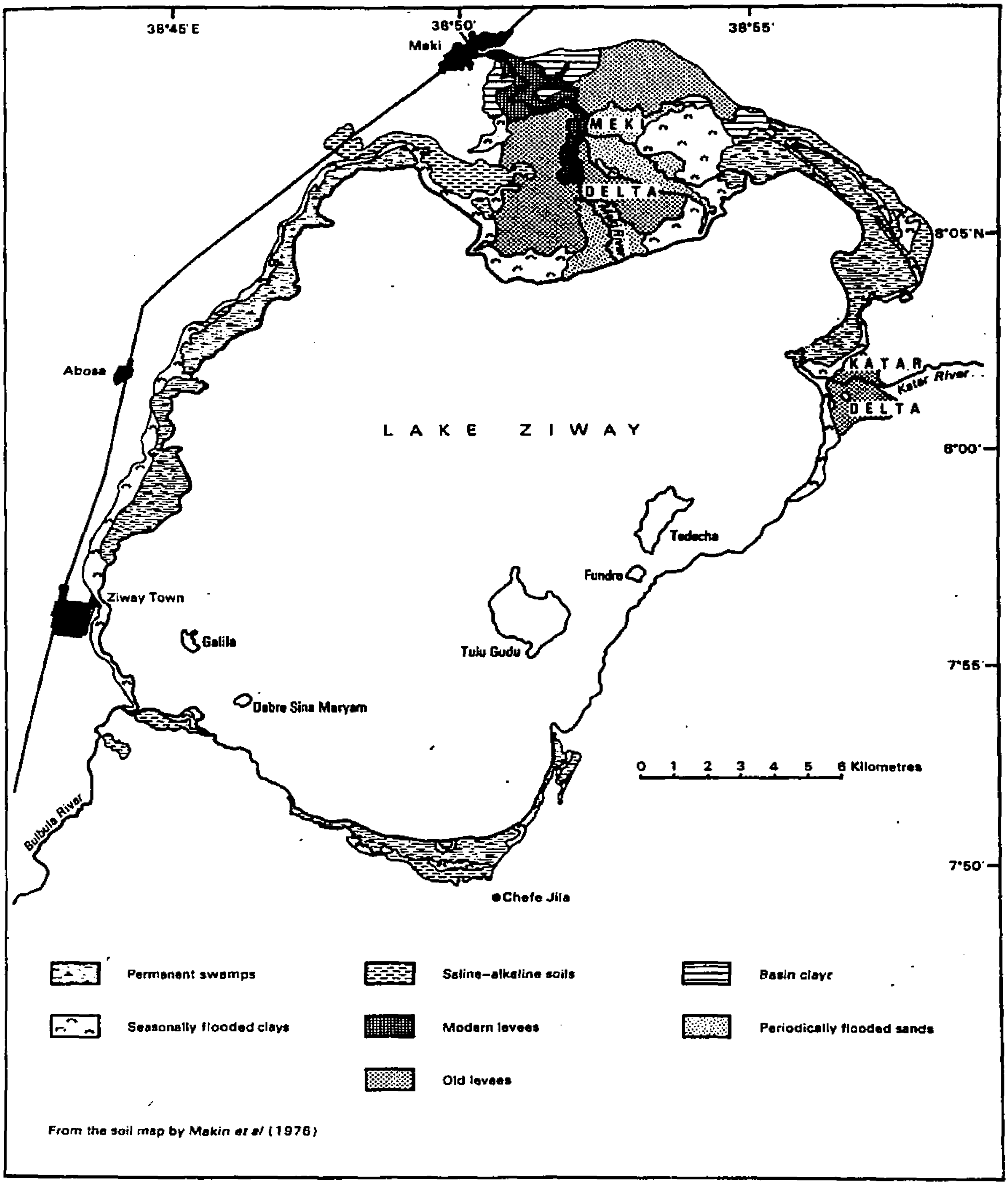


Fig. 2.9: Marginal environments around Lake Ziway (simplified from Makin <u>et al.</u>, 1976).

estimated 580 m of Quaternary deposits (Searle and Gouin, 1972). It is separated from Langano by the Katlo horst (Mohr et al., 1975) (Plate 6). Along its northern and northwestern shores, the lake abuts entirely on soft sediments (Plate 12).

Abiyata can essentially be regarded as an evaporating basin – and in keeping with its terminal position, the water level has fluctuated up and down by about 8 m in the last 40 years, although the greatest depth recorded is only 14.2 m (Figs. 5.12 – 5.14). The lake is presently surrounded by a rather melancholy belt of drowned Acacia tortilis trees, which dates from a very abrupt rise in 1967-70 (Plate 13). Its waters are sodic and strongly alkaline, containing an average of 8.09 g/l dissolved solids. They are more dilute during highstands, for example in 1938 (Talling and Talling, 1965).

V. Lake Shala

Lake Shala possesses the distinction of being the deepest lake in the Eastern Rift System of Africa, partly owing to its youth. Its volume is nearly five times greater than the other three main lakes combined (Fig. 2.8). The central basin is a very large, funnel-shaped caldera whose sides plunge steeply to depths exceeding 250 m (Mohr, 1967; United Nations, 1973). To north and south, cliffs of ignimbrite, rhyolite and pumice rise sheer to heights of 300 m above the water surface (Plates 2, 7). However, in the less steep northeastern and western embayments, the slopes overlooking the lake are veneered by up to 15 m of Late Pleistocene and Holocene lake sediments (Plate 14).

Although Shala has no outlet, its level seems to have fluctuated by only 5-6 m since 1933 (chapter 5) and may be partially stabilized by groundwater inflows, as suggested above. The dissolved solid content of the surface water amounts to some 15.7 g/l, largely sodium carbonate and bicarbonate, making this the most alkaline of the present Galla lakes apart from the tiny spring-fed lake Chitu Haro (Table 2.3).

VI. The bed of the Holocene united lake

If the outflows from Ziway and Langano were to increase, Abiyata would rise until it flooded the Gidu-Gorgeza isthmus and overflowed into Shala. This would cause a rapid drop in salinity in the Abiyata Basin. From then on, if inputs continued to exceed evaporation, Shala would increase in depth till it merged with Abiyata, shortly afterwards engulfing Langano, and eventually Lake Ziway (Fig. 2.7).

Obviously the former lake floor has been modified by the deposition of up to 15 m of Holocene sediments, but its general morphology is easy to reconstruct, consisting of a northern and a more extensive southern basin, separated by a topographic high at about 1652 m in the central Bulbula Plain (Plate 15). This threshold, which is called Hondola, must have greatly affected the configuration of the united lake. Gèze (1974) has suggested that it marks a major E-W fault, but if so then the faultscarp lies buried beneath at least 50 m of sediments (Fig. 6.1). It seems more likely that the threshold is related to the deposition of pyroclastics downwind from Alutu.

The area-altitude curve of the present basin floor is shown in Fig. 8.2. The main point to note is the very rapid increase in evaporating surface with depth once the three southern lakes are united.

2E LIMNOLOGY AND LACUSTRINE SEDIMENTATION

I. Physical limnology

Although the climate of the Rift floor is somewhat modified by altitude, all the lakes are essentially tropical in character, since surface water temperatures average 24-25°C (Vatova, 1941; Baxter et al., 1965). A distinctive feature of tropical lakes is that their water circulation, and hence productivity, tend to be largely determined by wind stirring, and secondarily by any inputs of saline spring water, rather than by the seasonal march of radiation and temperature (Beadle, 1974).

Wave action tends to prevent persistent stratification of the lake waters and is responsible for the sorting of modern beach deposits. The four large lakes have maximum fetches ranging from 19.5 km (Abiyata, NW/SE) to 31.5 km (Ziway, NE/SW), which is sufficient to generate quite high waves: breaker heights of 0.5 m or more are common on southwest Langano during windy nights in the dry season. Although detailed measurements of wind speeds are not available, a simple wave-prediction approach suggests that the depth of the wave-mixed layer in all four lakes will be less than 6-8 m under average conditions ($<$ Force 3), although it could increase to 20 m or more in gale-force winds ($>$ Force 8) (from Smith and Sinclair, 1972).

The mean residence time of the lake waters varies from 1.5 to 56 years (Table 2.3). As expected, the shallow lakes Ziway and Abiyata are polymictic, whereas complete mixing occurs less frequently in Langano and Shala (Baxter et al., 1965). Although Baumann et al. (1975) found that the Shala deep water below 70 m was 5°C colder and 15% more saline than the surface water, there is still sufficient mixing to allow dissolved oxygen to reach the deepest parts of the lake (Vatova, 1941). Persistent deoxygenation would tend to prevent stirring of the bottom sediment by mud-dwelling animals such as chironomid (midge) larvae (McLachlan and McLachlan, 1971, 1976), thus allowing the preservation of seasonal laminations in the sediment (Ruttner, 1963). It is therefore of some interest that the only rhythmically laminated fossil diatomites occur on the slopes overlooking this lake (chapter 6D).

II. Water chemistry

All five lakes fall into the sodium bicarbonate-carbonate-dominated category of Kilham (1971a) (Table 2.3). Electrical conductivity, pH, total dissolved solids and carbonate alkalinity all increase from the highest lakes towards Shala and Chitu Haro. Most of the major ions like Na^+ and Cl^- show high positive correlations ($p < 0.001$) with each other and with total carbonate (Fig. 2.10). This also holds for the hot springs. The two exceptions, Ca^{++} and Mg^{++}, are discussed below.

The evolution of water chemistry at the present day, which will be used to interpret past conditions, is portrayed in Fig. 2.11a-d. These are modified Schoeller diagrams (Schoeller, 1962), on which chemical constituents are plotted on the horizontal axis and their concentrations in meq/l on the

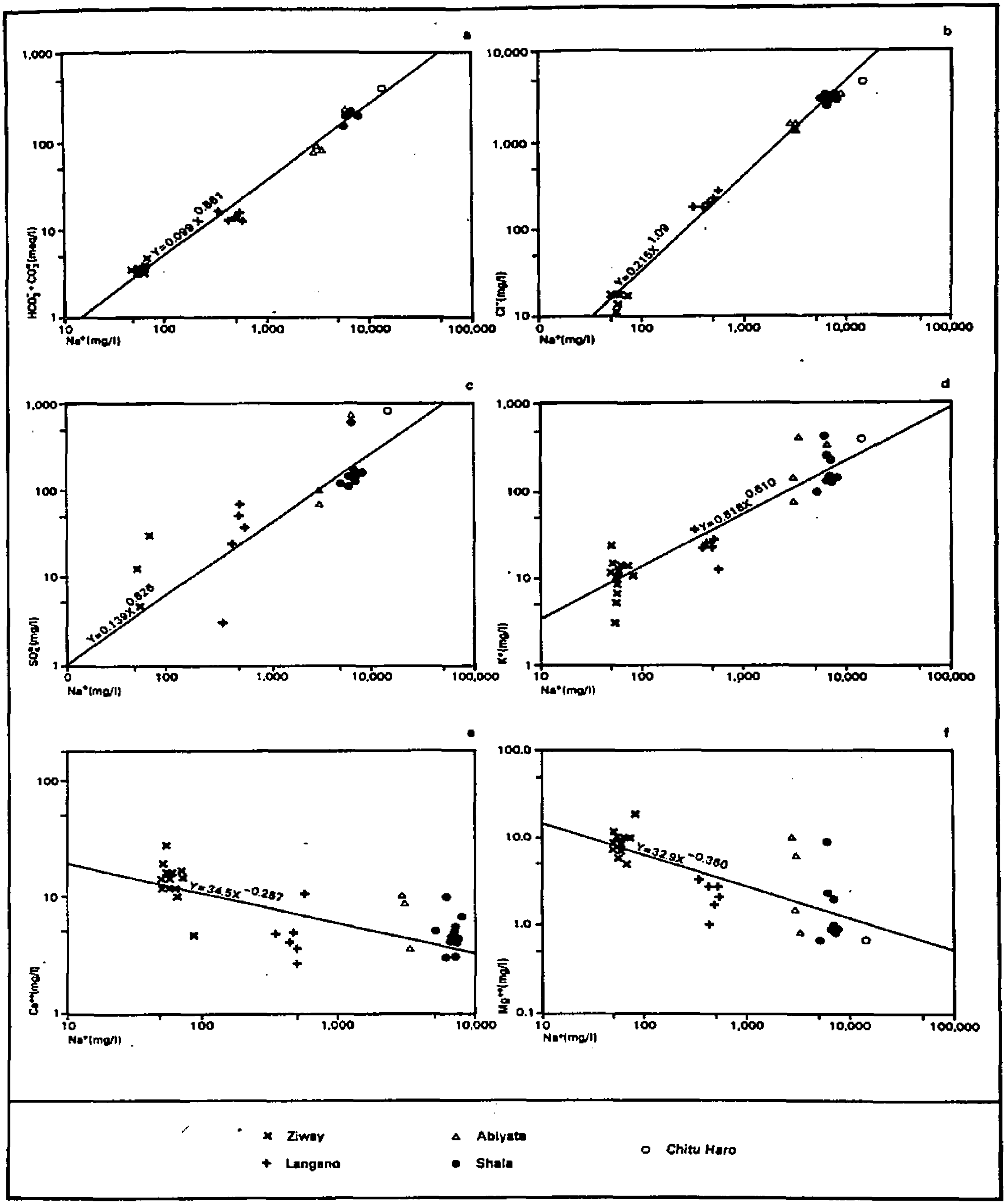


Fig. 2.10: Correlations between major ions in the modern lake waters. All relationships shown are significant at the 0.001 level. Data sources are listed in Table 2.3.

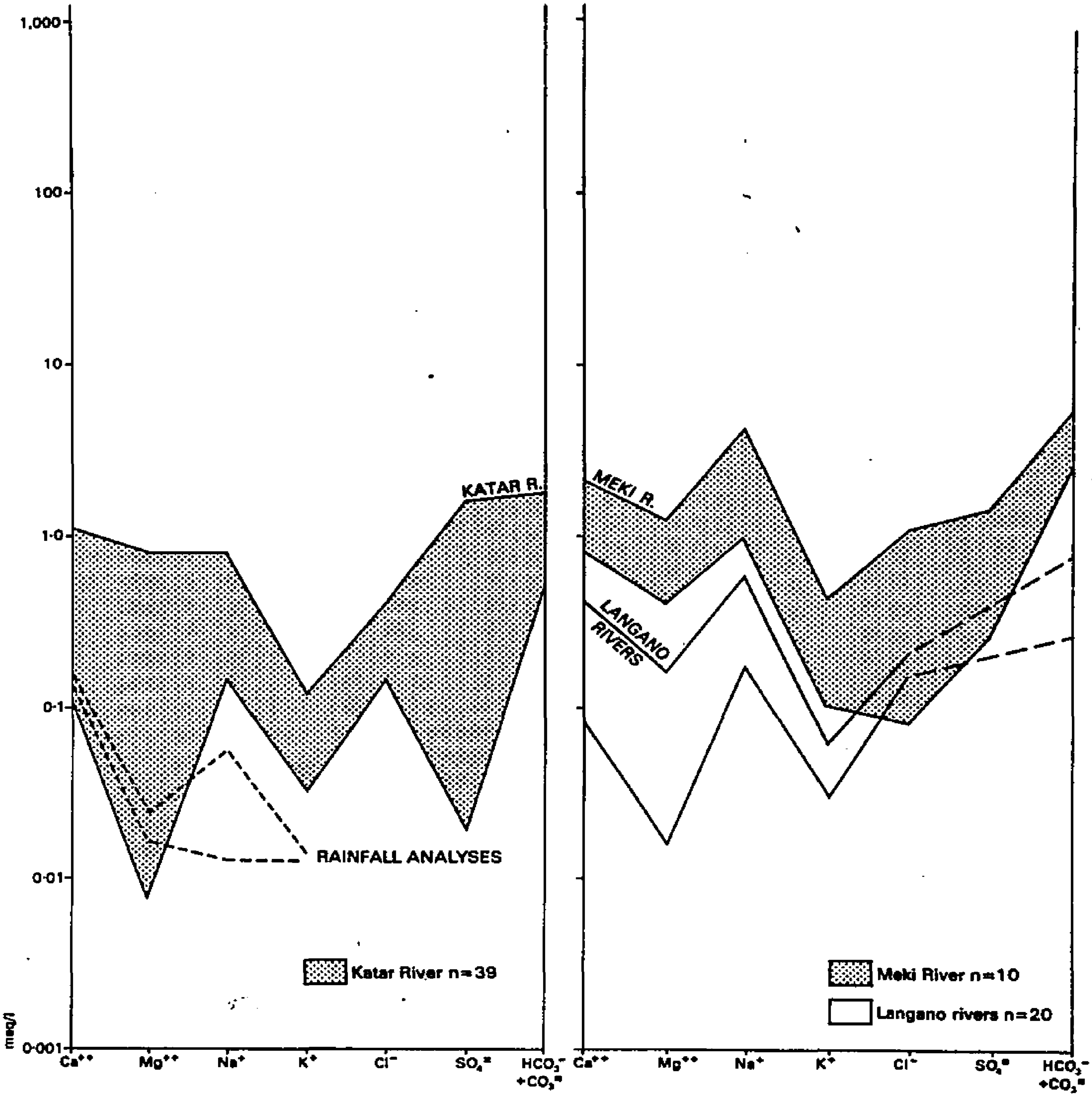
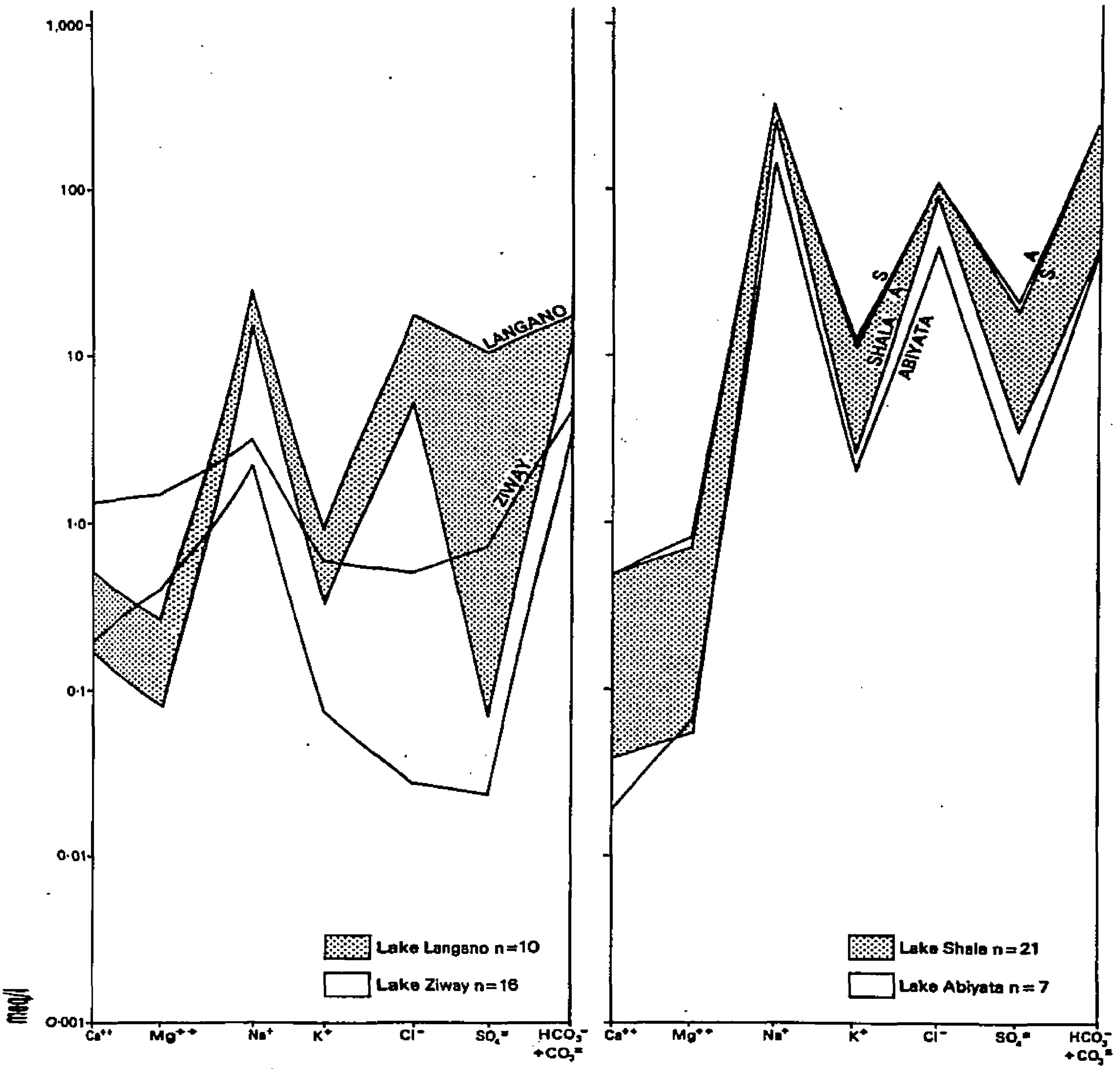


Fig. 2.11: Schöeller diagrams for modern surface waters in the Ziway-Shala Basin. Data sources are listed in Tables 2.3 and 2.4.

a) (left) Rainfall at Asella, Katar River
b) (right) Meki River, Langano Rivers
c) (overleaf, left) Lakes Ziway and Langano
d) (overleaf, right) Lakes Abiyata and Shala



1,000
100
10
1·0
0·1
0·01
0·001
meq/l
LANGANO
ZIWAY
Lake Langano n=10
Lake Ziway n=16
Ca^{++}
Mg^{++}
Na^{+}
K^{+}
Cl^{-}
$SO_4^{=}$
HCO_3^{-} $+CO_3^{=}$
S
A
SHALA
ABIYATA
Lake Shala n=21
Lake Abiyata n=7

Table 2.4: Na^{+}/Ca^{++} equivalent ratios for rainfall, rocks, surface and ground-waters in the Ziway-Shala Basin

	No. anals.	Range of Na^{+}/Ca^{++} values
Rainfall (Asella)	2	0.0902 - 0.343
Bedrock		
Basalts and hawaiites	4	0.269 - 0.509
Rhyolitic trachytes	2	1.48 - 4.25
Pantelleritic ignimbrites	3	5.30 - 22.0
Pantelleritic lavas and pumice	6	10.2 - 26.1
River waters		
Headwaters of the Katar	36	0.361 - 1.86
Katar	3	0.350 - 1.00
Meki	9	0.826 - 3.90
Langano tributaries	20	0.435 - 3.38
Bulbula	10	1.00 - 7.70
Horocallo	2	32.8 - 113
Groundwater (boreholes not in thermal areas)		
Katar Basin (plateau)	11	0.308 - 1.29
Mid-Meki Valley	6	0.630 - 2.67
Rift	12	1.01 - 4.39
Hot springs		
Ziway	2	13.9 - 43.6
Langano	17	5.12 - 1500
Shala	40	8.15 - 2010
Chitu Haro	9	46.1 - 212
Lake waters		
Ziway	14	1.77 - 15.9
Langano	6	61.2 - 175
Abiyata	5	250 - 6720
Shala (surface)	6	513 - 6850
(deep water)	9	1030 - 1510
Chitu Haro	1	4880

Sources:
Rainwater: Wenner (1973a)
Bedrock: Di Paola (1972a,b); Mohr (1970)
River water: Makin *et al*. (1976); Wenner (1970, 1973b); Pittwell (1967); United Nations (1973); F.A. Street (unpubl.)
Groundwater: Wenner (1973a); Makin *et al.* (1976); United Nations (1973)
Hot Springs: United Nations (1973); Baumann *et al* (1975); Pitwell (1967)
Lakes: Listed in Table 2.3.

vertical, semi-logarithmic axis. Na^{+} and K^{+} have been shown separately, since they are very abundant. This format facilitates comparison of ratios between various elements in solution. In waters of common derivation, lines joining the various elements are parallel. In addition, Na^{+}/Ca^{++} equivalent ratios calculated from rock and water analyses in the basin have been listed in Table 2.4.

Rainfall, which contains 2-3 mg/l Ca^{++} but only 0.5 - 1.0 mg/l Na^{+}, must be an important calcium input into the lakes (Fig. 2.11a). Its composition is modified by a number of important processes:-

1) Contact with soil and rock minerals, and with the clastic load of streams, leads to an increase in total ionic concentration, and to a shift in ionic proportions. The incongruent solution of silicate minerals can be schematically represented as follows:

$$\text{Cation Al-silicate} + H_2CO_3 + H_2O = HCO_3^- + H_4SiO_4 + \text{cation} + \text{Al silicate(s)}$$

(Stumm and Morgan, 1970, p.391).

These reactions give rise to the high Na^{+}, HCO_3^- and SiO_2 contents, typical of drainage waters from peralkaline silicate rocks (White et al., 1963; Garrels and Mackenzie, 1967; United Nations, 1973; White and Claassen, 1975). In this way, variations in bedrock geology and soil type exert considerable influence on runoff concentration and composition. The streams draining the well-watered Katar and Langano catchments are significantly more dilute than the Meki River ($p < 0.001$) (Table 2.5). The Katar headwaters, flowing off

Table 2.5: Average dissolved load of major rivers

River	Mean concentration $\pm 1\sigma$ (mg/l)	No. anals.
Meki	224 $\pm$ 53	9
*Katar	82 $\pm$ 22	39
*Langano rivers	76 $\pm$ 34	20

* Not significantly different at the 0.05 level.

Sources: listed in Table 2.4.

the trachyte shield volcanoes, are rather less sodic than surface runoff from the Meki and Langano Basins (Fig. 2.11, Table 2.4). Ground waters, except in the thermal areas, reflect the ionic composition of river waters but are more concentrated and marginally richer in calcium (Table 2.4).

2) Evaporative concentration in rivers and lakes increases the ionic strength of the water and also its carbonate alkalinity (Garrels and Mackenzie, 1967). As the pH rises above 9, silica becomes increasingly soluble (Krauskopf, 1967, p. 168), with the result that concentrations of more than 100 mg/l are often recorded in Abiyata and Shala.

3) Ca^{++} and Mg^{++} show an absolute decrease in amount from the inflowing rivers towards Shala and Chitu Haro (Fig. 2.11). Most of the reduction occurs in passage through Lake Ziway. Ca^{++} and Mg^{++} show their strongest negative correlations with pH ($p < 0.001$). This is easiest to explain in the case of calcium. Carbonate ion is present in increasing proportions above pH 8, due to the reaction:

$$HCO_3^- + OH^- \rightleftharpoons CO_3^= + H_2O \quad ,$$

thus tending to cause calcite ($CaCO_3$), which has a low solubility product, to come out of solution (Talling and Talling, 1965; Krauskopf, 1967; Morgan and Kalk, 1970; Hutchinson, 1975). The processes involved are discussed in more detail in chapter 7DIII. Magnesium is incorporated in small amounts into the calcite lattice (Lippmann, 1973) but can also take part in clay-mineral synthesis. The net result is an extreme depletion of both elements in the waters of Shala and Chitu Haro (Fig. 2.11).

4) Biological sequestration of solutes is likely to be most important in the case of SiO_2, which is utilized by diatoms (Kilham, 1971b), and $CaCO_3$, which is required by mollusca (Beadle, 1974, p. 164) and ostracods. Although molluscs are only known to occur in Lake Ziway, significant calcite extraction by ostracods may take place in Abiyata and Shala. Removal of dissolved P, N and Ca^{++} by fringing swamps, which has been observed in Lake Chad and the White Nile swamps (Beadle, 1974; Viner, 1975) has never been studied in an Ethiopian context, but cannot be discounted in Lake Ziway.

5) Silica and certain ions, notably K^+ and Mg^{++}, may be fixed by the formation of lacustrine clays (Roche, 1977), particularly Mg-smectite and illite, which are both recorded from Lake Shala (Baumann et al., 1975). Authigenic zeolites, which would preferentially remove Na^+ and Ca^{++} (Hay, 1966) have not been observed forming in the present lakes. Trona ($NaHCO_3 . Na_2CO_3 . 2H_2O$) is being deposited around the shores of Lake Shala, but in negligible amounts.

6) The most uncertain factor is the influence of the hot springs. These have a gross ionic composition very similar to the adjacent lakes, but Ca^{++} and Mg^{++} are relatively less depleted (United Nations, 1973; Baumann et al., 1975). Variations in hot spring activity through time could therefore be one factor leading to changes in carbonate deposition (chapter 7D III).

The combined result of all these processes is a tendency towards very high concentrations of Na^+, $CO_3^=$ and Cl^- in the lowest lakes. Although the UNDP and Preussag teams both suggested that the salinities of Abiyata and Shala are derived from the springs, it seems more likely that all surface waters in the basin are undergoing a convergent evolution governed primarily by bedrock composition, evaporative concentration, and the atmospheric partial pressure of carbon dioxide; as suggested by Talling and Talling (1965) for Lakes Magadi and Natron. No firmer conclusions can be reached without much more sensitive chemical and isotopic tracing. It does seem clear that during highstands we can expect the lakes to have been much more dilute and to reflect the composition of rain and river water more closely.

III. Biology and marginal habitats

Because of the high alkalinity of the more southerly lakes, Ziway is the only one to be bordered by a well-developed belt of marginal swamp, which ranges from 100 m to 2 km in width (Fig. 2.9, Plate 16). These wetlands appear to be the closest analogue for the freshwater lake-margin environments at times of highest lake-level.

The present vegetation zonation around Ziway is controlled by seasonal lake-level fluctuations, as shown in Table 2.6. There is also a close relationship with substrate. Away from the Meki and Katar Deltas, where deposition of heavy organic clays is occurring, the lake-floor sediments consist of inert pumice sand which is overlain by about 0.3 m of clay on the higher-lying, occasionally flooded Cynodon grassland and by thin organic loams in papyrus, Typha and Juncus swamps near the edge of the lake (Makin et al., 1976).

Table 2.6: Vegetation zones around the shores of Lake Ziway (after Makin et al., 1976, p. 154)

Site	Plant community
Open water under sheltered conditions	Floating grasses (Plate 8) and *Nymphaea caerulea* (water-lilies)
Lake edge: permanent shallows	*Cyperus papyrus* (papyrus) and *Typha latifolia* (cattail) swamp
Lake edge: seasonally inundated	
- 1636 and 1637 m contours closely spaced:	*Aeschynomene elaphroxylon* (ambatch) thicket with sheltered lagoons (Plate 16)
- Elsewhere:	*Typha* thicket with *Juncus* and *Panicum repens*
Lakeshore with high watertable:	
- unaffected by alkali:	*Cynodon plectostachyus* (Bermuda) grassland
- affected by alkali:	*Sporobolus spicatus* grassland
- strongly influenced by alkali:	Bare ground
Hinterland with seasonal high watertable	*Acacia albida* woodland with *Ficus*, *Croton macrostachys* and *Acacia tortilis*

Around the other lakes, aquatic vegetation is limited to a discontinuous belt of cattail (Typha), backed by wet meadows of Cynodon and the alkali-tolerant Sporobolus (Bolton, 1969) (Plate 11).

The present microflora of the lakes is poorly known (Gasse, 1975). Although fossil diatom studies have helped to back up the conclusions of chapter 8B, they have been based largely on modern distributions from the Afar and East Africa. Ziway has yielded an oligohalobous diatom assemblage dominated by littoral species with high nutrient requirements. Abiyata, like many shallow soda lakes, supports a very productive phytoplankton, consisting largely of green algae (Oocystis sp.) (Baxter et al., 1965) with some alkaliphilous diatoms (Navicula elkab and Nitzschia spp.). The productivity of Lake Shala is difficult to assess, since diatom frustules are at present being dissolved on the lake bed, but the flora appears to consist of salt- and alkali-tolerant diatoms with a preference for eutrophic waters: Cyclotella meneghiniana is commonest in littoral environments and Nitzschia spp. in the phytoplankton. Lake Chitu Haro supports dense blooms of blue-green algae.

The only elements of the lake faunas which are really useful for environmental reconstructions are mollusca and fish. Gastropods have so far only been found alive in Lake Ziway and some of the freshwater rivers.. A list of species is given in Table 2.7. The apparent lack of bivalves, especially the tiny pisidia, may well be due to inadequate collecting. The specimens of Bellamya unicolor, Corbicula pusilla and Unio dembeae collected by O. Neumann from the Bulbula (Thiele, 1933) were almost certainly subfossil.

Two major factors can be suspected to influence the present distribution of mollusca: the rise in salinity of the lakes southwards and the depletion of Ca^{++} to potentially limiting levels (Beadle, 1974, pp. 51-2). It is not impossible that tolerant species may be found in Lake Langano, since the gastropods Melanoides tuberculata, Cleopatra pirothi, Gabbiella rosea and Ceratophallus natalensis are believed to live in Lake Turkana, which has an alkalinity of about 20-25 meq/l (Ferguson, in litt. 19.7.76; Van Damme and

Table 2.7: Freshwater mollusca collected alive in the Ziway-Shala Basin

	Ref.
MEKI RIVER	
Bulinus "truncatus" (Audouin)	1
Biomphalaria pfeifferi ruppelli (Dunker)	4
KATAR RIVER	
Succinea hararensis Connolly	
HULUKA RIVER (tributary of L. Langano)	
Ancylus regularis Brown	3
LAKE ZIWAY	
Melanoides tuberculata (Muller)	2,4
Lymnaea natalensis (Krauss)	1,2,4,5
Bulinus "truncatus" (Audouin)	2,4,5
B. forskali (Ehrenberg)	2,5
Afrogyrus coretus (De Blainville)	4
Ceratophallus natalensis (Krauss)	1,2,4,5
C. bicarinatus (Mandahl-Barth)	4
Gyraulus costulatus (Krauss)	1,4
Segmentorbis (Segmentorbis) angustus (Jickeli)	2,4
Biomphalaria sudanica (Martens)	2,4,5
B. pfeifferi ruppelli (Dunker)	1
BULBULA RIVER	
Melanoides tuberculata (Muller)	4
Lymnaea natalensis (Krauss)	4
Bulinus "truncatus" (Audouin)	1,4
B. forskali (Ehrenberg)	1
Ceratophallus natalensis (Krauss)	4
Biomphalaria pfeifferi ruppelli (Dunker) (in small pond near entry into lake)	1

References:
1. Connolly (1928)
2. Brown (1965)
3. Brown (1973a)
4. Goll and Aweitu (1974a)
5. Goll and Aweitu (1974b)

Taxonomy follows Brown (1965), Brown and Wright (1972) and Brown and Mandahl-Barth (1973).

Gautier, 1972; Beadle, 1974). However, Brown (in litt., 4.1.71) believes that the worn, opaque shells collected in 1970 from the south shore of Langano by A.T. Grove were reworked from older deposits.

Fish species are even fewer in number, and only the large ones have interested collectors. Notwithstanding, there appears to be a clear decline in diversity from Ziway to Shala which can be attributed directly or indirectly to water composition. Lake Ziway supports the richest fauna: five species of cyprinids (Barbus intermedius, B. ethiopicus and possibly the hybrid B. microterolepis, Garra quadrimaculatus, and G. makiensis) plus the cichlid Tilapia nilotica (Banister, 1973; Tedla, 1973). Fishermen on Langano catch abundant T. nilotica, B. intermedius and an unidentified catfish. The only large fish known to breed in Abiyata and Shala is the salt-tolerant T. nilotica; and in the latter it has only been caught in the dilute waters at the mouth of the Adabar River.

IV. Sedimentation

a. Sources of clastic sediment

The modern Rift-floor sediments have only been studied to a very limited extent. Until now, interest has centred on Lake Ziway, because of its potential for irrigation (Italconsult, 1970; Makin et al., 1976) and on Lake Shala, where heavy-metal deposits were anticipated (Baumann et al., 1975). A thorough study of depositional environments was beyond the scope of this investigation. The data given below are gathered from published reports, supplemented by a limited amount of new information. They nevertheless provide some useful clues to the origin of older deposits.

Today all the lakes seem to be characterized by predominantly clastic sedimentation. In this they resemble the chemically very similar North Basin of Lake Turkana (Yuretich, 1975) rather than brine lakes such as Magadi (Surdam and Eugster, 1976), Manyara (Stoffers and Holdship, 1975) and Asal (Gasse and Stieltjes, 1973) in which evaporites and other unusual authigenic minerals are forming.

Clastic inputs enter the lakes in five principal ways:-

1) in perennial rivers such as the Meki, the Katar and the Gidu (Plate 7).
2) from ephemeral gullies entrenched into the rift-floor sediments.
3) by wind transport.
4) through wave erosion of rocky cliffs and pre-existing deposits.
5) by direct ash falls onto the lake surfaces.

1) Modern rates of fluvial and aeolian transport are thought to have been greatly accelerated by overgrazing and deforestation (Tato, 1970). The Meki carried 3.7×10^5 tonnes of suspended material into Ziway in 1963-64 (United Nations, 1965, p. 93); corresponding to a minimum erosion rate of 161 $t/km^2/y$. At this rate the lake will be filled up in 11,200 years. 94.5% of the suspended load was carried between July and September. Judging from the size of its Holocene delta, the Meki has carried the largest volume of sediment over the long term, followed by the Gidu, but this is misleading because the Katar probably drops much of its bed load in the swamp-filled Shetemata graben east of Ziway (Lloyd, 1977). The Katar's suspended load comprises silts and fine sands (Fig. 2.12a). A single sample of Meki River bed load, collected 20 km upstream from Meki Town, consisted of moderately sorted (σ_ϕ 0.944), medium to coarse pumice sand (Fig. 2.12b).

2) Large amounts of sand and gravel are carried during the rains by ephemeral streams, especially those which are entrenched into unconsolidated lacustrine and pyroclastic deposits, for example at the foot of Alutu (chapter 6C). The gully-floor deposits (Fig. 2.12b) are poorly to very poorly sorted sands and gravels (σ_ϕ 1.78 - 2.75), and characteristically contain a mixture of rock types of different densities, such as obsidian, ignimbrite and pumice.

3) Little is known about the magnitude of aeolian inputs. Very high local rates of soil deflation resulting from overgrazing have been recorded in the Rift near Dera (Tato, 1970), but the relative contribution of long-distance dust transport by the NE Trades (Turekian, 1968, fig. 2.7) is quite unknown. The latter may conceivably be an important source of calcium carbonate (Goldberg and Griffin, 1970). The locally transported dust seems

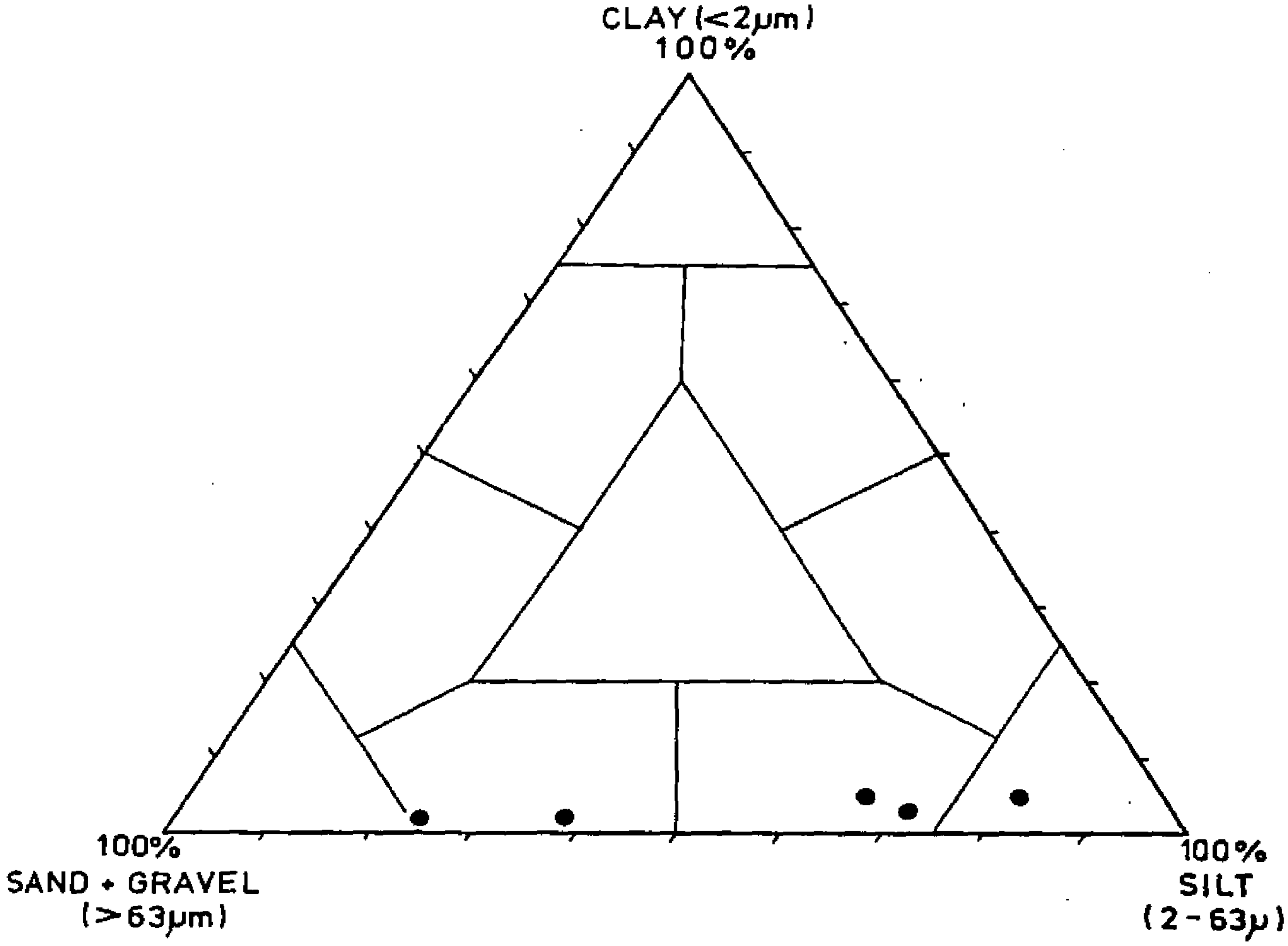
Fig. 2.12: Grain-size distributions of recent sediments in the Ziway-Shala Basin

a) (above) Suspended load, Katar River (data from Wenner, 1973a)

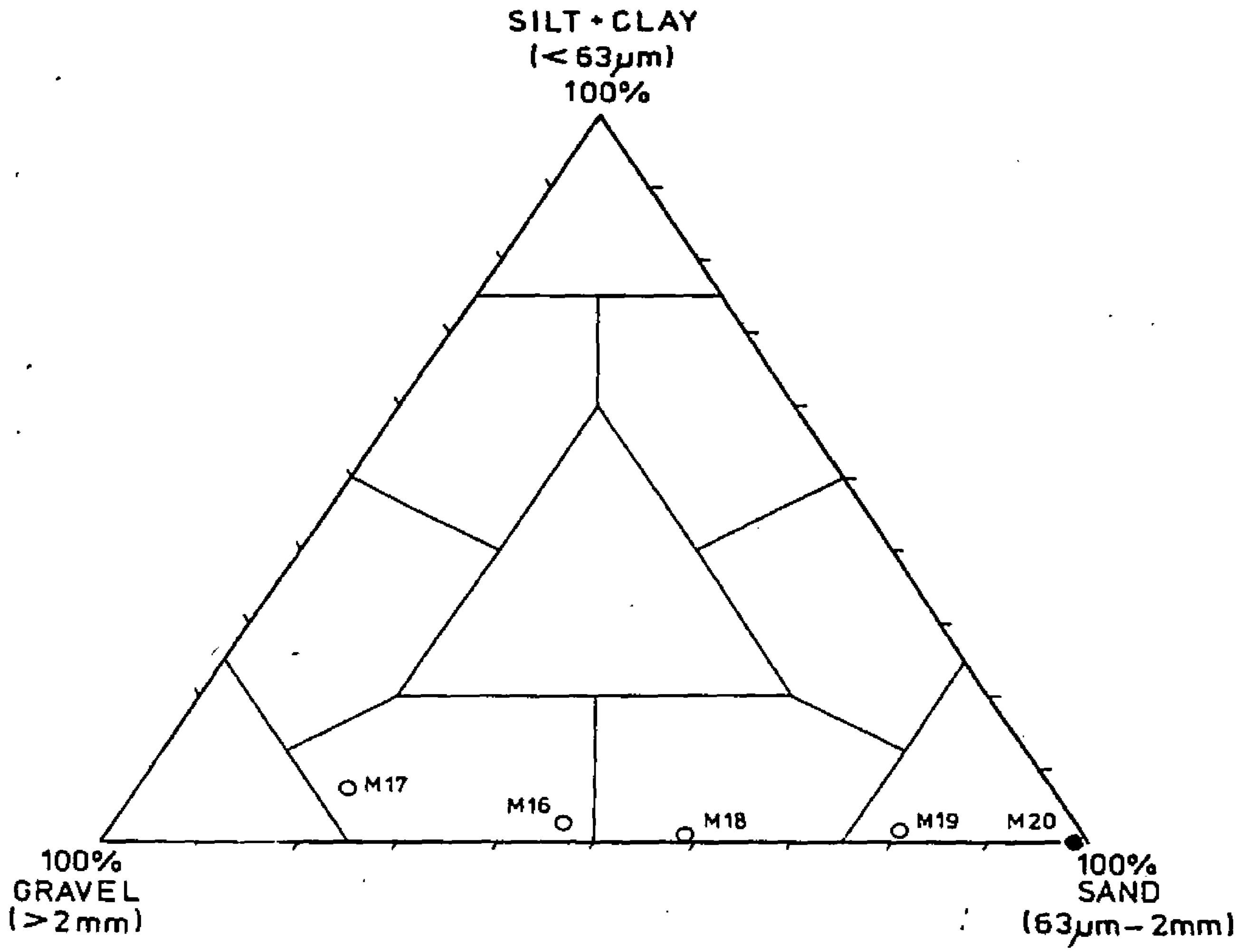
b) (below) Coarse-grained fluviatile sediments (Appendix 7a)

Open circles: gully-floor gravels

Black dots: Meki River bedload.



CLAY (<2μm)
100%
100%
SAND + GRAVEL
(>63μm)
100%
SILT
(2-63μ)



SILT + CLAY
(<63μm)
100%
M17
M16
M18
M19
M20
100%
GRAVEL
(>2mm)
100%
SAND
(63μm-2mm)

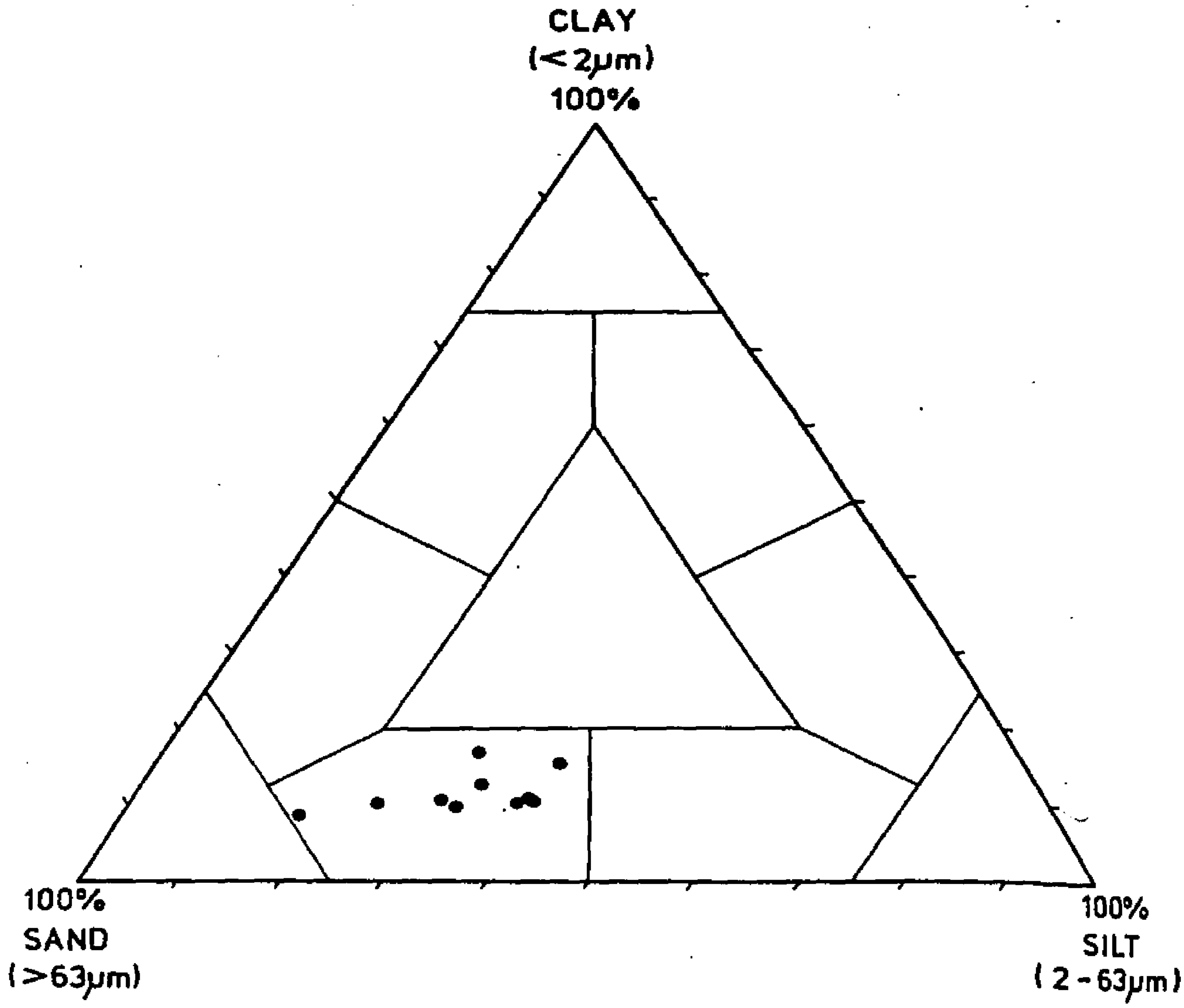
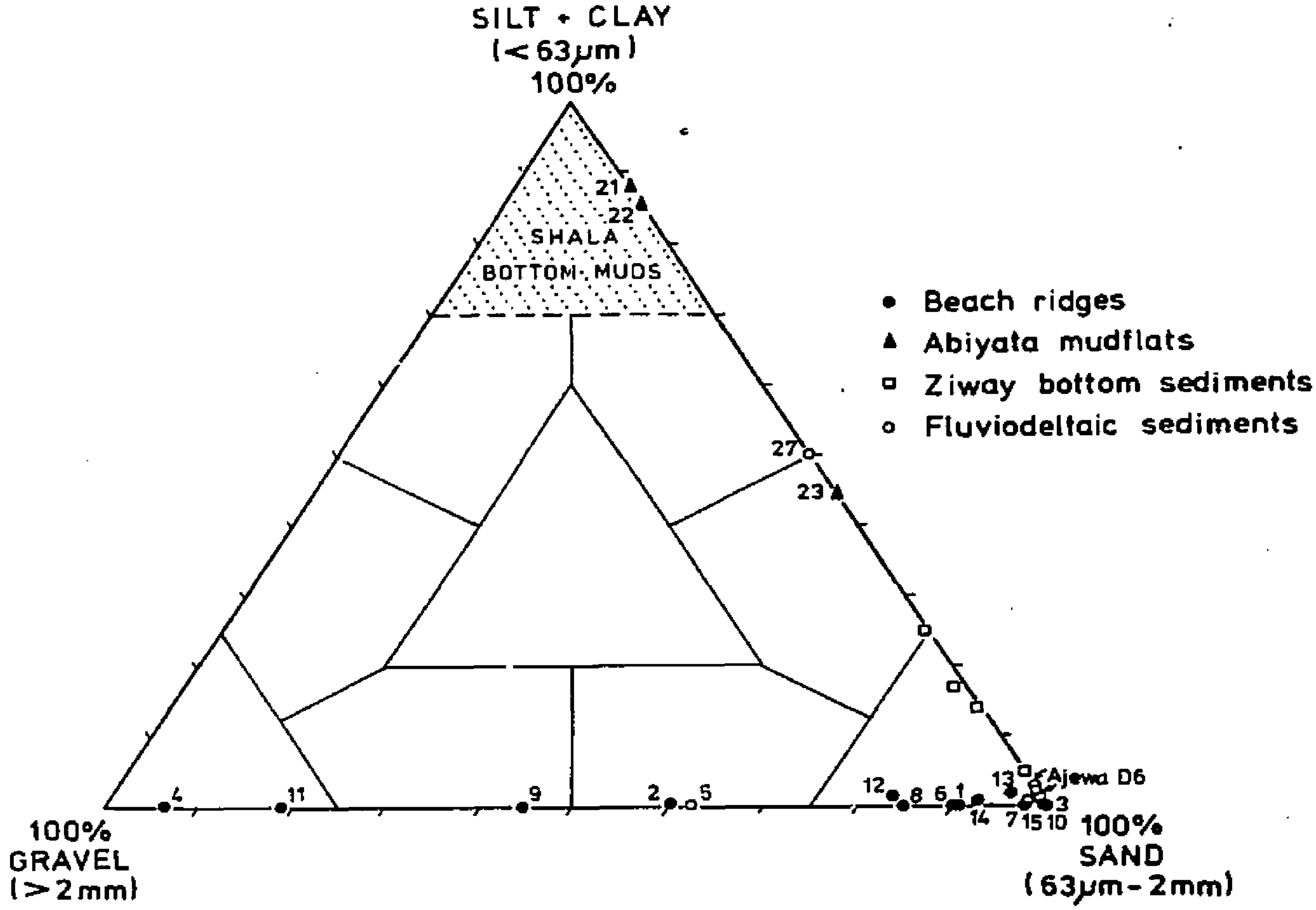


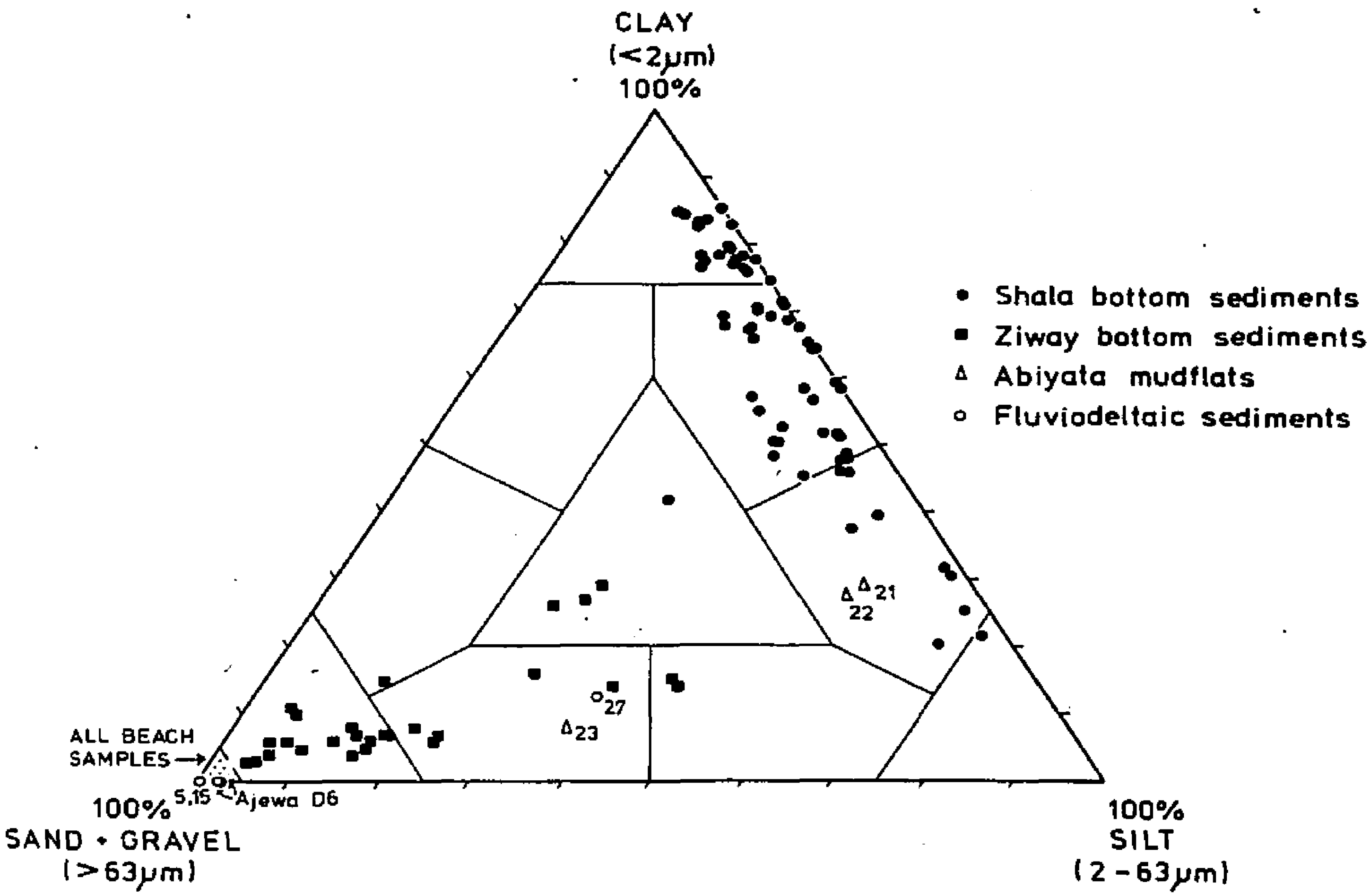
Fig. 2.12 (contd.): c) Wind-blown soil material, Rift floor near Dera (dated from Tato, 1970).

Fig. 2.12 (contd.):

d) (above)
Coarse-grained lacustrine and fluvio-lacustrine sediments (data from Appendix 7a (samples M1-15, M21-23, M27, Ajewa D6) and Makin et al., 1976 Lake Ziway)).

e) (below)
Fine-grained lacustrine and fluvio-lacustrine sediments (data from Appendix 7a (samples M5, M15, M21-23, M27, Ajewa D6); Italconsult, 1970 (Lake Ziway) and Baumann et al., 1975 (Lake Shala)).


SILT + CLAY
(<63μm)
100%
SHALA
BOTTOM MUDS
21
22
27
23
4
11
9
2
5
12
8
6
1
13
Ajewa D6
14
7
15
10
3
100%
GRAVEL
(>2mm)
100%
SAND
(63μm-2mm)
Beach ridges
Abiyata mudflats
Ziway bottom sediments
Fluviodeltaic sediments


CLAY
(<2μm)
100%
Shala bottom sediments
Ziway bottom sediments
Abiyata mudflats
Fluviodeltaic sediments
21
22
27
23
ALL BEACH
SAMPLES
5,15
Ajewa D6
100%
SAND + GRAVEL
(>63μm)
100%
SILT
(2-63μm)

to consist of silty sand (Fig. 2.12c), although its mineralogy has not been studied. Small shore dunes of quartz sand are presently found at the southern tip of Abiyata (Gèze, 1974) (Plate 7) and east of Lake Chitu Haro, but otherwise aeolian deposits are absent.

4) Undercutting of rocky cliffs by waves is one possible source of gravelly beach deposits. During highstands, this source would have been supplemented by the erosion of slope deposits (McLachlan and McLachlan, 1971; McLachlan, 1974). Ignimbrite, rhyolite and obsidian shingle is abundant around Langano, E. and NE. Shala, and exposed shores of Lake Ziway. Around Abiyata, which is at present bordered by pumiceous sediments, resistant gravels are rare.

5) Primary air-fall pyroclastics in the Alutu piedmont range from vitric ash (silt and fine sand sizes) to pumice lapilli[1] (pebbles and granules). Unweathered samples are poorly sorted (σ_ϕ 1.47 – 1.98). Lithic fragments are scarce. Since large pumice lapilli may have specific gravities as low as 0.22 (Pettijohn et al., 1972, ch.7), they can float for long distances in rivers and lakes before sinking (Plate 17).

b. Marginal sedimentation (above wave base)

Extensive marginal accumulations of fine-grained organic sediments are restricted to Lake Ziway, along shores protected by dense aquatic vegetation (chapters 2D II and 2E III). The more exposed coastlines of Langano and Shala are fronted by steep, rocky cliffs or by low ridges of clean, coarse sand and shingle (Plate 18). Samples collected from the swash zone are generally monomodal and moderately well to poorly graded (σ_ϕ 0.58 – 1.38) (Fig. 2.12d). Wave action also sorts effectively by density as well as grain size, so that accumulations consisting largely of pumice or rhyolite are common. Ephemeral streams entering Shala are building out small, Gilbert-type deltas (Gilbert, 1885) into the lake. Where exposed along the eastern short by the pre-1933 drop in lake level (chapter 5B III), the deltaic

1. Terminology follows Fisher (1966).

deposits are seen to consist of moderately to moderately well sorted, ostracod-bearing, fine sands (σ_ϕ 0.69 - 0.90) with steeply inclined foreset bedding and ripple marks. Lenses of pumice pebbles, convolute bedding and gravel-filled channels are common.

Lake Abiyata is being slowly filled in from the north end by the Bulbula and Horocallo Rivers, which supply mainly pumice and fine-grained sediments (Plates 6, 12). The lake is bordered by gently sloping, salt-encrusted flats of very poorly sorted, clayey silt to silty sand (σ_ϕ 0.69-0.90). These are punctuated by low, parallel ridges of fine sand or well-rounded pumice gravel. Shallow pools near the shore have a pH of around 10. The silts contain clusters of disarticulated fish bones which have probably been regurgitated by the large flocks of pelicans and cormorants which feed on the lake (Plate 12).

The fine-grained character of the Abiyata muds reflects not only the sediment supply, but also the very gently shoaling offshore profile (1:35 to 1:620) and the rapid flocculation of suspended clays on entering the lake (Mohr, 1966). McLachlan and McLachlan (1976) have shown by experiment that the settling of fines in $NaHCO_3$-dominated waters is accelerated at conductivities above 500 μmhos/cm (K_{20}). Above 10,000 μmhos, precipitation is almost immediate, which accounts for the rapid deposition at the river mouths and the very low turbidity of Lake Shala (Plates 6,7).

c. Deep-water sedimentation (below wave base)

Even in lakes of this size, the depth range accessible to wave action is very limited (Smith and Sinclair, 1972). The bottom sediments become fine-grained within a relatively short distance from the water's edge.

No offshore samples from Langano or Abiyata have so far been studied. This is a pity, because lacustrine sands similar to the modern marginal facies of Lake Abiyata are common in the Bulbula formation. Lake Ziway is largely floored by silty sands which tend to become finer with depth below the sediment-water interface (Fig. 2.12).

The deep-water deposits of Lake Shala have been extensively studied by the Preussag Company, who took 32 short gravity cores from the lake bed in

1971 (Baumann <u>et al.</u>, 1975). The bottom muds, which consist mainly of volcanic glass, are extremely fine-grained (mean size 10.1 ϕ) and very poorly sorted (σ_ϕ 2.30 - 4.65) (A. Baumann, unpubl. data). Below 4 m, there is no variation in grain size with water depth. Poorly crystallized Mg-smectite, illite, chlorite, low-Mg calcite, quartz and feldspar are also present in small amounts. Dissolution of diatom frustules in the sediment is so advanced that only "ghosts" are left (Gasse, 1975, p.57). The accumulation and preservation of diatomite in the Ajewa formation clearly requires a significant reduction in lake-water alkalinity.

d. Authigenic minerals

Precipitation of carbonates and other evaporites is currently of minor importance. Lake Shala deposits a thin white crust on mud and rock surfaces (Plate 19) which was identified by X-ray diffraction as a mixture of <u>trona</u> ($NaHCO_3 . Na_2CO_3 . 2H_2O$) and <u>thermonatrite</u> ($Na_2CO_3 . H_2O$). Trona is a very common evaporite mineral in soda lakes with pH $\geqslant$ 9 (Eugster, 1971). Thermonatrite probably forms through dehydration of trona in strong sunlight (Lippmann, 1973).

<u>Calcium carbonate</u> has been identified in small amounts in deep-water sediments: 0-10% in Ziway, where it becomes more abundant with depth, and 0-9.6% in Shala. Baumann <u>et al</u>. (1975) found that low-Mg calcite occurred in Lake Shala in both clay and sand grades. They suggested two different origins: precipitation for the finer fraction and transport for the coarser (see chapter 2 E II). The presence of the low-Mg form is to be expected from the composition of the lake water (Mg/Ca < 1) (Muller <u>et al</u>., 1972). It is interesting that a plume of higher Ca and Mg contents occurs in the sediments of the northeastern corner of Lake Shala, offshore from the largest group of hot springs (Baumann <u>et al.</u>, 1975, fig. 4).

The shoreline muds of Lake Abiyata contain 3-10% low-Mg calcite, partly in the form of ostracod fragments. But it is likely that inorganic carbonate precipitation is also taking place as the Bulbula and Horocallo inflows mingle with the brackish lake water (Müller <u>et al.</u>, 1972).

Algal limestones are not forming in the present lakes, although fossil crusts are known from Shala and Dakadima. In the past, limestone formation appears to have been linked to hot-spring activity. The present springs deposit small quantities of siliceous sinter but no travertine (United Nations, 1973), which seems to indicate quite different geochemical and/or temperature conditions from those prevailing when the tufas and freshwater marls were deposited. The origin of the fossil carbonate deposits is discussed in chapter 7DIII.

CHAPTER THREE

Previous Investigations

3A SUMMARY

Research into the Quaternary history of the lakes in the Ziway-Shala Basin was initiated at the turn of the century by the Erlanger expedition, and subsequently continued by a number of Ethiopia- and foreign-based expeditions. The first ^{14}C dates became available in 1971. The picture which had emerged by 1975 can be summarized as follows:

1) Pliocene - Early Pleistocene[1]

Fine-grained tuffaceous sediments were deposited in lakes which formed in the developing Rift. The age, extent and topography of these water bodies is unknown, but some of the sediments predate the Gademotta (1.05 m.y.) and Shala (0.20-0.24 m.y.) calderas (Lloyd, 1977).

2) Middle Pleistocene

Beach gravels, tentatively dated 0.15-0.18 m.y., were deposited on Gademotta Ridge by a precursor of Lake Ziway. Although these deposits reach elevations $\geq$ 65 m above the present Meki-Awash watershed, there was a suggestion at least that the highstand was climatically controlled.

3) Late Pleistocene

The occurrence of lake-level maxima was demonstrated by the dating of freshwater mollusca from marginal sediments at 26,780 BP (1641 m) and 14,400 BP (1626 m).

The faulted West Ziway diatomites described by Taieb (1974) could be Middle or lower Late Pleistocene in age.

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1. These terms are defined in Appendix 2a.

4) Holocene

A major lacustrine highstand took place between 9500 and 5000 BP. All four principal lakes were united at about 1670 m, the elevation of the most prominent lake shoreline, and overflowed into the Awash. Rapid retreat after 5000 BP created the sequence of recessional shorelines fringing the three southern lakes.

3B RESEARCH PRIOR TO 1970

Early in 1970, when A.T. Grove, A.S. Goudie and G. Dekker began the first modern investigation into the fluctuations of the Galla Lakes, a quite detailed, though oversimplified, picture had already emerged from earlier work.

At the turn of the century, the Erlanger expedition established the existence of the four main lakes (Neumann, 1901, 1902). The zoologist in their party, Oscar Neumann, was the first to recognize the evidence for former highstands. He discovered a 25-30 m deep section of shelly lacustrine deposits along the Bulbula (Suksuk) River, which he attributed to "a great Tertiary or Diluvial lake". Neumann also identified a number of volcanic hills which had formerly been islands. The subfossil mollusca he collected were examined by Thiele (1933), and proved to be a freshwater fauna including a small number of so-called "palaearctic" species (Bacci, 1951). Members of the 1937-38 Italian mission (Brunelli et al., 1941) subsequently added to the list of palaearctic taxa (Bacci, 1940).

The uppermost Holocene shoreline was first noticed in 1904 by H.W. Blundell (1906): he described two distinct terraces at about 1660 m.a.s.l. around the north end of Lake Ziway. Negri (1913, cited in Taieb, 1974) measured much lower strandlines at 20 m, 16 m and 8-10 m above Lake Shala (1566 - 1578 m), and explained them by the progressive desiccation of an Early Quaternary lake.

The first coherent and detailed reconstruction of the Holocene palaeolake was published by Nilsson (1935, 1940). He established that all four modern lakes had once been united at a level controlled by an outlet on the

Ziway-Awash watershed to the west of Mt. Bora. His careful survey showed that their shorelines were horizontal. The most prominent was a terrace 'at about 40 m' above Lake Ziway (1676 – 81 m), which separated the gently sloping basin floor from the gullied slopes above. Nilsson correlated these fresh morphological features with his "Last Pluvial", but felt that older, consolidated sediments of "Kamasian" type had been laid down in an arm of the Red Sea. On the eastern shore of Lake Shala, Nilsson discovered drowned trees which recorded a rise in lake-level during historic times. He later climbed Mt. Badda (erroneously called Kaka) and mapped the glacial moraines which he discovered west of the summit (1940, fig. 38). Although he was unable to resist correlating the high lake levels with the glaciation, Nilsson's map of the "Last Pluvial" lake is astonishingly accurate considering the contemporary lack of roads, maps, and aerial photographs.

Further details of Holocene palaeogeography were added during geological reconnaissance work by P.A. Mohr (1960, 1966, 1971). He described the 'beautifully preserved' erosional bench at about 1660 m which runs from southeast Shala into the Abiyata Basin (1966, p.66), but he was uncertain about the relationship of such wave-cut forms and featureless deposits high above the lakes, which he felt must be Late Pleistocene in age, to the recessional beaches at lower elevations. He was the first to trace the well-marked strandline at about 1600 m, previously distinguished by Nilsson (1940, fig. 27), around all three southern lakes. By then it was clear that Mt. Alutu was the source of much of the pumice in the Holocene beach gravels. Mohr believed, however, that the fine-grained water-laid tuffs of the Langano area had been deposited in a much more extensive Plio-Pleistocene lake, before the development of the eastern escarpment (Mohr, 1966).

3C RECENT INVESTIGATIONS

Research on the Galla Lakes entered a new phase in 1971, when A.T. Grove and A.S. Goudie published radiocarbon dates of 9,220 ± 190 BP and 5,610 ± 100 BP on mollusca from the beaches above Lake Shala. From then onwards, it became possible to relate the Holocene lacustral phase in the

Ziway-Shala Basin to the climatic history of intertropical Africa in general (Grove and Goudie, 1971a; Street and Grove, 1976).

Although UNDP (United Nations, 1973) had already mapped the lake sediments around Shala and Alutu from air photographs, Grove and Goudie were the first to use photo-mosaics to reconstruct the outlines of the Holocene united lake (Grove et al., 1975). They called it "Pluvial Lake Galla". The former outlet on the Awash divide was shown to lie at ca. 1670 m (Grove and Dekker, 1976). Curiously enough, the abandoned spillway had already been mapped by Italian geologists working for Italconsult (Venzo, 1971b), but they had completely failed to grasp its significance. Dekker estimated that a 30% increase in rainfall over the catchment would be necessary to maintain the water surface at the overflow level, even allowing for a 10% drop in evaporation (Grove and Dekker, 1976). The resulting dilution of the Holocene lake waters was clearly apparent from the abundance of freshwater mollusca. Samples collected in 1970 from exposures northeast of Shala were identified by B.W. Sparks (Grove and Goudie, 1971a; Grove et al., 1975). D.S. Brown subsequently pointed out that the collections once again contained an important palaearctic element, and that one species in particular, Pisidium moitesserianum Paladilhe, was a first record from Africa (Brown, 1973b).

Grove and Goudie also found the first definite stratigraphic evidence for high lake levels during Late Pleistocene time. A shelly limestone at 1626 m east of Lake Shala yielded a date of 14,400 ± 750 BP (Grove et al., 1975). However, because of the paucity of height control, they were unable to demonstrate the pre-Holocene age of the terraces above the outlet level on the west side of Ziway (Grove and Goudie, 1971a, p.4). The general picture of an early to mid-Holocene Lake Galla remained a relatively simple one.

Meanwhile, a French geologist, Maurice Taieb (1974), had been studying outcrops in the northern part of the basin. Taieb described three sections of uplifted, pre-Holocene diatomites along the West Ziway faultscarp (Figs. 5.9, 5.15), and one in the Meki Valley. He attributed these sediments to a large Middle Pleistocene lake stretching from Ziway to Mojo, north of the

Awash. This correlation now appears very doubtful, particularly in view of the archaic diatom flora present at Mojo (Taieb, 1974, p.227). On the other hand, the view held by Frederic Gèze (1974), that the West Ziway lake beds are of Holocene age, seems equally untenable (chapter 5c).

Gèze produced detailed maps of the Holocene strandlines, but his main contribution was the description of a number of other important sedimentary exposures in the Bulbula Plain and around the southern lakes (Gèze, 1974, 1975). He obtained two ^{14}C dates, of 9360 $\pm$ 210 BP and 4960 $\pm$ 140 BP, which bracket the diatomites northwest of Alutu (his "Abelosa Section"). This confirmed the main conclusions of Grove and Goudie (1971a). However, more recent work (chapter 6) has shown that Gèze made a serious error in correlating these marginal deposits with the thick Upper Pleistocene diatomites outcropping in his "Bulbula Section", 3 km north of Bulbula Village. Hence the palaeoecological conclusions reached by F. Gasse, who studied the diatom floras from Bulbula (Gèze, 1975), relate to the Late Pleistocene and not to the Holocene. Gèze was nevertheless the first to point out the important role of the Hondola sill in the evolution of the North and South Basins of the Holocene lake (1974, pp. 180-181). On the subject of the overflow channel, he inclined towards the Italconsult view that it represented a former course of the Meki, now abandoned due to uplift of the drainage divide, and accordingly he played down its role as a spillway (<u>ibid</u>., pp. 182-185). He firmly rejected Taieb's idea of a Middle Pleistocene Galla-Awash lake, but accepted that the subaqueous volcanics of Alutu and Shala indicated the existence of lacustrine conditions well before the Holocene highstand (<u>ibid</u>., p.190).

A detailed study of the West Ziway-Gademotta area has recently been carried out by an archaeological expedition from Southern Methodist University (Wendorf and Schild, 1974; Wendorf <u>et al</u>., 1975; Laury and Albritton, 1975). Their findings provide further circumstantial evidence for a highstand of the Galla lakes during late Middle Pleistocene time. The SMU team excavated four Middle Stone Age (MSA) occupation levels within the Gademotta Formation, which veneers the back slopes of the Gademotta caldera rim, west of Lake Ziway (Fig. 5.9). All the MSA living floors occur in calcareous palaeosols within a sequence of clastic slope deposits and laharic mudstones. The stratigraphy is summarized in Fig. 3.1. The oldest MSA site immediately underlies a tuff K/Ar dated (on sanidine) at 0.181 ± 0.006 m.y. A channel fill, tentatively correlated with this oldest level, furnished hippo bone refuse, implying that permanent water (probably the ancestral Lake Ziway) existed within the hunting territory of the occupants.

Traces of an important phase of gully erosion, followed by aggradation, occur higher in the sequence, just below a second tuff dated at 0.149 ± 0.013 m.y. Laury and Albritton (1975) attribute the cut and fill cycle to increased surface runoff followed by a rise in lake level. Their argument gains some support from the presence of rhyolitic beach pebbles contouring the ridge of altitudes $\geqslant$ 100 m above Lake Ziway (Fig. 5.9, localities 1,2). Unfortunately the stratigraphic position of these gravels remains ill-defined, and several further problems of interpretation arise. Incision seems more likely to have resulted from a fall in lake-level, as it did during the terminal Pleistocene and late Holocene (Gasse and Street, 1978a). However, the inter-linkages between climate, vegetation and gully incision are very complex, and do not lend themselves readily to simple climatic interpretations (Cooke and Reeves, 1976). The elevation of the ancient beach, $\geqslant$ 65 m above the present drainage divide to the north, implies substantial warping of the Rift floor. Furthermore, the K/Ar dates themselves remain highly controversial, since they revise most previous ideas on the antiquity of the MSA tradition (Wendorf et al., 1975).

Laury and Albritton (1975) also published a very useful map of the lake benches west of Ziway (Fig. 5.9), together with a series of radiocarbon dates on shell (Haynes and Haas, 1974). The break between their terraces IV and V corresponds closely to the uppermost early and mid-Holocene shoreline as mapped on Fig. 5.1. Terraces V and VI are underlain by the older, more indurated lake deposits first described by Taieb (1974). The ^{14}C dates obtained range from 4800 to 9330 BP, and were obtained from terraces I to IV. An additional date of 26,780 ± 440 BP (SMU-61) was derived from lacustrine sediments at the northern end of the Bulbula Plain, confirming that a large freshwater lake had existed during Late Pleistocene time.

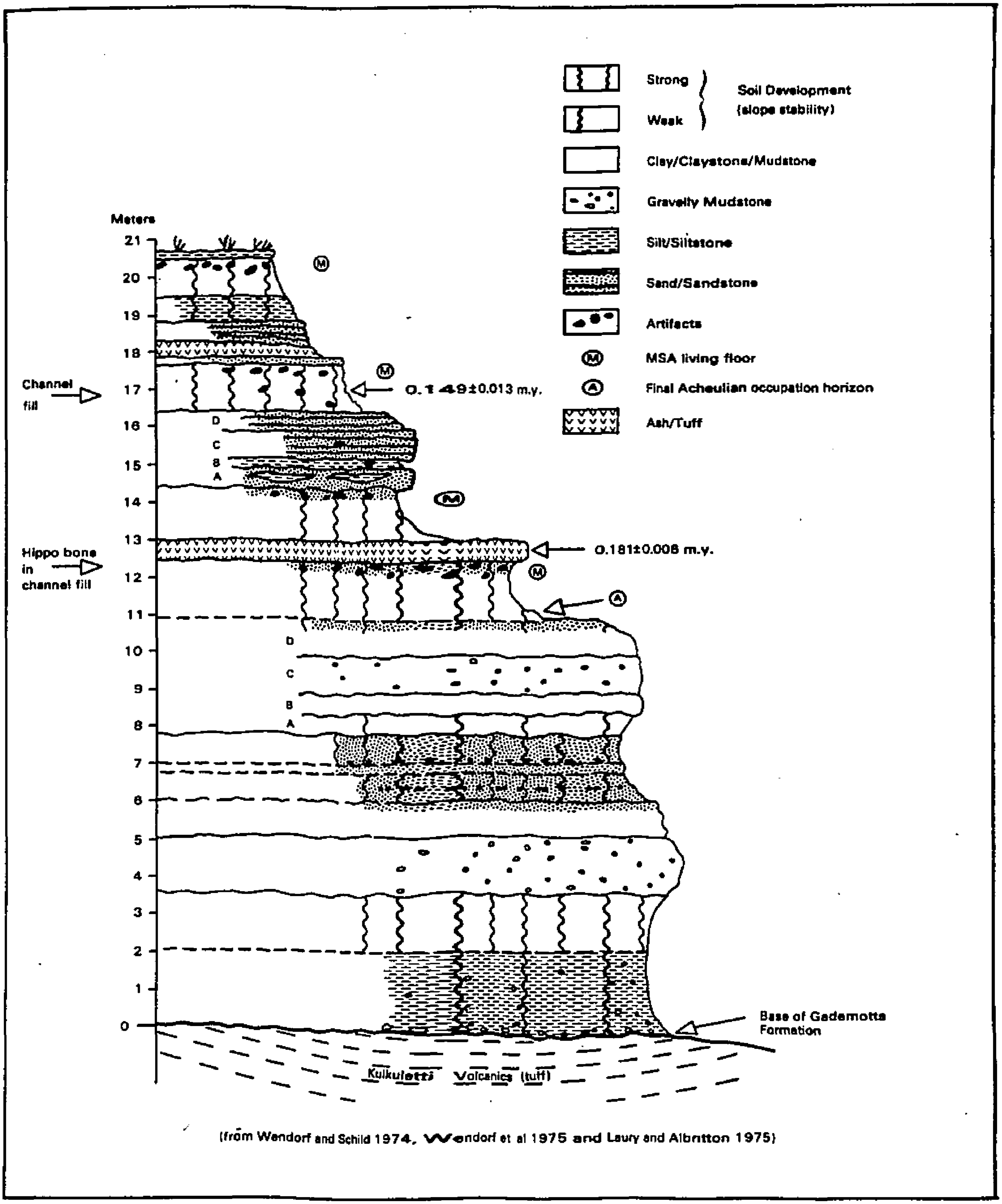


Fig. 3.1: Composite stratigraphy of the Gademotta Formation in the Gademotta and Kulkuletti areas (localities G and K on Fig. 5.9).

CHAPTER FOUR

Methods Used In This Study

The methodology used in this investigation was developed during the classic studies of lake-level fluctuations in the Great Basin by G.K. Gilbert (1890), R.B. Morrison (Morrison and Frye, 1965; Morrison, 1966) and G.I. Smith (1968). These workers integrated the information gained by detailed mapping of individual lake shorelines and their associated landforms with careful stratigraphic analysis of the lacustrine and intercalated subaerial deposits. According to Morrison (1966, pp. 79-80):

> The combination of areal mapping and measured stratigraphic sections provides a three-dimensional picture that is essential for comprehension of the intricate relationships between the many thin lacustrine units, their numerous changes in lithologic facies, and the interlacustrine unconformities separating the various lacustrine units, through the full altitude range between the lower and higher lake levels. Studies of inter-relationships of landforms alone ... have not yielded nearly so complete and unambiguous information on lake fluctuations as have the stratigraphic studies. Landform relations generally give information only on the younger lake cycles, and virtually none on the lake recessions.
>
> ... Lake maxima are determined by mapping the upper altitude limits of the deposits of each lake cycle and (in the case of the younger lake cycles) by observing the relations of these deposits to shore-geomorphic features. Lake recessions, much more difficult to determine fully, are represented in the successions by unconformities caused by subaerial erosion, by weathering profiles (soils), by wedges of alluvium, colluvium, eolian sand, or loess between the lacustrine sediments, and by shore deposits intercalated with deep-water deposits. Such criteria allow identification that a lake recession definitely went at least as low as a certain altitude.

This approach was adapted for use in Africa by Françoise Gasse in her study of the Afar lakes (Gasse, 1975, 1977a; Gasse and Street, 1978a). Her thesis confirmed the value of diatom analysis, which had earlier been applied to a core from Lake Naivasha by Richardson and Richardson (1972), as a tool for correlating lacustrine cores with surface exposures and for the interpretation of palaeoenvironments. It also indicated the fundamental importance of a good radiocarbon-dating framework, especially when dealing with Late Pleistocene sediments (Gasse and Delibrias, 1977).

Since the ultimate aim of the present study was to reconstruct the water balance of the lakes in the Ziway-Shala Basin at different times in the past, the extensive approach advocated by Morrison was felt to be more appropriate than the intensive sedimentological and biostratigraphic investigation of individual cores as carried out by Professor D. Livingstone and his students (Kendall, 1969; Richardson and Richardson, 1972; Hecky and Degens, 1973; Harvey, 1976; Holdship, 1976). This is because the first approach yields direct estimates of the parameter of greatest palaeoclimatic significance - lake area - and its variation through time. The procedure adopted in this study was therefore as follows:

1) <u>Geomorphology</u>

Individual lake shorelines were mapped from the 1971-72 D.O.S. aerial photographs at a scale of 1:40,000 to 1:56,000. Their altitudes were determined by levelling or by careful altimeter measurements (chapter 5). Although Celia Washbourn-Kamau, in her pioneering study of the Nakuru-Elmenteita Basin in Kenya, relied exclusively on levelling because of its greater precision, altimeter traverses were frequently used in this study in order to allow more time for stratigraphic work. The altitudes quoted are believed to be accurate within $\pm$ 5 m. This is quite adequate for water-balance modelling given the uncertainties attached to the estimates of hydrological variables such as evaporation and runoff.

2) <u>Stratigraphic description</u>

Detailed descriptions of sedimentary exposures in different parts of the basin were made in the field (chapter 6). Properties routinely recorded included dry Munsell colour, texture, sedimentary structures, carbonate content, pH and mollusc fauna. The presence of soils, erosional unconformities or artifacts was also noted. The principal exposures have been numbered according to the D.O.S. 1:50,000 grid square in which they lie (Fig. 4.1). For example, the Deka Wede gully, 0738-B20, is locality 20 in map area 0738B. The grid reference of each site is given in Appendix 1. Individual vertical profiles measured at intervals along a large gully system of this sort are numbered 0738-B20/1, B20/2 and so on. On the maps and in the text these are occasionally abbreviated to P1, P2 etc. when the context is clear. In all, a total of 173 profiles were measured, although only a small proportion are discussed in detail.

In addition to surface exposures, records were made of the stratigraphy of the Abiyata Core, five incomplete cores obtained by Italconsult from the Meki Valley, and the cuttings from the AIDBANK water-supply borehole in Adamitulu.

3) Dating and laboratory analyses

Samples of sediment and datable materials were brought back to the U.K. for analysis. The laboratories which have provided radiocarbon dates are listed in the Acknowledgments. All the shell samples submitted for dating were first analyzed by X-ray diffraction (XRD) to check for recrystallization (revealed by the presence of a calcite peak) which might cause their measured ages to be systematically too young (Appendix 2c). The majority of radiocarbon assays on shell were made on single species fractions. A complete ^{14}C date list for the basin is also given in Appendix 2b. The lake-level chronology used in this study is based entirely on this internal evidence and correlations with other areas are not introduced until chapter 10.

The sections analyzed in most detail were the type section of the Bulbula formation (0733-B20) (Gasse and Street, 1978a,b; Street, in press, a) and the Abiyata Core (chapter 6CIVa). Laboratory investigations carried out on 0738-B20 include grain-size analysis, measurement of carbonate and organic-carbon content, X-ray diffraction study of the carbonate and clay fractions, and petrographic examination of impregnated samples. The techniques used are listed in Table 4.1. The Abiyata Core, which was drilled by the Ethiopian Ministry of Mines, was sampled in Addis Ababa by Frédéric Gèze, Dr. Raymonde Bonnefille and myself with the aid of several assistants. Its diatom and phytolith stratigraphy is currently being studied by Mlle. Claudine Descourtieux. I have been responsible for the measurements of carbonate and organic-carbon content and for correlating the stratigraphy with surface exposures.

Analyses of other sections were carried out on a less systematic basis when this was important for correlation or for interpreting depositional environments. Samples of charcoal, mollusc shells, ostracods and crab remains were sent for identification to the various experts mentioned in the Acknowledgments. In addition to her work on the Mt. Badda Core, Dr. Françoise Gasse has very kindly examined a number of diatomite samples from the most important Rift-floor exposures. The corresponding species lists are given in Appendix 4b. Because she has not made formal counts on these samples, she prefers her ecological interpretations to be regarded as provisional.

The mollusc species present in bulk samples of the principal shell beds have been recorded on a presence or absence basis. At the beginning of this study, it was hoped to make detailed species counts on all these beds following the method developed by Sparks (1961, 1964). However, as the complex depositional history of the shell material was revealed by radiocarbon dating, this approach had to be abandoned because the number of samples collected was insufficiently large to allow for the bias resulting from reworking and sorting processes. Only very simple and conservative palaeoecological interpretations have therefore been attempted on the basis of presence/absence data for the basin as a whole.

4) Local correlation

The sequence of lacustrine and nonlacustrine sedimentary units in each area was established from the field and laboratory data. The stratigraphic procedures used are described in chapter 6. Past variations in water depth were then reconstructed from each local sequence using a variety of sedimentological evidence (chapter 7), and checked against the diatom assemblages when these were available.

5) Regional correlation

The sequences of lacustrine units in different parts of the basin were correlated with each other and with the shoreline evidence (Table 6.9). Criteria used for correlation include relative position and thickness, lithofacies, diatom assemblages and radiocarbon age. Soils have not been employed to such an extent as in the Great Basin (Morrison and Frye, 1965; Smith, 1968) because they are both less common and less distinctive.

6) Constructing a lake-level curve

The combined sequence of lake-level fluctuations was verified by comparison with the altitudinal distribution of ^{14}C dates on lacustrine and nonlacustrine deposits, and with the evidence on palaeosalinity and alkalinity derived from the diatom and mollusc data, to yield a lake-level curve for the basin as a whole (chapter 8).

7) Estimating past lake areas

The area of the lake(s) at different times was reconstructed from the shoreline evidence (for the younger lacustrine units) and from information on lake depth and bathymetry (for the older lacustrine units) (chapter 8).

8) Comparison with conditions in the catchment

The glacial moraines on Mt. Badda were mapped from aerial photographs and the snowline elevation at the last glacial maximum was approximated by Hofer's mid-height method (Osmaston, 1975). An estimate of the corresponding temperature depression was then made using the modern lapse rate (chapter 8). Samples from a 3 m-long peat core collected from a small bog in a glacial cirque at 4000 m were sent to Dr. Alan Hamilton and to Dr. Françoise Gasse for pollen and diatom analysis (Hamilton, 1977a,b; Gasse, 1978; Gasse and Descourtieux, in press). Their palaeoecological conclusions have been used as a guide to past conditions in the catchment; in particular to select more realistic values for the runoff coefficient than could be arrived at on the basis of modern hydrological evidence alone (chapter 9).

9) Water-balance calculations

The data on past lake areas, palaeotemperatures and vegetation changes were used to compute the precipitation required to sustain the lakes at different time periods (chapter 9). The water-balance model employed was a slightly modified version of the one developed by Washbourn (1967b). The results are compared with other eastern African lakes in chapter 10.

Note on terminology

The geomorphological and geological terms used in this study are used as defined in the Glossary of Geology (Gary et al., 1974), with the following emendations:

Subdivisions of the Quaternary (Gasse, 1975, annexe VB; Butzer and Isaac, 1975; Haq et al., 1977).

Stratigraphic nomenclature (Hedberg, 1976).

Salinity classification of surface waters (Gasse, 1975, annexe IIA—2, p.34).

Alkalinity classification of surface waters (Richardson et al., 1978).

Textural classes for unconsolidated sediment (Shepard, 1954).
Note: The terms gravel, sand, silt and clay are used as defined on the Wentworth scale except that the silt/clay boundary is taken as 2 μm (9ϕ) instead of 4 μm (8ϕ). The sand-silt-clay class is referred to as "loam" for the sake of brevity.

Field textural classes for soils (Soil Survey Staff, 1951).

Sedimentary environments and structures (Born, 1972; Picard and High, 1973; Reineck and Singh, 1973; Collinson et al., 1978).

Volcaniclastic rocks (Fisher, 1966).

Details of the salinity and alkalinity classifications, the terminology for different lacustrine environments, and the time scale employed are given in Appendices 4, 9 and 1 respectively.

Table 4.1: Field and laboratory methods used in this investigation

Item	Method	Equipment	Reference
Altitude:-			
(in m.a.s.l. (metres above sea level) or m.S.D. (metres above Shala datum, 1558 m))	Levelling traverses from U.S. Corps of Engineers bench marks on Addis Ababa-Shashamanne road; D.O.S. trigonometric stations; or lake surfaces	Wild N-2 level and metric staff	
	or		
	Altimeter traverses from datum points listed above. Readings corrected for temperature and diurnal pressure variations.	Wallace and Tiernan altimeter reading up to 15,000 ft, with 20 ft scale divisions	
Field description:-			
Thickness of beds (in m)	Measured to nearest 5 cm on exposure cleaned with pick and/or trowel	30 m tape	
Dip and strike (in degrees)		Brunton compass	
Textural class	Subjective estimate. Best results for glassy, low-clay sediments obtained with dry samples.		Shepard, 1954
Carbonate content	Reaction with dilute HCl		Soil Survey Staff 1951
pH	By comparison with standard chart, using universal soil indicator and $BaSO_4$	BDH soil test kit	
Munsell colour	(Dry sediment)	Revised Japanese Standard Soil Colour Charts	
Soil description			Soil Survey Staff 1951; F.A.O. 1974

Table 4.1 continued:

Item	Method	Equipment	Reference
Laboratory analyses:-			
Texture (unconsolidated sediments) (in Ø units)	Dry sieving (sand and gravel fractions)	Set of graduated sieves at 1 Ø intervals from -5Ø to +4 Ø	Folk, 1968
	and/or		
	Wet sieving followed by pipetting (silt and clay fractions). Samples were dispersed using 2.5 g/l of Na $(PO_4)_6$ and adjusting pH to 8 with ammonia solution		Folk, 1968; S. Bieda, C.N.R.S pers. comm.
	(Carbonate and organic matter were removed when necessary by pretreatment with 10% HCl, followed by boiling gently for 1 hr on a hot plate with 7.5% H_2O_2)		(Day, 1965)
	(Textural parameters (M_ϕ, σ_ϕ etc.) were calculated by the method of moments and grain-size histograms plotted using a package programme)	Hewlett-Packard desk computer and plotter	
Texture (soils) (in % sand, silt, clay)	Hydrometer analysis, following carbonate removal and dispersal as above.	U.S. standard hydrometer (A.S.T.M. 152H)	Day, 1965
Organic carbon content (in %)	Wet oxidation with potassium dichromate (Tinsley's method as modified by Dr. R. Perrin, Dept. of Applied Biology, Cambridge)		Bremner and Jenkinson, 1960
Carbonate content (in %)	Simple manometric measurement of CO_2 evolved on adding excess 10% HCl. Calibrated with pure $CaCO_3$	Bernard calcimeter	L. Casta, C.N.R.S., pers. comm.

Table 4.1 continued:

Item	Method	Equipment	Reference
Carbonate mineralogy (sediments)	XRD. Finely ground samples scanned from 25-32° 2θ across aragonite (3.40 Å), calcite (3.04 Å) and dolomite (2.89 Å) peaks (Settings: scan speed $\frac{1}{2}^{\circ}$ 2θ /min, chart speed 10 mm/min, time constant 1, 300 or 1000 c.p.s.) Weight % $MgCO_3$ in calcite lattice determined from spacing of (100) peak using CaF_2 as internal standard	Phillips diffractometer (Cambridge: 35 kv, 20 mA, with mono-chromator) (Oxford: 36 kv, 30 mA, no mono-chromator). CuKα radiation, Ni filter, 1° slits, randomly oriented powder mounts on glass slides	Griffin, 1971
Carbonate mineralogy (shells)	XRD. Shells ground to a smooth paste with acetone in an agate mortar before mounting. Grinding minimized to avoid conversion of calcite to aragonite (Burns and Bredig, 1956). Samples scanned three times from 25-30.5° 2θ. (Settings: $\frac{1}{4}^{\circ}$ 2θ /min, chart speed 10 mm/min, time constant 1, 1000 c.p.s.) Calcite content obtained from the peak-height ratio (measured above background) of aragonite (3.40 Å) to calcite (3.04 Å). A calibration curve was constructed from known mixtures of ground Iceland spar and non-biogenic aragonite.	" "	Lowenstam, 1954, Griffin, 1971
Other evaporites	XRD. Finely ground samples scanned from 5-50° 2θ. (Settings: $\frac{1}{2}^{\circ}$ 2θ /min, chart speed 10 mm/min, time constant 4, 1000 c.p.s.) Minerals identified from A.S.T.M. powder diffraction file.	" "	

Table 4.1 continued:

Item	Method	Equipment	Reference
Clay mineralogy (sediments)	XRD. Randomly oriented mounts made from the < 2 µm fraction obtained by sampling the fines remaining in suspension after 7 h 24 min ($24^{\circ}C$), placing a few drops of the suspension on a glass slide and allowing to dry overnight. Pretreatments described under pipette analysis. Scanned from $2^{\circ}-50^{\circ}$ 2θ without further treatment; after glycolation; and after heating to $550^{\circ}C$ for 1 hr. (Settings: 2° 2θ /min, chart speed 10 mm/min)	Phillips diffractometer, C.N.R.S. 46 kv, 26 mA, with monochromator; CoKα radiation	Carroll, 1970
Impregnation of unconsolidated sediments with polystyrene for thin-section work	Small samples placed in wine-bottle caps (lead) or small polythene bags and allowed to soak in polystyrene mix at room temperature. Proportions of monostyrene, catalyst and accelerator adjusted to allow polystyrene to set slowly after about 3 weeks, gradually expelling all air bubbles. When hard, stands up very well to sawing and mounting.		L. Casta, C.N.R.S., pers. comm.
Photomicrographs	Thin sections photographed at 25x to 400x magnification, using FP4 film with green filter.	Zeiss Ultraphot photomicroscope	
Bulk density of sediment (in g/cm^3)	Samples coated with either hot paraffin wax (porous diatomite) or a solution of saran F310 resin (obtained from K. & K. Greeff Chemical Group Ltd., Croydon) in methyl ethyl ketone (less porous samples). Density measured by weighing in air and in water, allowing for weight of coating.		Brasher et al., 1966

CHAPTER FIVE

Geomorphic and Pedologic Evidence for Former Lake Levels

5A SUMMARY

This chapter presents a preliminary model of past lake-level fluctuations. This is based on the sequence of strandlines and surface soils which surround the present lakes. With the exception of Lake Ziway, bathymetric surveys have so far failed to reveal any submerged landforms dating from lower water levels than present.

Shoreline remnants of Pleistocene age are equivocal and seldom well preserved. Since radiometric dates covering this interval are still sparse, it was necessary to fall back on artifacts or soil development as an indication of relative age. The geometry of the various Holocene lakes has been reconstructed from their well-preserved erosional and depositional shoreline features which have been dated in several localities by ^{14}C. For the period since 1967, detailed water-level records are available for the three northern lakes. These have been supplemented by isolated historical observations back to 1933.

The reconstructed sequence of lake-level highstands is as follows:

1) A very extensive lake or lakes may have existed during the Middle Pleistocene (Chapter 3C), but its geometry can no longer be reconstructed due to warping of the Rift floor.

2) An extensive, brackish lake formed in the area southwest of Meki, prior to the development of the West Ziway fault. Its sediments, which appear on geomorphological grounds to be younger than the Gademotta Formation (0.15 m.y.) outcrop up to at least 1696 m between Ziway and Meki. Undeformed terrace remnants at 1680 m west of Ziway Town are probably younger in age and may relate to a pre-Holocene outlet at 1675-1680 m.

3) A prominent Holocene shoreline at 1669 $\pm$ 1 m can be followed around all the present lakes. This level was reached more than once.

Overflow through the Meki-Dubeta col originally took place at ca. 1673 m, but lake-level eventually stabilized between 1667 and 1671 m. This highstand or highstands left well-preserved erosional landforms and some rather denuded beach ridges and deltas. Its recessional shorelines retain little original morphology except in the area northwest of Abiyata.

4) A later Holocene highstand brought lake level to 1595-1600 m, uniting the three southern lakes. The level of Lake Ziway was probably little affected. This oscillation is chiefly represented by fresh depositional features.

5) Secular fluctuations of the lakes have occurred in response to variations in rainfall totals over the Ethiopian Highlands. Maxima of the terminal lakes Abiyata and/or Shala occurred around 1938 and 1971 and minima around 1933, 1956, 1967 and post-1974. The peaks and troughs correspond closely to the Addis Ababa rainfall curve (Wood and Lovett, 1974) if a lag of 1-4 years is allowed for.

In the following sections, the evidence is set out by age and by area. The discussion begins with the less controversial Holocene features.

5B HOLOCENE HIGHSTANDS

Within the limits of the Holocene palaeolakes ancient shorelines are strikingly visible on aerial photographs (Plates 6,7,34). They are marked by wave-cut benches, cliffs and stacks as well as by depositional forms - deltas, shingle ridges, spits and tombolos (Table 5.4). In general only the erosional features are obvious on the ground. Figure 5.1 shows the two principal shoreline levels, at 1670 m and 1595-1600 m, which have been mapped from aerial photographs at a scale of 1:40,000 to 1:56,000 (chapter 4). These shorelines are described below.

I. The 1670 m united lake

The uppermost shoreline level of Holocene age can be followed right round the present lakes with little difficulty. It links up with the Meki-Dubeta overflow channel in area 0838D sheet 4 (Fig. 4.1). Measurements of

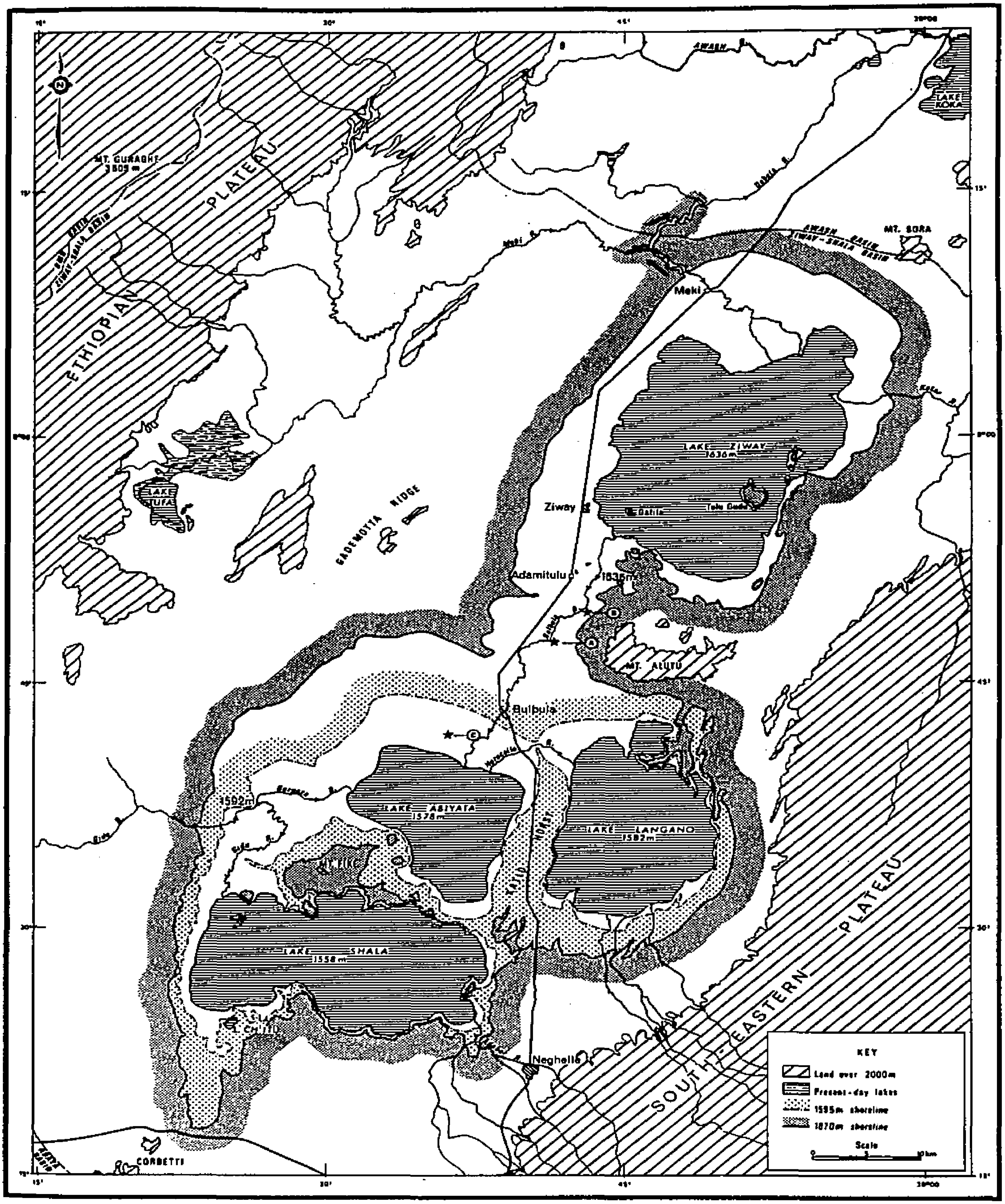


Fig. 5.1: Extent of Lake Ziway-Shala during the early and mid-Holocene (1670 m shoreline) and the late Holocene (1595-1600 m shoreline).

its altitude are very consistent, and do not reveal any significant tectonic or isostatic warping (section IVe) although local fault displacements occur along the West Ziway fault and northeast of Shala (Figs. 5.2 and 6.13).

The former lake margins are described in detail below, beginning with the overflow channel and proceeding clockwise around the palaeolake:

a. <u>The Meki-Dubeta overflow channel (area 0838D sheet 4)</u>

The overflow channel crosses the Awash divide at its lowest point (Fig. 5.3). It runs for 6.75 km as the crow flies from the north side of the Meki valley, 20 km upstream from Meki Town, to the Dubeta river, an intermittent tributary of the Awash. The channel has a general NE trend, though meandering widely in its northern half. Italconsult (1970) produced a plan of the area with 2 m contour intervals, and a geological report (Venzo, 1971b). From these it is clear that the spillway, which is incised 25–40 m below the general level of the saddle, is cut through largely unconsolidated air-fall pumice beds of the Wonji Series. These are well exposed in a quarry north of Meki (Plate 20). It is likely that a substratum of trachytic lava or ignimbrite exists at depth (Venzo, 1971b, p.248).

The abandoned channel floor is flat, and 100–120 m wide in its central part. It keeps to 1667 m for most of its length, descending towards the Meki River (1654 m) at its southern end, and rising to the Dubeta Valley (1675 m) in the north. This overall slope has been attributed to later downcutting by the Meki River and to blockage of the northern end by a plug of sediment (Grove and Dekker, 1976).

Fieldwork in the area was directed towards answering several important questions:

1) What is the altitude of the "bedrock" threshold of the Ziway-Shala Basin?
2) At what level did overflow actually take place?
3) Are there signs of earlier outflows through this col?
4) Is the NE trend of the channel in any way related to the Wonji Fault Belt?
5) How is the outlet connected to the shorelines at the northern end of Ziway?

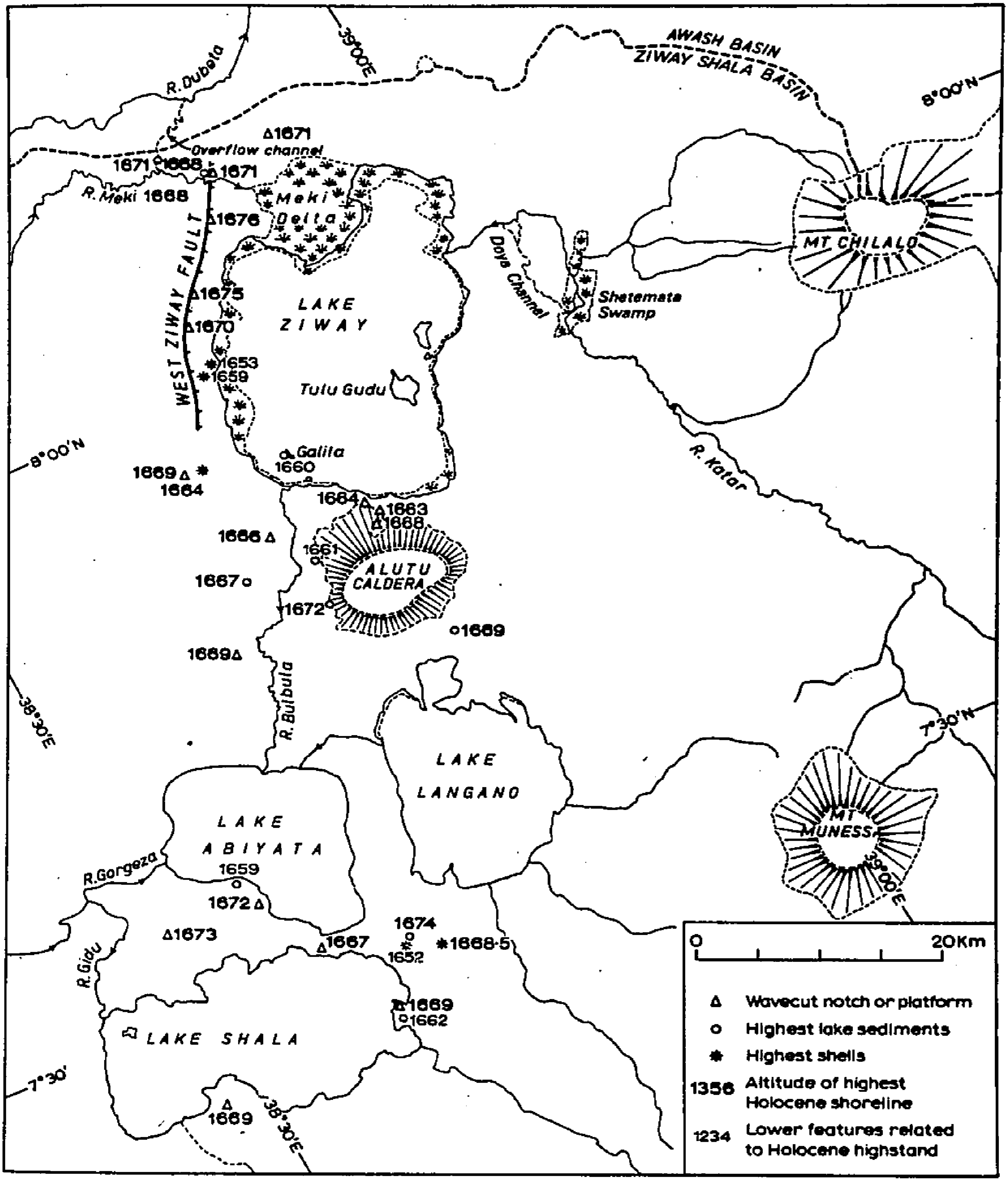


Fig. 5.2: Altitudes measured along the highest Holocene shoreline.

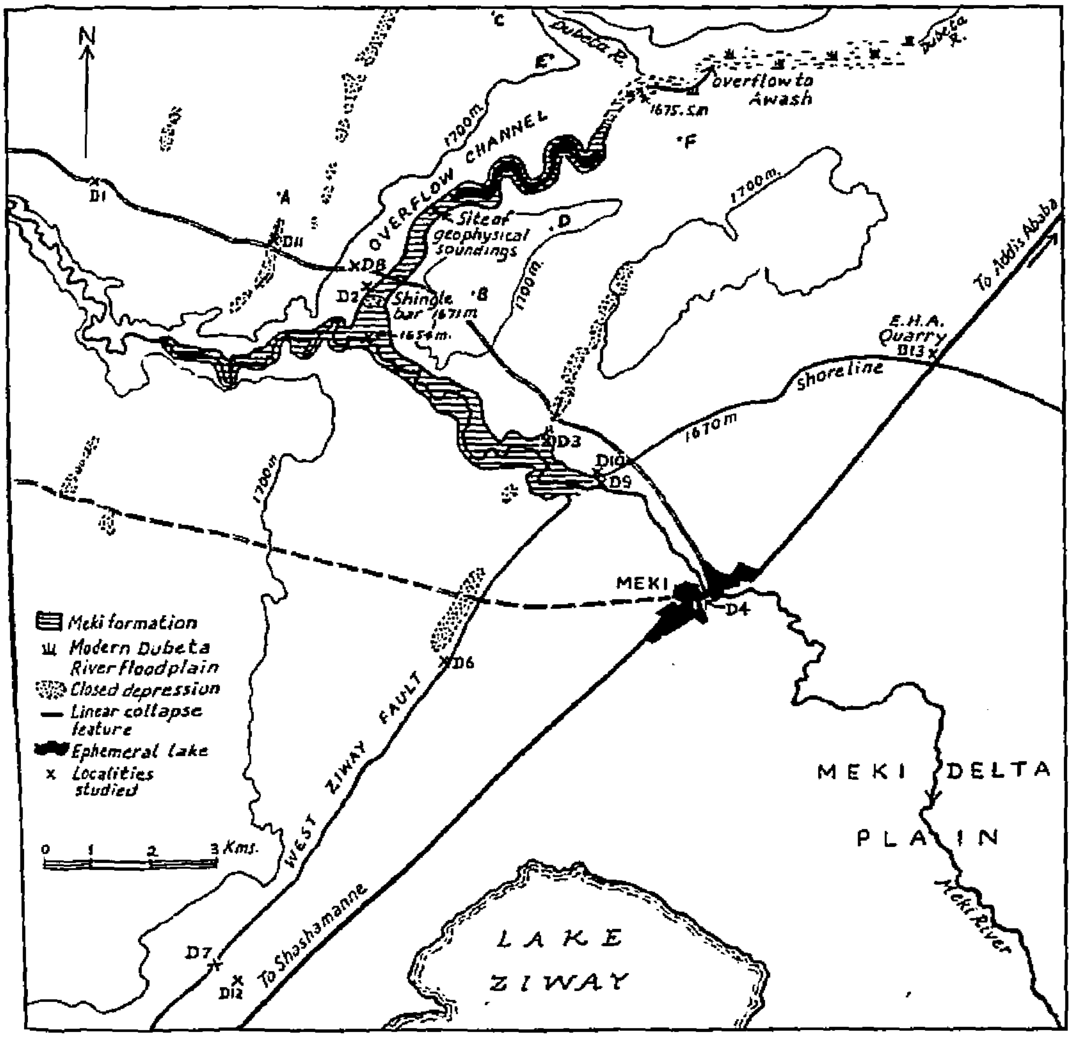


Fig. 5.3: Map of the Meki-Dubeta overflow channel showing the distribution of the Meki formation.

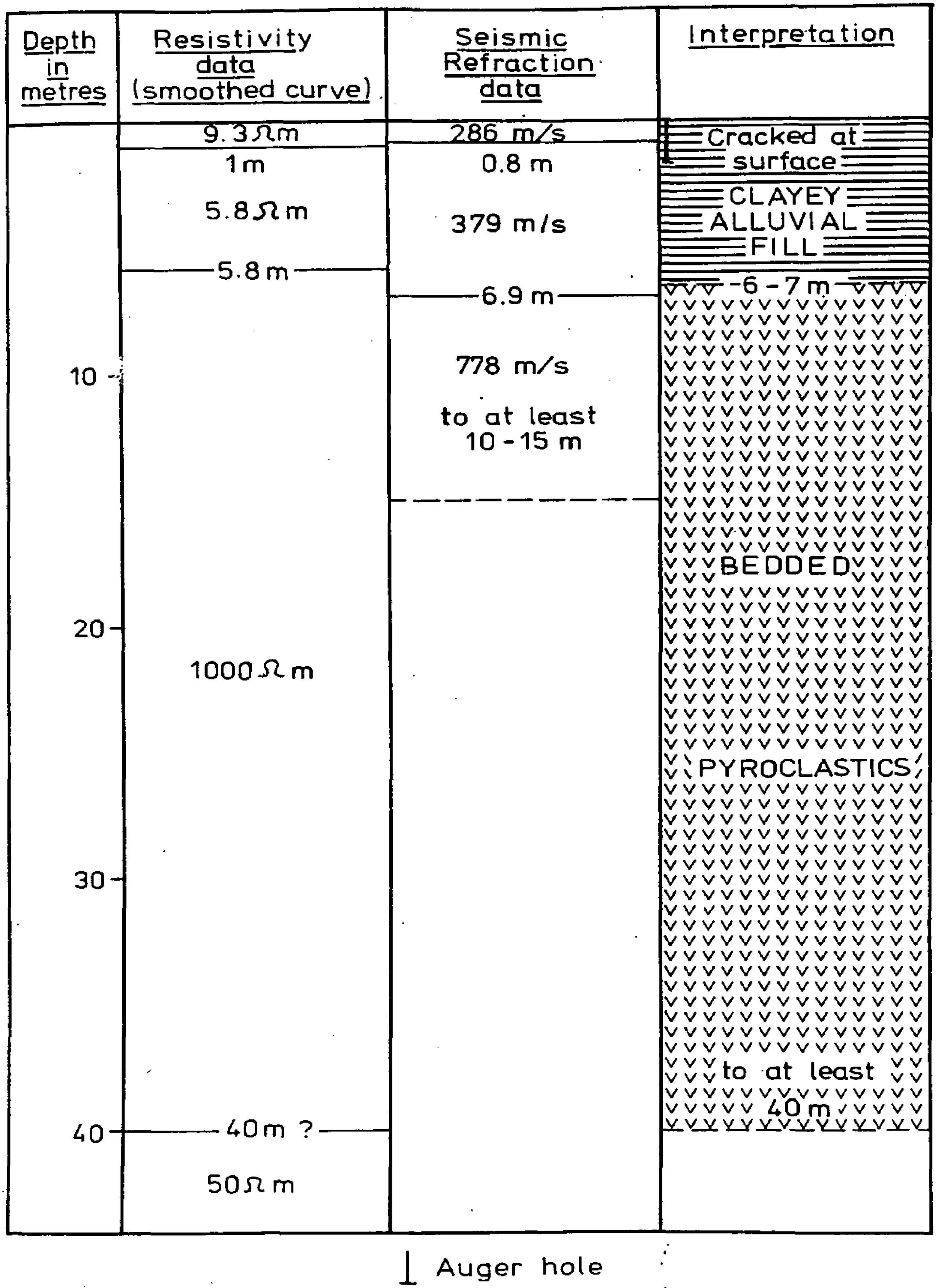


Table 5.1: Interpretation of geophysical results, Meki-Dubeta overflow channel

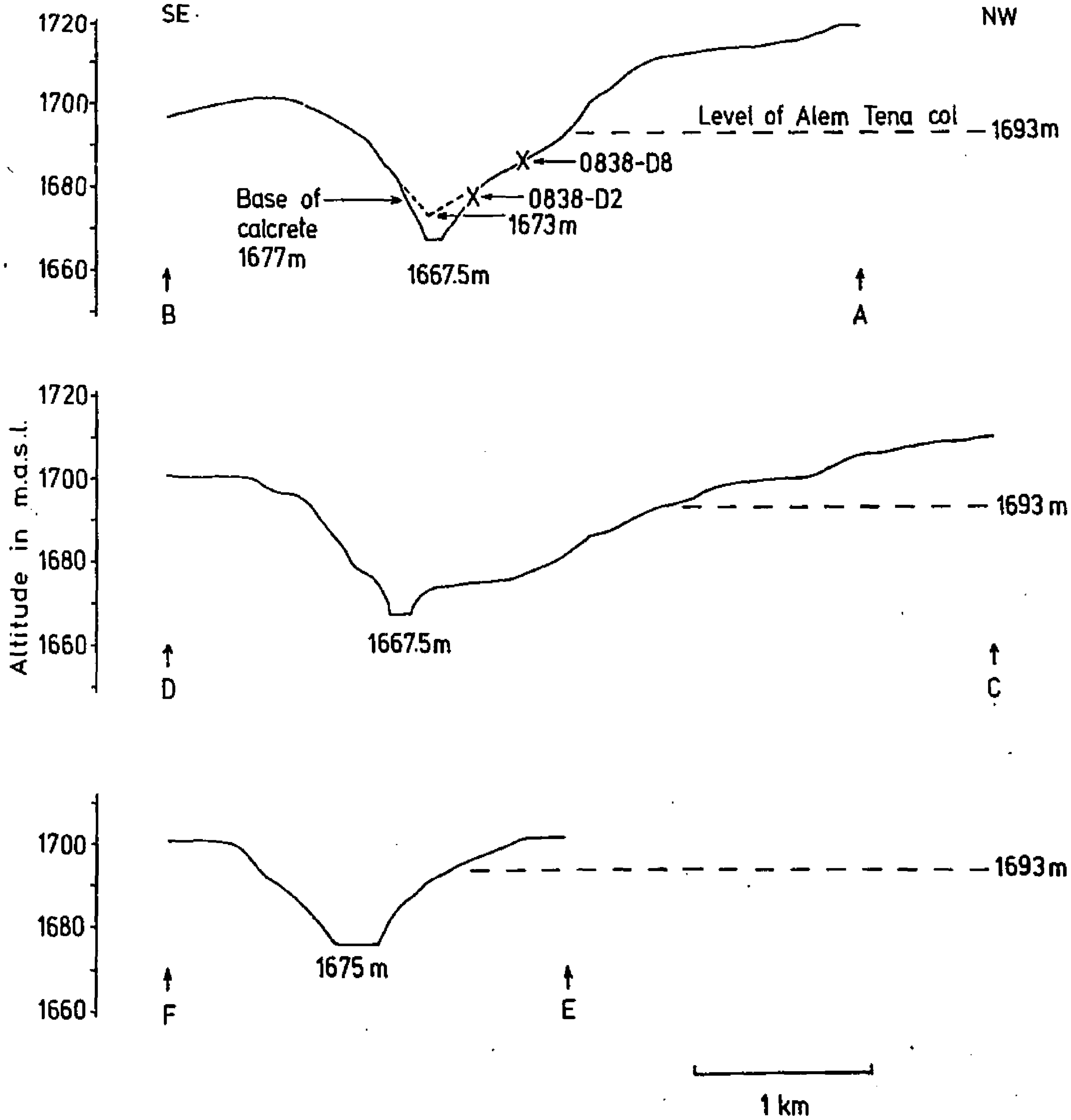
1) Cross profiles of the overflow channel (Fig. 5.4) show that the valley sides slope gently down to a few metres above its floor. The central section is a swampy "dambo" underlain by cracking clays. During the rains it becomes a shallow, winding lake more than 5 km long (Plate 21). Dr. David Emilia and Dr. Brian Last, of the Geophysical Observatory and Department of Physics, Addis Ababa University, kindly made resistivity and seismic refraction soundings to determine the depth to "bedrock" in the central part of the depression. They used a Soiltest ER2 Seismograph (employing the seismic hammer technique) and a Scintrex RAC-8 Resistivity Meter (Plate 22). Both lines were laid out along the axis of the dried lake floor. The resistivity measurements were interpreted using the Gosh method (Gosh, 1971).

The results, shown in Table 5.1, are somewhat ambiguous, but can be evaluated in the light of previously published information (Table 5.2). The seismic velocities down to 6.9 m are very similar to those attributed to "weathered" and "dry" silty-sandy alluvium. The presence of a sedimentary infill was confirmed by an auger hole in the centre of the survey line which penetrated 1.70 m of dark brown, vertisolic clay. This was very moist below 0.6 m. From 6-7 m downwards there is a puzzling, relatively low-velocity,

Table 5.2: Seismic velocities measured in the Ziway-Meki-Koka area (from Italconsult, 1970)

Interpreted rock type	Mean velocity $\pm 1\sigma$ (m/s)	Number of sites
Surface soil ('weathered surface')	222.5 ± 80.8	15
Silty-sandy alluvium (dry)	346.5 ± 53.2	8
Water in alluvium	1045.4 ± 455.3	5
Pumice-tuffs (dry)	677.4 ± 145.1	9
Pumice-tuffs (wet or more indurated)	863.3 ± 192.7	8
Hard rock	2438	1

Fig. 5.4: Cross sections of the Meki-Dubeta overflow channel (data from Italconsult, 1970).



SE
NW
1720
1700
1680
1660
Level of Alem Tena col
1693m
0838-D8
0838-D2
Base of calcrete 1677m
1673m
1667.5m
B
A
Altitude in m.a.s.l.
1720
1700
1680
1660
1693 m
1667.5m
D
C
1700
1680
1660
1693m
1675 m
F
E
1 km

but high-resistivity (1000 Ω m) layer, extending to at least 40 m. A velocity of 778 m/s is far too low for solid rock, although the high resistivity indicates a material which is at most damp, and far above the water table. The seismic results fall in the quoted ranges for "dry" and "wet/more compact" pumice-tuffs (Table 5.2). The 7-40 m layer is therefore interpreted as dry, unindurated or partly indurated pyroclastics in situ, sealed by an overlying clayey infill.

We can assume from these results that the erosional threshold of the basin lies at 1660 m and is cut into relatively unindurated pumice beds.

2) Immediately to the south of the road the depression is closed off by a shingle ridge consisting of uncemented, subrounded to well-rounded, lava and obsidian pebbles. Its elevation is 1671 m. Presumably the channel was plugged by a gravel bar from the Meki River at the time when overflow ceased (cf. Beadle, 1974, p. 231). Minimum and maximum altitudes for the lake level at overflow are set by the "bedrock" threshold (1660 m) and by the next highest col at Alem Tena (1693 m). A conservative estimate of 1670 m, 3 m above the present floor of the depression and 1 m below the top of the gravel bar, fits in well with evidence from the rest of the basin (Fig. 5.2).

3) Cross profiles of the Holocene channel suggest that it may have been incised into the floor of a pre-existing valley (Fig. 5.4). Exposures near the road confirm this interpretation. The valley sides are armoured by calcrete down to elevations of 1677-78 m. Upslope, the duricrust can be traced beneath outliers of the original soil cover (Plate 5). At site 0838-D2, in a small gully, the channel can be seen to cut through the calcrete, which has developed in the top of a 1.75 m thick section of bedded gravels (Plate 23). These are subrounded to well-rounded. They range from pumice cobbles at the base to lava and obsidian pebbles at the top, and were previously interpreted as a beach deposit (Gasse and Street, 1978a). But it now appears more likely that they originated in the same way as the shingle bar mentioned above.

If the calcrete-covered slopes are projected towards the centre of the channel, an estimate of 1673 m is obtained for the pre-Holocene valley floor.

This would imply about 13 m of downcutting and 7-11 m of backfilling during the Holocene. A corollary is that any Late Pleistocene overflow took place at a somewhat higher level (chapter 5C).

4) Fieldwork in 1975 also revealed a connection between the alignment of the outlet and extensional tectonics along the Wonji Fault Belt. The northern and southern sections of the channel run parallel to a field of linear collapse depressions and vertical fissures up to 10 m deep which trends in the same direction as the West Ziway fault (N20°-30°E) (Fig. 5.3). These have developed in gravelly slope deposits or in the underlying bedded pumice and ash.

Swarms of parallel vertical fissures have commonly been reported along the Fault Belt (Gibson, 1967; Mohr, 1967; Catlin et al., 1973; Clark and Williams, in press). In ignimbrites and lavas, the cause is usually said to be tensional separation, whereas in unconsolidated pumice or alluvium they have been attributed by Albritton (1974) to subsoil piping and consequent collapse. These explanations are not mutually exclusive. Collapse features in unconsolidated materials have been observed to form in 1956 and 1966 near Welenchiti (Gouin and Mohr, 1967) and in 1960 and 1969 near the Koka Dam (Italconsult, 1970, vol. 3, pp. 39-41). Their appearance coincided with heavy rains (1956, 1966, 1969) or, in one case, with the impounding of the artificial lake (1960). In both areas the fissures are clearly related to regional fault patterns. At Welenchiti, they occur in a graben filled with up to 100 m of sandy alluvium. The development of surface cracks and sink holes was preceded by seismic activity which was apparently caused by fracturing of the solid bedrock at depth. Collapse along the resulting lines of underground weakness was initiated by the first heavy downpours. Near Koka, large sink holes and fissures have developed along the downthrown side of fault scarps, where lacustrine or alluvial sediments, overlying pyroclastics, abut onto jointed lavas. The fissures now form an integrated subsurface drainage network well above the water table. The Italconsult geologists (1970) attributed them to subsurface erosion by water at the contact between the pyroclastics and the less permeable lavas.

The open fissures and linear depressions on the Meki-Awash watershed seem closest to the Welenchiti examples, which suggest that they originated by collapse of the stratified pumice deposits into fractures in a hard substratum of lava or ignimbrite lying at > 40 m depth. Very large volumes of sediment can be swallowed in this way (Williams and Dakin, 1976). The NE/SW valley exploited by the Holocene and presumed Late Pleistocene overflows from the Ziway-Shala Basin was probably a linear depression of this type.

5) It seems remarkable that the uppermost Holocene shoreline crosses the Meki without any change of direction (Figs. 5.1, 5.3). It runs along the foot of the West Ziway fault scarp to the river, rejoins the 1680 m map contour on the far side, and then curves round in a broad sweep to parallel the Ziway-Awash watershed. No delta morphology developed, although lake deposits are present in the vicinity of Meki Town (sites 0834-D4 and D9).

Investigations by Italconsult in the overflow area (1970, vol. 3) show that the Meki is bordered by cut-and-fill terraces of fine-grained alluvium, which can be followed into the entrance of the spillway at 1670 m. One plausible hypothesis is that rising lake level "backed up" the Meki drainage, causing the river to overflow through a pre-existing col and to dump its sediment load before reaching the lake. Aerial photographs show clearly that the Dubeta River runs in a narrow, incised valley upstream from the junction, whereas downstream it meanders in a broad, flat-floored, alluvial valley which is underlain, according to Venzo (1971b) by several metres of pebbly, pumiceous silts and sands closely resembling the Meki terrace deposits.

The nature of these alluvial fills and of the lacustrine sediments near Meki Town will be discussed further in chapter 6E. On its own, the geomorphological evidence suggests the interesting possibility that the bulk of the Meki inflow was diverted immediately through the outlet, so that the lake waters at maximum were essentially replenished by the Katar, Langano and Gidu drainages. This would be similar to the situation in Lake Mobutu Sese Seko (Albert), into which the Victoria Nile enters immediately adjacent to the outlet, thus contributing little to the water chemistry, although it helps to maintain lake level 'in the manner of the inflow to a constant-level water still' (Beadle, 1974, p.139).

b. <u>North Ziway (area 0838D sheet 4)</u>

The limit of the Holocene lake can easily be followed on air photos from the Meki to Mt. Bora, but it becomes less and less distinct on the ground. A break of slope at 1671 m is still clearly evident where the shoreline crosses the Mojo-Shashamanne road, just below the large road-metal quarry (0838-D13). This exposes the undisturbed Wonji Series volcanics into which the cliffline was cut (Plate 20).

On the gently-sloping flanks of Mt. Bora, the edge of the palaeolake shows up clearly as a lower limit of ephemeral gullies (Grove <u>et al</u>., 1975) (Plate 24), but is obscured on the ground by slope-wash. It runs between the 1660 m and 1680 m contours to a point due north of the northeast embayment of Lake Ziway, where there is an indentation. Here it crosses a N/S tectonic lineament, but there is no sign of fault displacement. The recessional strand-lines in this sector are low ridges which slow up as vegetation stripes, in contrast to the terraces on the west side of the lake (Fig. 5.9).

c. <u>East Ziway and the islands (areas 0838D sheet 4 and 0738B sheet 2)</u>

Along the step-faulted eastern margin of Ziway, the Holocene limit is very problematical. In general, the uppermost strandline seldom departs far from the eastern shore of the lake although it is uncertain how far it penetrated up the Katar Valley. The Shetemata Swamp, at 1680-1700 m, appears to have formed a stilling basin on the Katar just above its entrance into Ziway. During Late Pleistocene lake-level maxima it may even have been an adjunct of the main lake (Fig. 5.2). This may explain why the Katar River has abandoned its original outlet at the southern end of the swamp in favour of one near Ogelcho. The meandering Doya channel is very fresh-looking and deeply incised (Di Paola, 1976). However, disruption of the drainage by recent faulting cannot be ruled out, since the Doya stream now flows back into the Shetemata graben.

At least five of the larger islands in Lake Ziway emerged above the waters of the palaeolake, as well as a number of hyaloclastite and scoria cones on the south shore. The largest island, Tulu Gudu, is a faulted volcanic edifice rising to over 1940 m. Extensive agricultural terracing on its slopes has destroyed any trace of former shorelines. Galila (1694 m) is a smaller, basaltic tuff cone, breached towards the northeast. Thin algal limestones were found coating boulders on its slopes up to 1660 m.

d. <u>The Alutu piedmont (area 0738B sheets 1-4)</u>

The shoreline around the Alutu volcanic complex consisted of alternating, shallow embayments and stretches of low cliffs where outlying eruptive centres were trimmed by the lake. The outer slopes of the massif are formed of rhyolite domes, hyaloclastite rings, and viscous, obsidian flows.

On the south shore of Ziway, a cliffline and wave-cut platform, incised into promontories of soft, subaqueous tuffs, show up quite clearly. A former stack is present on the rock platform at grid ref. DU 761694. In the embayments there are distinct terraces which run in smooth curves from one headland to another, parallel to the modern shore. They are largely covered by spreads of slope-wash and modern pumiceous alluvium, so that exposures of lacustrine sediments are rare; but at site 0738-B29 the uppermost terrace, which corresponds in elevation to terrace IV, West Ziway (Fig. 5.9), is underlain by thin-bedded, ripple-marked, very fine pumice sand, with some large floated pumice clasts near the base. Although no diatomites outcrop here, probably because the waters were too turbid with fine sediment, a bed of fairly pure diatomite with fishbones is locally exposed at 1658 m, in the more protected Abelosa embayment on the northwest side of the massif (site 0738-B28).

To the west of the Bulbula River at Adamitulu, there is a former rocky island, a tuff cone 1702 m high which has been breached by the lake on the northeast side (Fig. 5.9). A wave-cut notch is preserved on its southwestern flank at 1666 m. As already mentioned on p. 17 the cone is one of a line of hyaloclastite centres running N 35° E towards the island of Galila. Two others, situated in the centre of Adamitulu Village and north of the prominent tuff cone, have been planed off at, or just below, 1670 m.

In contrast, on the western and southern flank of Alutu, the lake transgressed across a low-angle fan of pumice deposits and rarely impinged on harder rocks (Fig. 5.5). In such gently-sloping sectors the exact position of the shoreline is often hard to locate unless gully exposures are available, because it has been obscured by spreads of alluvium or slope-wash. But there is a well-defined terrace of cross-stratified deltaic sands and beach gravels contouring the southern side of the mountain at 1669-1672 m (Plates 25-26). The upper surface of this terrace often consists of silicified puddingstone; probably formed by seepage of hot, SiO_2-rich spring waters into the lake (E.F. Lloyd, 1977). Similar indurated deposits are found today around the shoreline springs in O-itu (United Nations, 1973, pp. 87-88).

e. <u>East and southeast Langano (areas 0738B sheet 4 and 0738D sheet 2)</u>

The palaeolake flooded the North Bay of Langano (O-itu) on the west and north, leaving only a temporary island on the western rim (>1640 m) and a long rocky peninsula (1774 m) linking the eastern rim to Alutu. There is no clearcut shoreline notch developed on the hard rhyolite of the inner caldera walls, and the limit of the lake shown on Fig. 5.6 has been inferred largely from the 1670 m contour on the 1:20,000 Ministry of Mines base maps.

East of O-itu the faulted relief becomes very complex (United Nations, 1973, fig. 43). No good exposures of lake deposits were found. The lake apparently penetrated up the Haro Bu-a graben, and flooded the upper end of the Kelbo fault trough to the west, as far as the foot of Alutu; leaving high and dry a very narrow, flat-floored graben at 1680-1690 m, at the foot of the outer caldera wall. The soils just outside the inferred limit at 0738-B30 and -B31 have very thick (3-4 m) red-brown profiles containing LSA (Late Stone Age) artifacts, which include large numbers of scrapers and backed blades. Superficially, they bear a strong resemblance to the soils at 0738-B1 and 0838-D1 (chapter 5C). But no detailed descriptions were made. Lloyd (1977) refers to an extensive siliceous sinter deposit at ca. 1680 m, just south of B30, which was probably formed by hot springs discharging into the lake.

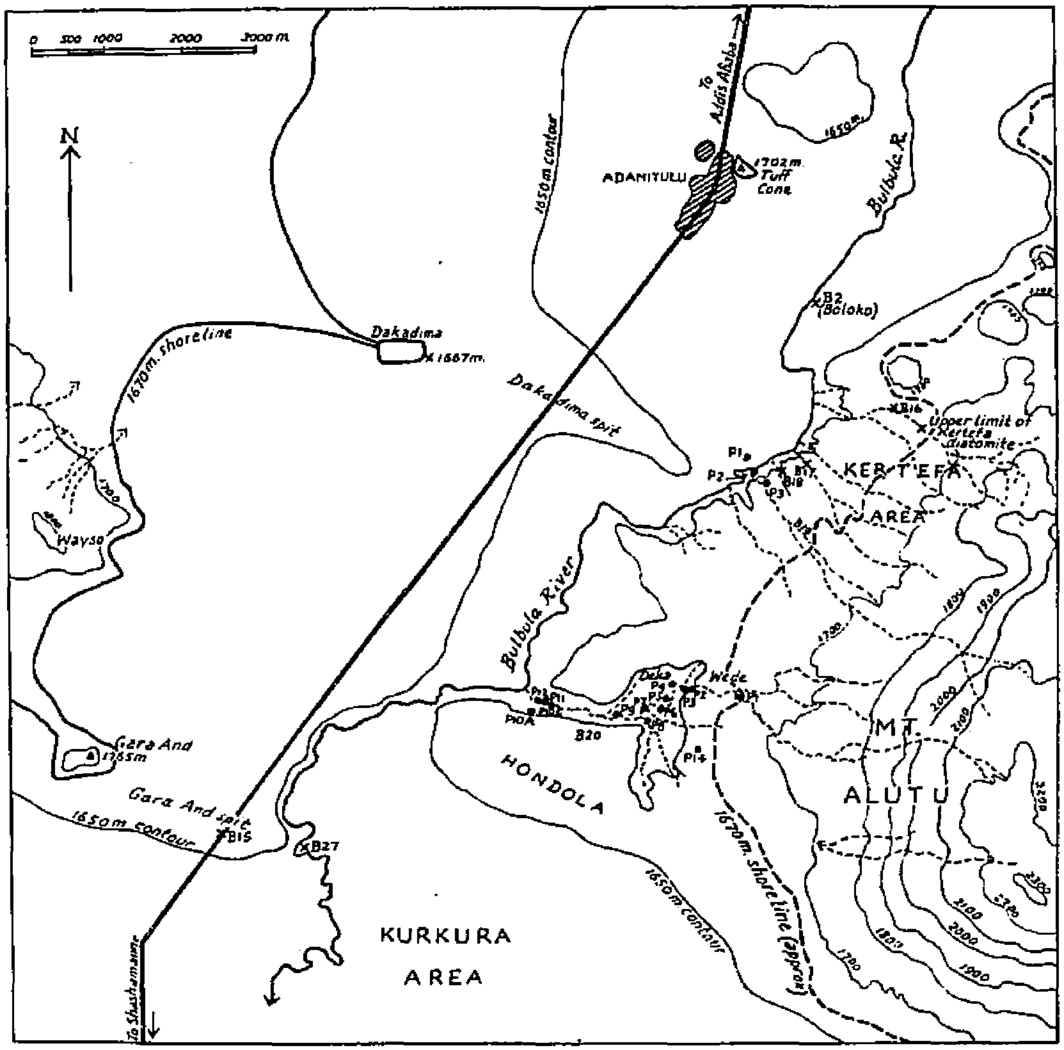


Fig. 5.5: Location map of the central Bulbula Plain showing the 1670 m shoreline.

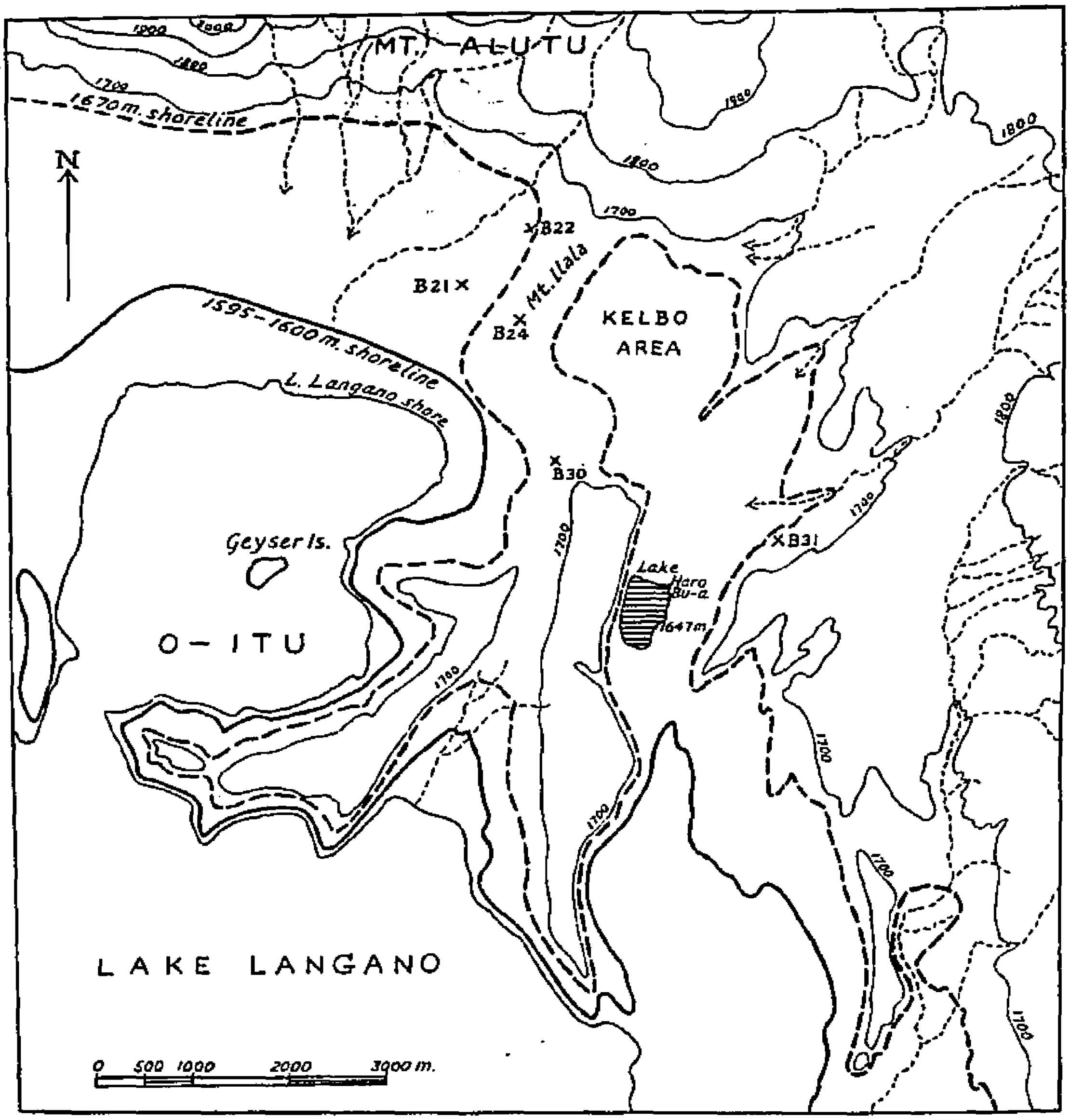


Fig. 5.6: Location map of the northeast Langano area showing the 1595-1600 m and 1670 m shorelines.

Haro Bu-a is a small, fault-bounded lake at 1647 m, surrounded by reed swamp. It shows promise as a site for pollen analysis and for reconstructing the chronology and palaeoecology of the lake-level maxima.

The eastern margin of Langano is thickly forested and even less accessible than the eastern side of Ziway. Once again, it is impossible to be sure of identifying the uppermost shoreline correctly where it parallels the strike of major faults. Some non-linear cliff-like features are apparent on the southern side of the lake, but in the absence of field confirmation the limit shown on Fig. 5.1 must be regarded as provisional.

f. The Langano-Abiyata-Shala isthmus (areas 0738B sheet 3 and 0738D sheet 1)

The geology of this rather complicated area, where ignimbrites and unwelded pumice breccias from the Shala caldera interfinger with ignimbrite sheets erupted (according to some authors) from the Munessa caldera on the present plateau, has been described by Di Paola (1972), Mohr et al. (1975), Lloyd (United Nations, 1973, and Lloyd, 1977) and Mohr (in press). The Wonji Fault Belt runs NNE from the northeast corner of Shala, creating a field of parallel horsts, grabens and open fissures (Mohr et al., 1975, fig. 3). The largest fault block, the Katlo horst, forms cliffs more than 65 m high along the west side of Langano.

Recessional strandlines in the grabens show that the ancient lake had a rather indented shore (Fig. 5.7). On the south side of Langano, cliffed promontories of soft "Munessa" ignimbrite alternated with gently shoaling embayments. The Galé horst (1714 m) stood up as an irregular island, whereas the Katlo horst was submerged. Two ^{14}C dates on shell have been obtained from sandy beach deposits underlying the strandlines in the Mirrga graben. An age of 5180 $\pm$ 55 BP (Q-1360) was measured on a sample from a culvert at 1668.5 m, which is very close to the overflow level (site 0738-D1). A second sample from 1651 m yielded an age of 6500 $\pm$ 120 BP (Grove et al., 1975). In the Ajewa embayment, northeast Shala, lacustrine diatomites (unit U2b) outcrop up to 1674 m, but their elevation has been disturbed by recent faulting on the lakeward side (chapter 6D; Fig. 6.13).

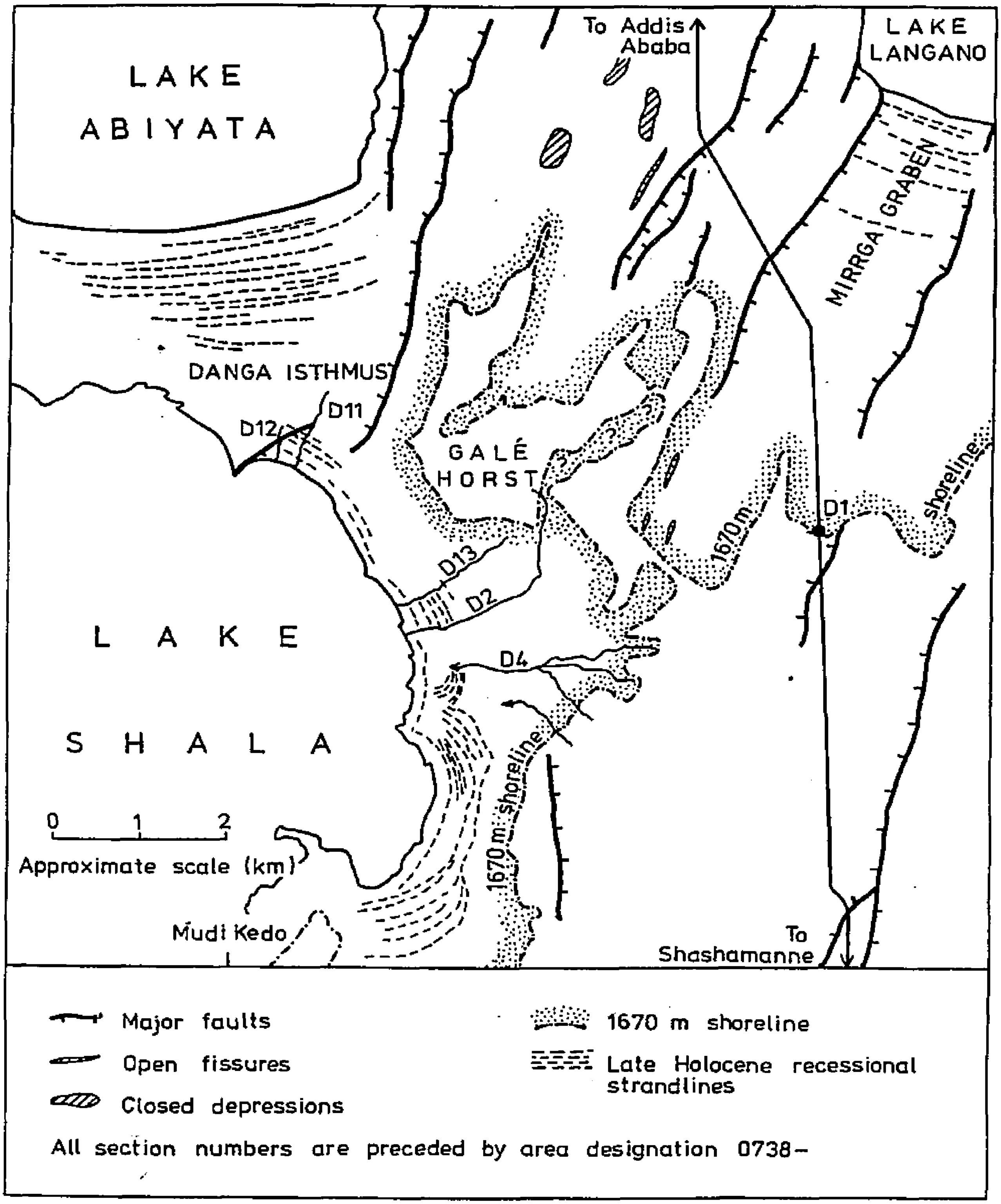


Fig. 5:7: Map of the Langano – Abiyata – Shala isthmus (from D.O.S. aerial photographs)

g. The Fiké-Lencha archipelago (areas 0738A sheet 4 and 0738B sheet 3)

At the culmination of the 1670 m transgression(s), the Abiyata-Shala isthmus was partly drowned, so that only the highest hilltops projected above the lake. The former islands fall into three geological categories (Mohr, in press): older volcanic centres, such as Fiké, Lencha Gudo and Lencha Tiko; large horsts such as Haroresa, Yebelo and Goljo; and younger volcanic centres (the only example on this side of the lake is Tulu Bila).

The first group was produced by explosive sublacustrine eruptions following caldera collapse. The best example is Mt. Fiké (32 km^2), which has a very prominent cliffline and wave-cut platform developed on its northern side (Plates 27-30). The two denuded Lencha cones have been described as "lacustrine guyots" by Lloyd (1977). They are built mainly of soft, yellow pumice-tuffs, which were easily abraded by the action of waves armed with hard pebbles of rhyolite: both centres are cliffed on their north and east sides. On Lencha Gudo (1750 m) there is a collapsed cave at the uppermost shoreline level. However the impressive rock shelter on Lencha Tiko (1740 m) opens at 1680 m (grid ref. DU 510393), and seems likely to have been formed by salt weathering as a result of lateral seepage of moisture through the rock, in the same way as the cave called "Shek Husen" at 1751 m on Mudi Kedo, rather than by wave attack. The susceptibility of the weakly consolidated pumice-tuffs to this kind of weathering is clearly demonstrated by the tafoni and basal undercutting on a group of stacks at DU 559300, just east of Haroresa (Plate 31). These pinnacles project from a horizontal rock platform at 1641 m. Although this elevation falls well short of the uppermost shoreline level, it seems quite likely that the stacks originated during a stillstand in the final retreat of the lake, and were subsequently moulded by subaerial weathering.

Unlike the Goljo horst, which is formed of resistant rhyolite, Haroresa (1780 m) shows a well-developed cliffline and wave-cut platform on its north-east side cut into soft tuffs and breccias. The small cinder cone called Tulu Bila (1760 m) (DU 497335) stood just on the margin of the lake. Its scoriaceous basalt flow has been partly buried beneath ca. 10 m of lake sediments, including a thick freshwater diatomite, which are exposed in the Mudi gully (0738-B14) at DU 485346. A Holocene age for the diatomite is probable, but not proven. Tulu Bila is joined to the eastern base of Fiké by a very subdued sand tombolo, delimited by the 1670 m contour. Younger recessional strandlines are much more clearly visible on air photos: on Lencha Tiko they show up as ridges of rounded rhyolite pebbles, and in the embayments as arcuate vegetation stripes.

Altitudes measured on the uppermost shoreline range from 1672-73 m on Fiké and Lencha Gudo to 1667 m on Haroresa. The validity of this discrepancy is dubious, but it may conceivably reflect slight updoming of the Fiké area. It could also be due to discrete fault movements, since Lloyd (1977) has found two small NNE-striking faults west of Lencha Tiko which cut Holocene sediments at DU 477389.

h. East, south and west Shala (areas 0738C sheet 2 and 0738D sheet 1)

Mapping the uppermost Holocene shoreline in this area presented few problems. It shows up very prominently on the new D.O.S. 1:50,000 maps, either as a bench running between 1660 and 1680 m, or as a bunching of the 1680, 1700 and 1720 m contours (Plate 32). A particularly striking bench at 1669 m on the Mudi Kedo horst (Plate 14) was first noticed by Mohr (1966). Best developed on the northern and eastern flanks, this wave-cut platform was eroded in massive ignimbrite, and is now covered by large boulders of pitchstone and ignimbrite interspersed with pumice shingle. Mudi Kedo is joined to the eastern caldera wall by a tombolo at about 1620 m, but horizontally-bedded beach gravels outcrop round its eastern slopes up to at least 1662 m.

The south shore of Shala is lined by precipitous cliffs and is largely inaccessible except by water. Fortunately, the southwestern embayment (Fig. 5.8) can be reached from the Shashamanne-Soddu road. Here the lake is intersected by the Shala-Corbetti sector of the Wonji Fault Belt, marked by basaltic flows and several eroded tuff rings. On the east side of the bay a well developed cliffline has been cut into soft pumice-tuffs at 1669 m. At

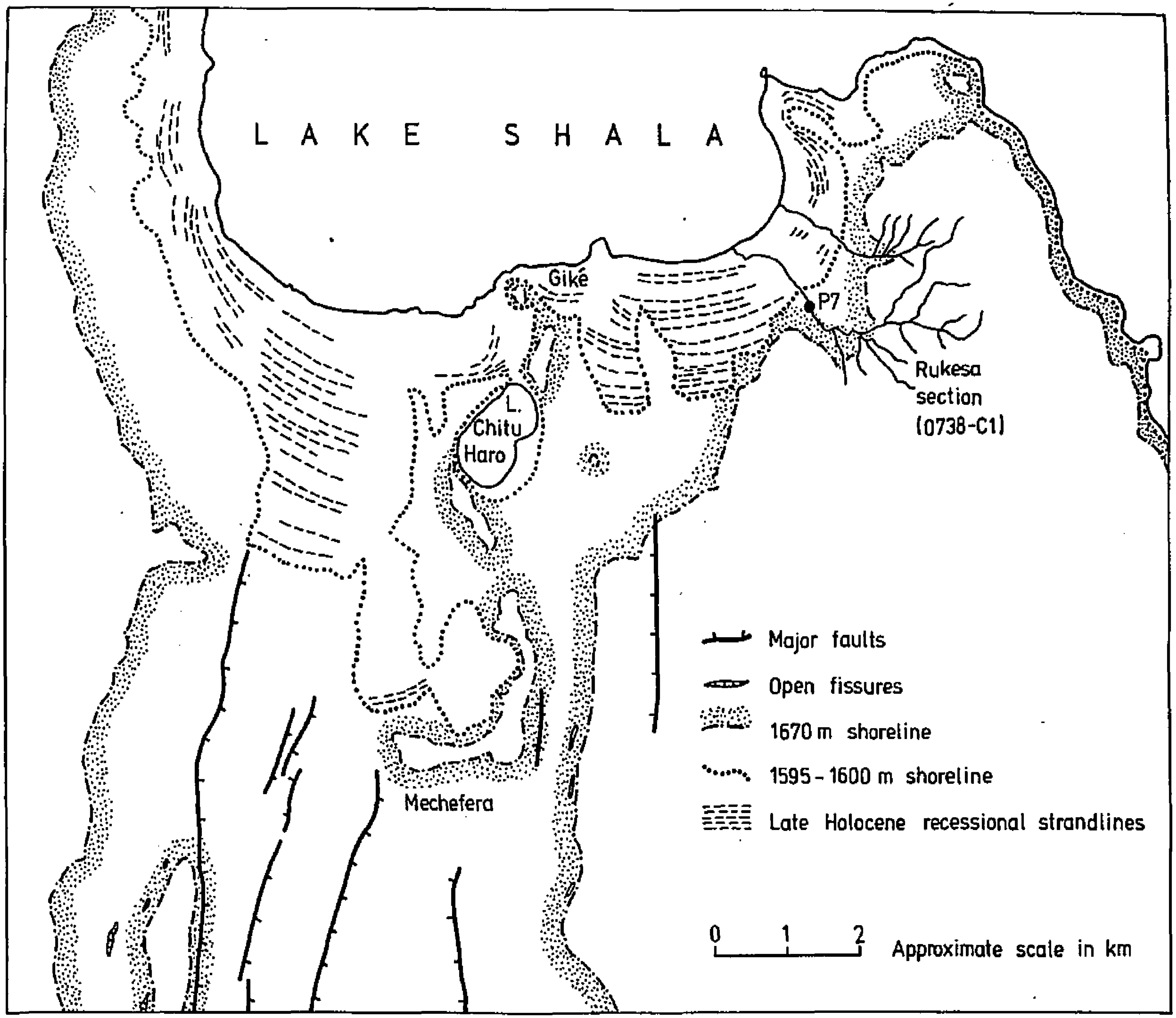


Fig. 5:8: Map of the Rukesa-Chitu Haro area, SW. Shala (from D.O.S. aerial photographs)

1629 m there is a prominent lower terrace.

The palaeolake penetrated more than 10 km up the fault troughs southwest of Shala, flooding the tuff rings Chitu Haro and Mechefera. The lowest point on the rim of Chitu Haro is < 1600 m. Greatly contorted diatomites are plastered over its inner walls, beneath pebbly strandlines of presumed late Holocene age (Plate 33). Mechefera is breached to the northwest and its floor is concealed by lake sediments.

The uppermost shoreline forms a pronounced cliffline all along the western side of Shala, petering out only where it crosses the Gidu River.

i. Northwest Abiyata and the western Bulbula Plain (areas 0738A sheet 4 and 0738B sheets 1, 3)

In this sector the broad sweep of the lake margin is easy to follow on photomosaics, although it is locally hidden by slopewash or river alluvium. Generally, the limit appears as a break of slope or a low cliffline. Running first NNE, it swings E and then SSE round the southern flank of Gademotta, towards the isolated hill Gara And (1765 m). Due north of Fiké, around Du 420540, traces of partly buried NNE-striking faults can be seen emerging through the sedimentary cover. Beyond Gara And, which was linked to the mainland by an ESE-trending tombolo, a sand-spit extended out into the middle of the Bulbula Plain at 1660 m (Plate 15). Whenever lake level dropped below 1652 m, which is the level of the Hondola ridge on the opposite side of the present river, the spit would have formed a barrier between the north and south basins of the palaeolake.

North of Gara And, the limit of the lake is cut into the gently sloping sedimentary infill of the Gademotta caldera (Fig. 5.9). It bears generally NNE, except where a tombolo linked the rhyolite plug Dakadima (1722 m) to the western shore. Once again the line of the tombolo was prolonged beyond the island by an ESE-trending spit. This meets the Bulbula River at about 1660 m. Laury and Albritton (1975) map this level as the lakeward limit of their terrace IV (Fig. 5.9).

Recessional strandlines are magnificently displayed over 200 km^2 south

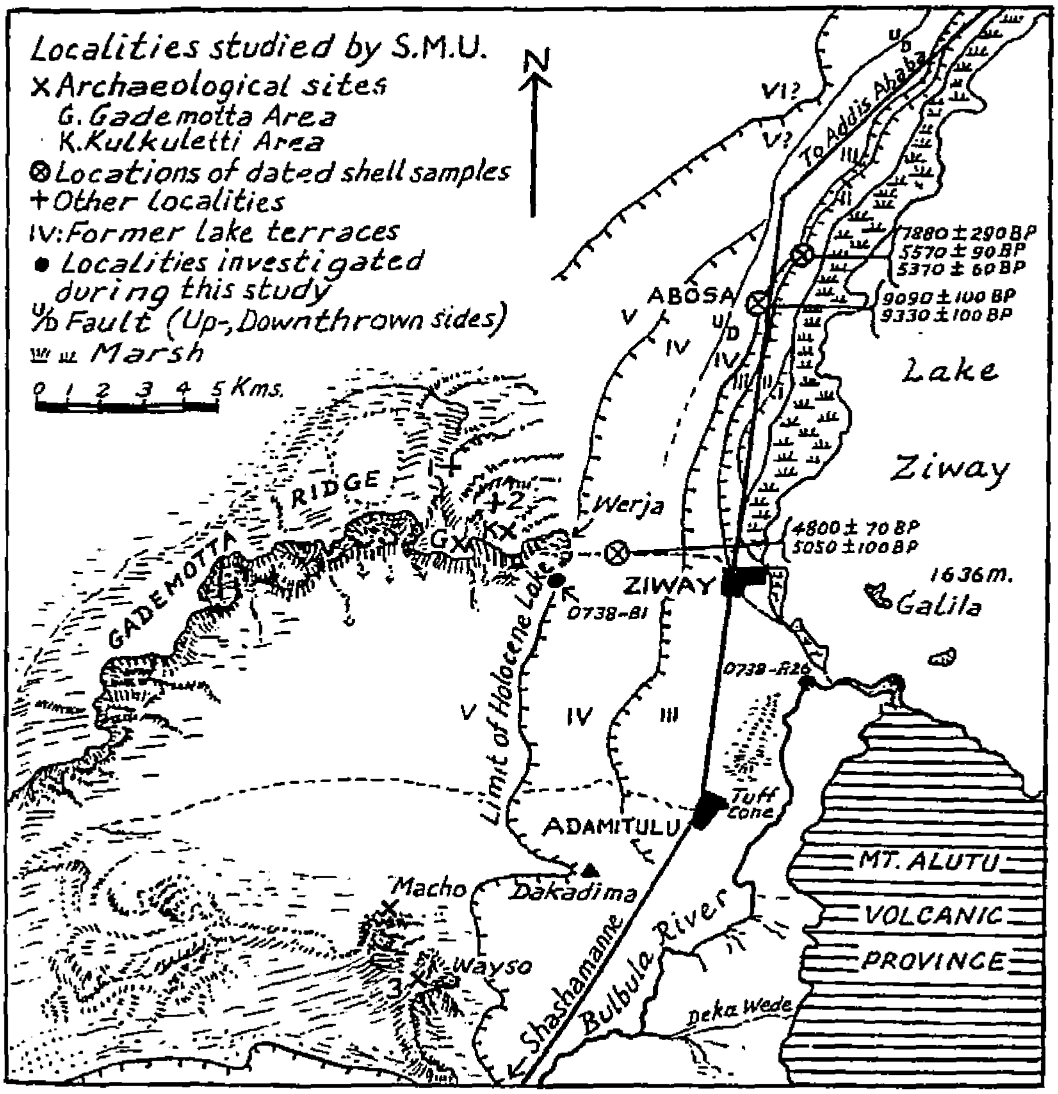


Fig. 5.9: Map of the West Ziway–Gademotta area (from Laury and Albritton, 1975 and Haynes and Haas, 1974, with additions).

and west of Gara And (Plate 34). They stop abruptly along the fault at DU 637663, which would have arrested longshore drift by northeasterly winds. Although the beach ridges here are unusually well preserved, they owe their prominence on aerial photos to the preference of Acacia trees for the tops of the ridges. Since the overall slope towards the lake is only 1:100 to 1:120, the strandline spacing of the order of 100-150 m, and the vertical contrast between crests and swales only 1-2 m, little can be seen on the ground.

The ridges seem to consist of asymmetric accumulations of pumice shingle and sand with the steepest face on the lakeward side, rather like those forming on the north shore of Abiyata today (see also Plate 18). At DU 433487 they enclose an oblong closed depression similar to those which are found around palaeolakes Bonneville and Lahontan. Probably it was once a shallow lagoon. The evenly spaced, parallel strandline curves provide little evidence for major oscillations of the lake shore. It seems that retreat from the 1670 m level was never interrupted for very long.

j. The West Ziway terraces (areas 0738B sheet 1 and 0838D sheets 3, 4)

The shorelines between Gademotta and the Meki have been described in detail by Laury and Albritton (1975) and Makin et al. (1976). Laury and Albritton identified five lake terraces, seen in cross-section in Fig. 5.10. Their excellent map (Fig. 5.9) needs little amendment.

Terrace IV corresponds to the overflow level. It follows the West Ziway fault from the Meki River to Abosa, and then swings SW across to Gademotta. Altitudes measured at the base of the IV/V cliffline range from 1669.5 to 1676 m (Fig. 5.2). The higher results were carefully checked, so recent uptilting of the fault block to the west seems a strong possibility. The lakeward edge of terrace IV follows the 1660 m map contour, and the foot of the III/IV riser is situated around 1654-58 m.

Laury and Albritton's terraces I to IV appear to be wave-cut benches eroded in partly cemented pre-Holocene lake sediments. Today the terrace surfaces are veneered by shelly sands and shingle of Holocene age. Samples of mollusca from terraces III and IV have been dated by Haynes and Haas (1974).

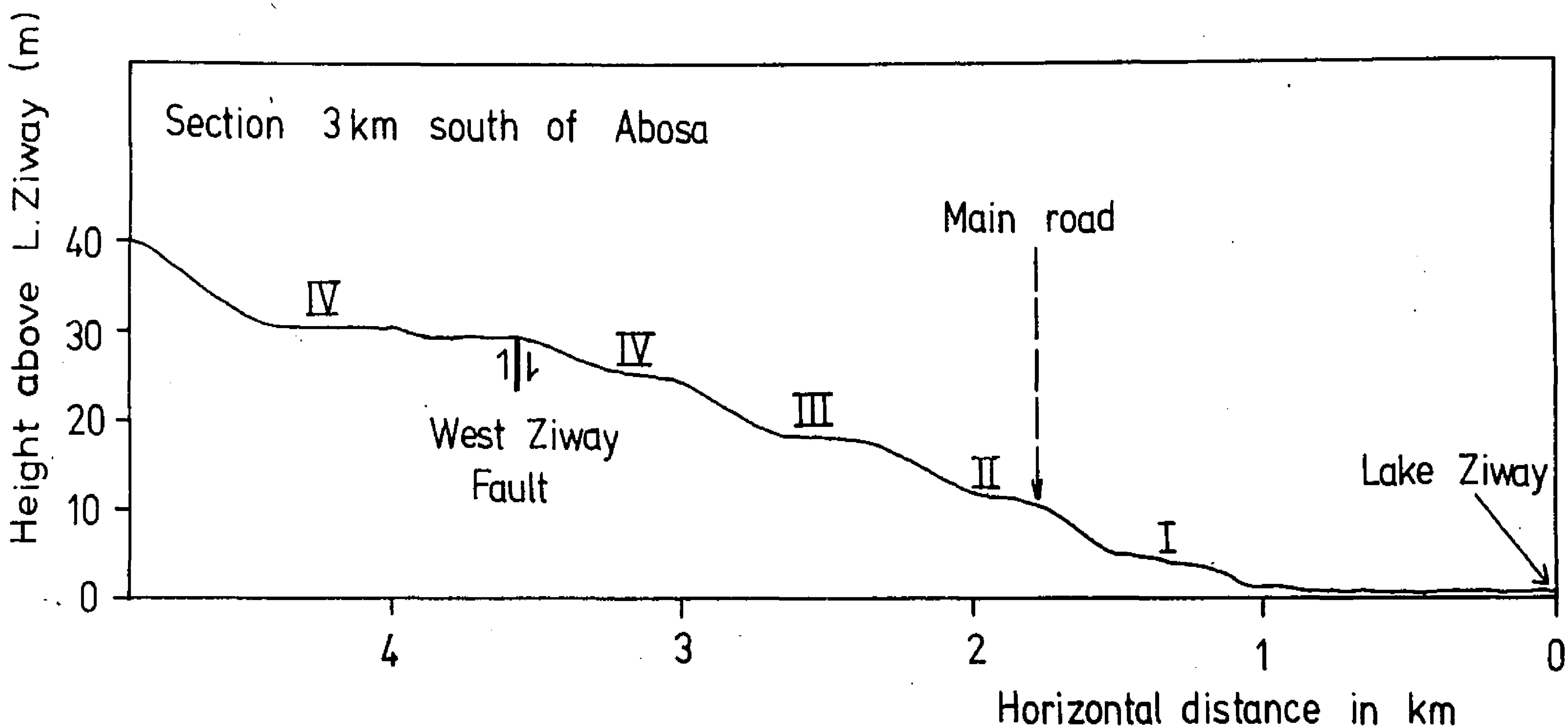


Fig. 5.10: Cross section of the West Ziway terraces (data from Makin et al. 1976).

The results are plotted on Fig. 5.9. Surprisingly, there is little relationship between age and elevation, although the dates clearly fall into two groups.[1] It is therefore uncertain whether or not the terraces represent a recessional sequence. Terrace III seems to coincide with the level at which Ziway separated off from the three southern lakes. The shells in it may well have been reworked from terrace IV during the final retreat of the lake.

In any case, we can conclude that the lake reached terrace IV at least twice during the Holocene. Since the lakeward edge of this terrace corresponds in elevation to the "bedrock" threshold of the basin, the lake probably overflowed during both episodes.

k. The elevation of the uppermost Holocene shoreline

All available altitude measurements on this shoreline are listed in Table 5.3. Only elevations obtained from erosional features or well-preserved depositional landforms (beaches and deltas) have been included. For simplicity, it is assumed that all errors result from measurement and are randomly distributed. The mean elevation, 1669.2 $\pm$ 0.8 m, is in excellent agreement with the evidence from the overflow channel (section Ia). Planimetric and bathymetric information on the 1670 m lake can be found in Table 8.3 and Figure 8.2.

1. The suspect sample SMU-66 (7880 $\pm$ 290 BP; Haynes and Haas, 1974) has been excluded from this discussion.

Table 5.3: Measured elevations on the uppermost Holocene shoreline (in m)

Elevation	Location	Feature	Grid Ref.	Survey Method
1670.8	W. Ziway fault scarp (0838-D10)	Cliffline	DV 786030	1
1670.9	Main road N. of Meki	Cliffline	DV 841051	1
1664.0	Lekanshu, S. Ziway	Cliffline	DU 764684	1
1663.1	Lekanshu, S. Ziway	Cliffline	DU 766681	1
1668.0	Awariftu gully, S. Ziway (0738-B29)	Top, terrace	DU 774667	1
1666.3	Adamitulu tuff cone	Wavecut platform	DU 679690	1
1672.3	Alutu piedmont (0738-B20/1)	Top, terrace	DU 684613	1
1668-6	O-itu, Mt. Ilala (0738-B22)	Top, terrace	DU 778540	1
1666.8	Haroresa	Cliffline	DU 548303	1*
1672.0	Lencha Gudo	Floor, collapsed cave	DU 518366	1
1673.4	Fike, N/W. side	Cliffline	DU 436367	1
1668.9	Mudi Kedo	Cliffline		1
1668.6	Rukesa, S/W. Shala	Cliffline		1*
1669.2	Gara And, E. side	Cliffline	DU 601598	1*
1667.1	Dakadima, E. side	Cliffline	DU 643664	1
1668.7	E.H.A. Quarry, Werja, W. of Ziway Town	Cliffline	DU 639774	1
1666.5	Werja, W. of Ziway Town	Cliffline		2
1666.5	3 km S. of Abosa, W. Ziway	Cliffline		2
1669.5	W. Ziway faultscarp (0838-D5)	Cliffline	DU 701909	1
1675.3	W. Ziway faultscarp (0838-D7)	Cliffline	DU 722938	1
1676.3	W. Ziway faultscarp (0838-D6)	Cliffline	DU 761995	1
	1669.2 $\pm$ 0.8 Average (n = 21)			

1 Altimeter traverse
2 Levelling traverse (Makin et al., 1976)
* Mean of two traverses

II. The 1595-1600 m united lake

During the last few thousand years a smaller lacustrine transgression linked the three southern lakes, but failed to reach Ziway. The maximum level attained was 1595-1600 m. This figure has been estimated by interpolation between benchmarks on the Addis Ababa-Shashamanne road south of Bulbula Village. In general the shoreline trace falls between the 1590 m and 1600 m contours on the 1:20,000 Ministry of Mines' maps.

Since Lake Ziway is impounded by a lava sill at 1635.6 m, its level is likely to have been little affected. Modern gauging records show that a rise in mean level by as little as 0.8 m would be sufficient to cause a six-fold increase in the monthly discharge at the Bulbula outlet (Makin et al., 1976), and initiate an expansion of the three southern lakes (section IIIb).

As the level of Lake Abiyata rose, it would unite with Lake Langano at 1582 m and begin overflowing into Shala at 1592.0 m. Above this elevation the evaporative surface would increase rapidly (Fig. 8.2): evidently the climatic anomaly was not large enough, or sufficiently prolonged, to increase lake area much beyond this point.

The geometry of the late Holocene palaeolake (Fig. 5.1) was first reconstructed by Gasse and Street (1978a). Its shorelines are very striking, and the beach ridges can often be followed on the ground. The only area where some doubt arises is the densely forested eastern shore of Langano. Estimates of the area and volume of the lake are given in Table 8.2. At maximum, Langano, Abiyata and Shala were separated by sand and shingle spits (Plates 27, 34), which appear to follow structural trends in the underlying bedrock. The Algé spit to the west of Fiké was first mapped by Gèze (1974, fig. 49).

Cliffed shorelines were rare around the 1595-1600 m palaeolake, except for the soft tuffs along the western and southeastern shores of Shala (Plate 32). Probably erosional features did not have sufficient time to develop. Low beach ridges are far more common, and are particularly obvious round the northern sides of Abiyata and Langano, and the southern tip of Abiyata, where the materials were reworked from older Holocene sediments. Small deltas formed off the mouths of many ephemeral streams (Plate 35). Along the southern shores of Abiyata and southwest Shala, windblown sand has accumulated on the ridges to give parallel dunes and slacks (Plate 7).

The strandlines run broadly parallel to the recessionals of the 1670 m lake. Around the northern margins of Langano and Abiyata they contrast strikingly with the older shorelines, due to an almost complete absence of woody vegetation. That this is not due to very recent formation is clear from the dense forest cover developed on similar ridges round southeast Langano and the Adabar embayment. Soil mapping by L.R.D. shows that the northern shores of Abiyata and Langano are underlain by solonetz soils (Makin et al., 1975). High salinity seems a more likely explanation for the absence of trees (Western and Van Praet, 1973), especially as large *Ficus* spp. are able to grow along the banks of the relatively fresh Bulbula inflow.

To the west of Fiké the Gidu River has deposited the extensive Gifato delta which was exposed by the fall from the 1595-1600 m level. This delta has not been studied in the field, but a map can be found in Gèze (1974, fig. 49). There are obvious morphological similarities with the late Holocene Meki Delta (Fig. 2.9). Traces of a fossil birds-foot distributary system can be seen below 1580 m northeast of the present Gidu mouth.

III. Minor historic fluctuations of the lakes

Apart from Lake Shala, little is known of the fluctuations of the lakes during the period of written records. Water levels have only been gauged since 1967 on Lake Ziway and since 1968 on Langano and Abiyata. Lake Ziway was known to the Portuguese missionary Manoel de Almeida around 1628, when he wrote that the Haoax [Awash] 'receives the waters of a big river named Machŷ [Meki] coming from Lake Zuaĵ [Ziway]' (Beckingham and Huntingford, 1954). This intriguing idea is also embodied in Coronelli's *Atlante Veneto* (1690) (Grove et al., 1975, plate VIIa), but should probably be regarded with scepticism because of the magnitude of the change involved.

Other travellers' descriptions are not generally very useful. Some information can be gleaned from Nilsson's description of Lake Shala (1940) if this is compared with recent field evidence. Scattered observations were also made on Lake Abiyata prior to 1968.

a. Lake Shala

Nilsson reported finding dead trees up to 8 m above the 1933 level of Lake Shala (1940, fig. 28), demonstrating that the water surface had previously risen by at least 8 m and then fallen again. By November 1974, the lake had once again submerged many of these trees, indicating a net rise of 5.5 m since 1933. The local people had cut down the trunks for firewood (Plate 36), but the highest carbonized stump was found to be 2.5 m above lake level at site 0738-D3 (Fig. 6.12a) in 1974.

Additional evidence for a recent rise can be found in many places around the eastern shores of the lake. Emerged Gilbert-type deltas are very common at the mouths of small gullies. In section 0738-D2, for example, cross-bedded sands extend up to 5.65 m above the water surface of 14/10/74 (Plate 68). On the small headland southeast of Haroresa, low-level bouldery strandlines contour the slopes below +5.1 m. And at site 0738-D9 (Fig. 6.12a), tufa deposits mark the site of inactive spring heads which lie 5.8 m higher than the present hot springs. All indications are that a recent transgression of Lake Shala reached about 5.8 m above its 1974 level; in other words, 11.3 m higher than in 1933.

When did this rise take place? In the absence of radiocarbon dates, we can guess that it occurred towards the close of the 19th century, when high levels were recorded in Lakes Abhé (Gasse, 1975), Chad (Maley, 1973), Chew Bahir, Turkana (Butzer, 1971), Rukwa (Gunn, 1973), Malawi, Tanganyika and Victoria (Lamb, 1966). But it could have taken place much earlier, since drowned trees around Lake Wonchi have been dated at 1400 ± 140 BP (Smeds, 1964).

There is no evidence for any significant dilution of Lake Shala during the pre-1933 rise. No unrolled mollusca or fish remains have been found in the deltaic deposits. However some mystery surrounds a sample of wood, collected on behalf of H. Smeds and purporting to come from the drowned trees, which was identified as the leguminous shrub Aeschynomene elaphroxylon (Guill. and Perr.) Taub. (Smeds, 1964). This is a surprising conclusion, since Aeschynomene typically grows on the outer fringes of freshwater swamps. It is common around oligohaline lakes such as Ziway, Abaya, Naivasha, Baringo and Chad; and in the White Nile swamps (Beadle, 1932; Brunelli et al., 1941; Rzóska, 1976). A second sample of wood collected from Lake Shala on 10/10/74 has been identified by the Jodrell Laboratory at Kew as Acacia sp., probably Acacia nilotica (L.) Willd. ex Del. Whereas Aeschynomene has a very soft wood like balsa, which rots easily, Acacia spp. have much tougher wood. Acacia nilotica is common on the margins of lakes, pools and rivers in Africa. It is able to tolerate seasonal flooding for several months each year (G. Wickens, in litt. 8/2/77). In addition, since it does not imply any marked decrease in the salinity of the lake, Acacia nilotica seems a more likely candidate. Perhaps Smeds' friend collected his wood sample from Lake Ziway, mistakenly or otherwise!

The cause of these fluctuations in the level of Lake Shala presumably lies in changing inflows from the Gidu and Adabar drainages. One slight clue suggests that Lake Abiyata, and therefore the northern rivers, responded in a similar way: in 1926 Omer-Cooper found dead trees on the shore of Abiyata, and inferred

> that within the last few decades the waters had reached an abnormal level, and that previously its waters had been lower for some period sufficiently long to allow the soil to become sweet and trees to grow to moderate dimensions.
> (1930, p. 200)

Unfortunately, there is no way to relate his observations to the 1968 datum.

b. Lakes Ziway, Abiyata and Langano

Since 1900, rainfall totals over the Ethiopian Highlands have varied markedly from decade to decade (Wood and Lovett, 1974; Kingham, 1975). This is clearly reflected in the hydrographs of the major rivers feeding the lakes (Fig. 5.11). There has also been considerable variation in the relative

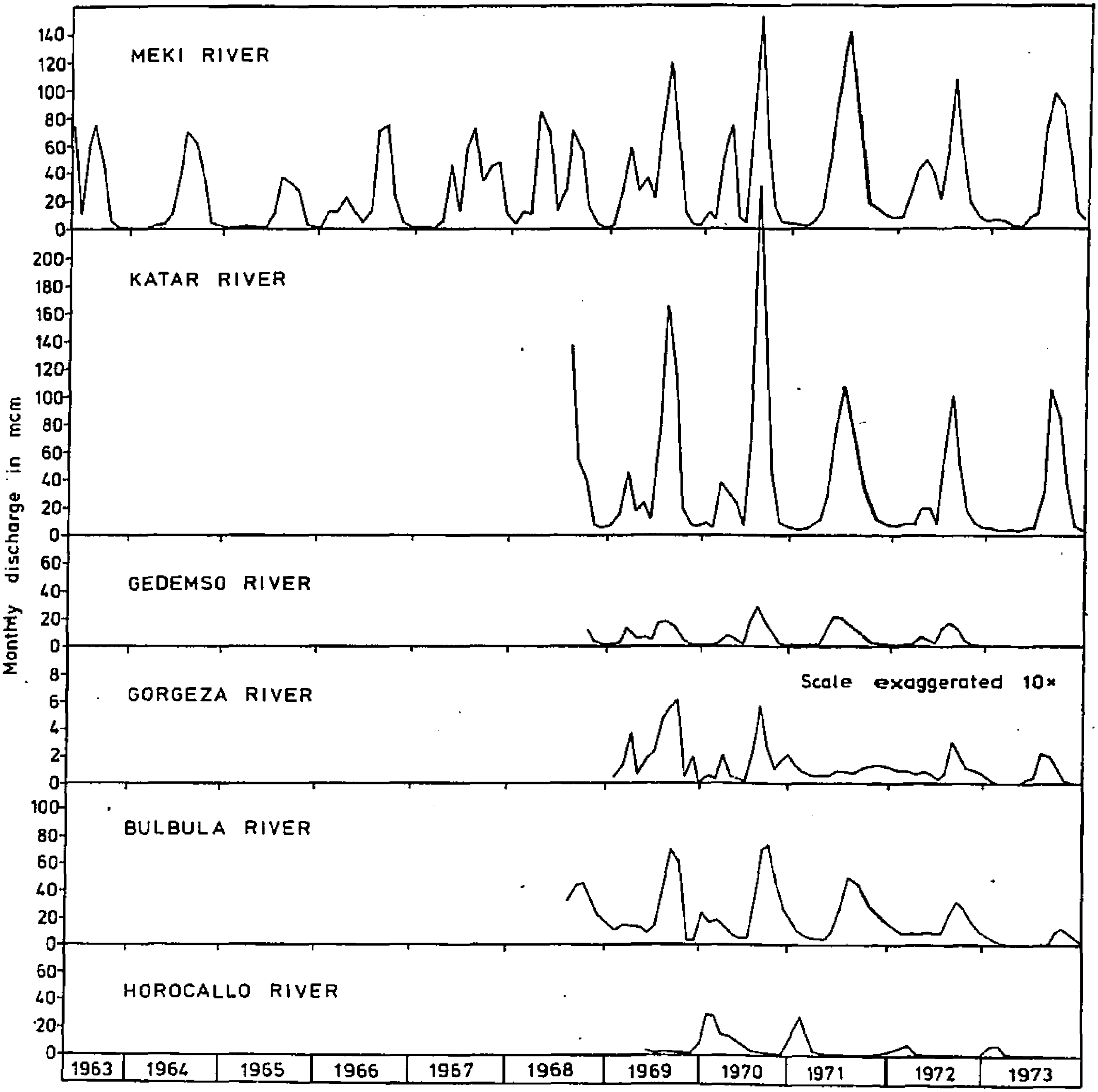


Fig. 5.11: Hydrographs of the rivers in the Ziway-Shala Basin, 1963 - 1973 (data from Makin et al. 1976).

importance of the two rainy seasons.

The annual discharge of the Meki at Meki Town rose from a minimum of 121-5 mcm in 1965 to a maximum of 486.4 mcm in 1970 (Fig. 5.11). By 1973 it had fallen to 362.4 mcm. Likewise, the Katar reached a peak of 618.1 mcm in 1970 compared with 261.6 mcm in 1974, and the Gedemso River (one of the principal affluents of Lake Langano) carried 94.9 mcm in 1970 compared with 66.4 mcm in the latest year for which data are available (1972). The input of the Gorgeza spillway into Lake Abiyata has also fallen from 28.7 mcm in 1969 to 1.2 mcm in 1974.

The response of the lakes to these fluctuating inputs can be seen in Fig. 5.12. The two lakes with outlets, Ziway and Langano, oscillated by only 1.97 and 1.67 m respectively between 1967/8 and 1974. However, both reached a rainy season maximum in 1969. Since then their mean levels have declined by 0.172 m/y and 0.152 m/y respectively. This has caused a very marked reduction in annual discharge in the Bulbula and Horocallo rivers (Fig. 5.11), since outflow is an exponential function of water-surface height above a critical threshold (Makin _et al._, 1976, figs. 11 and 26).

The response of Lake Abiyata has been cumulative, which is typical of a terminal lake. The gauging-staff record in Fig. 5.12 can be extended using evidence obtained from the 1956 Huntings aerial survey (Makin _et al._, 1976, p.72), and observations made by Dr Emil Urban in September 1964 and March 1967 (_in litt._ 23/9/75). Lake level was quite low in 1956, and again in 1964. Between 1964 and 1967 it receded a further 1-2 m. Beginning in June 1967 the water level rose dramatically at a rate of 1.9 m/y, reaching 1579.2 m by April 1970 and drowning a large number of _Acacia tortilis_ trees near the mouth of the Bulbula. It then continued high for three years, reaching its peak (1579.4 m) in November 1971. This was some 1.8 m less than the greatest depth encountered (14.2 m) in 1938 (Vatova, 1941). Since October 1972 the lake has fallen by about 1.1 m/y. Figure 5.13 portrays the variations in the shoreline of Lake Abiyata since 1956.

Comparison of the fluctuations of Abiyata since 1968 with the Meki River hydrograph suggests that its level lags only 1-2 years behind the variations in

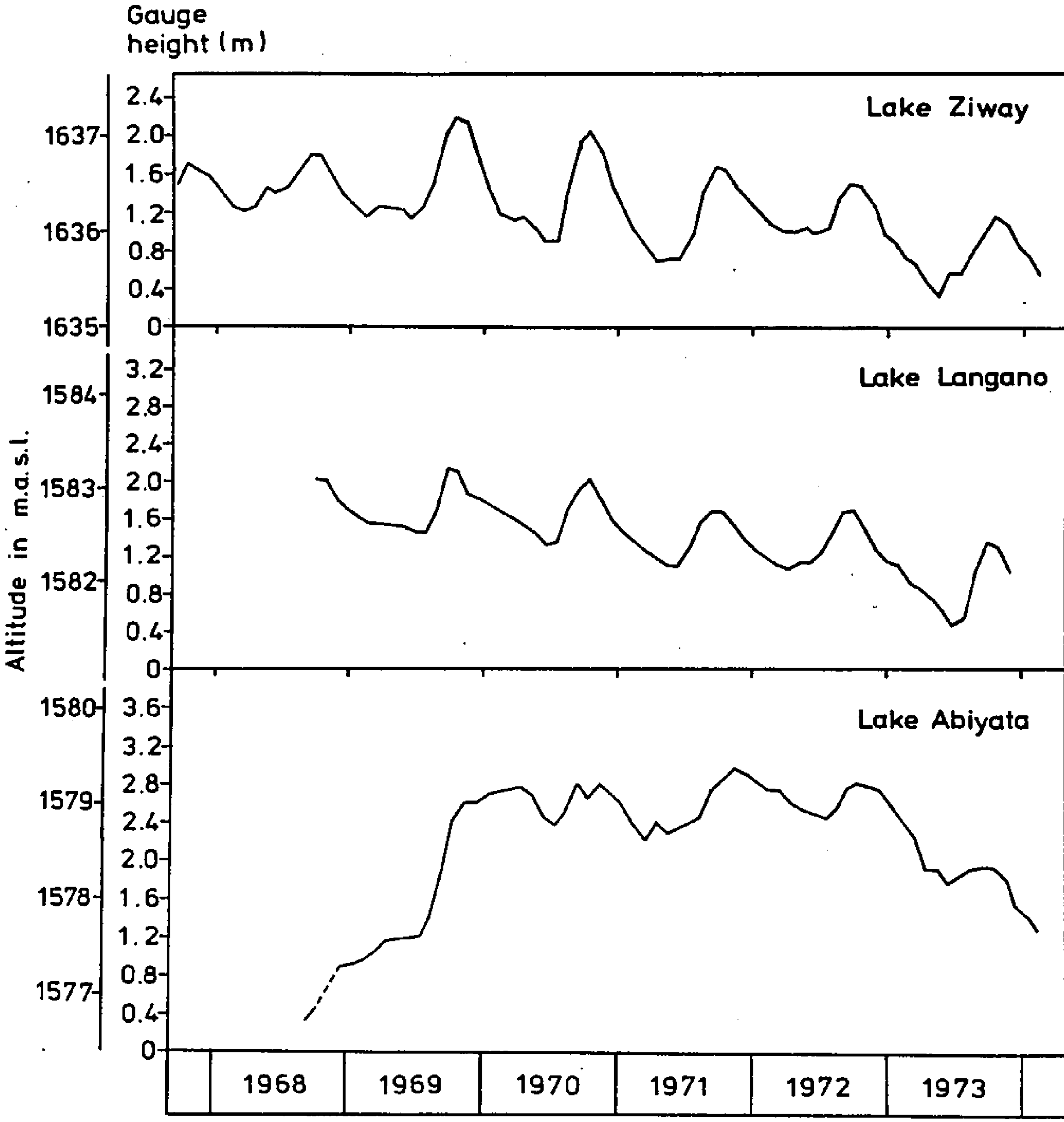


Fig. 5.12: Water-level fluctuations of Lakes Ziway, Langano and Abiyata, 1967-1974 (data from Makin <u>et al</u>. 1976).

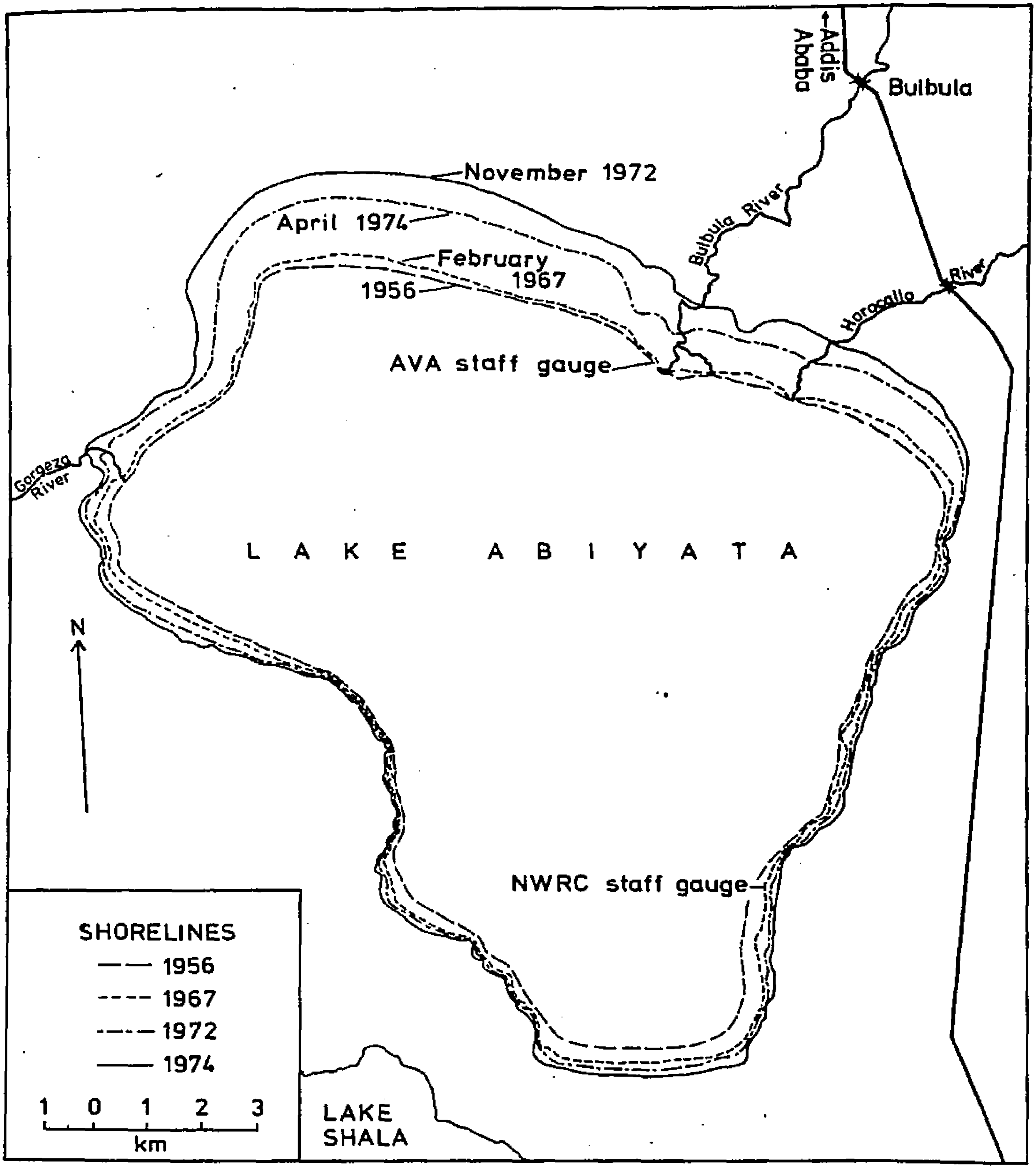


Fig. 5.13: Fluctuations in the extent of Lake Abiyata since 1956 (data from Makin <u>et al</u>. 1976).

river input into the higher lakes. It is interesting to compare the earlier part of the Abiyata curve with the simulated mean annual lake-level curve for Lake Ziway for the period 1954-1973 (Fig. 5.14). The latter was computed by extending the river-discharge series using the rainfall records from Addis Ababa (for the Meki Basin) and the Katar catchment, and then feeding the results into a calibrated water-balance model of the lake (McKerchar and Douglas, 1974). Figure 5.14 correctly predicts the level of Ziway shown on the 1956 Huntings photos to be slightly higher than in November 1972. The agreement with Lake Abiyata after a lag of about two years is allowed for is surprisingly good. Looking even further back, one can see that maxima and minima of lake level follow shortly after peaks and troughs in the Addis Ababa rainfall record (Fig. 5.14), suggesting that rainfall variations are the primary cause of changes in lake storage at the present day.

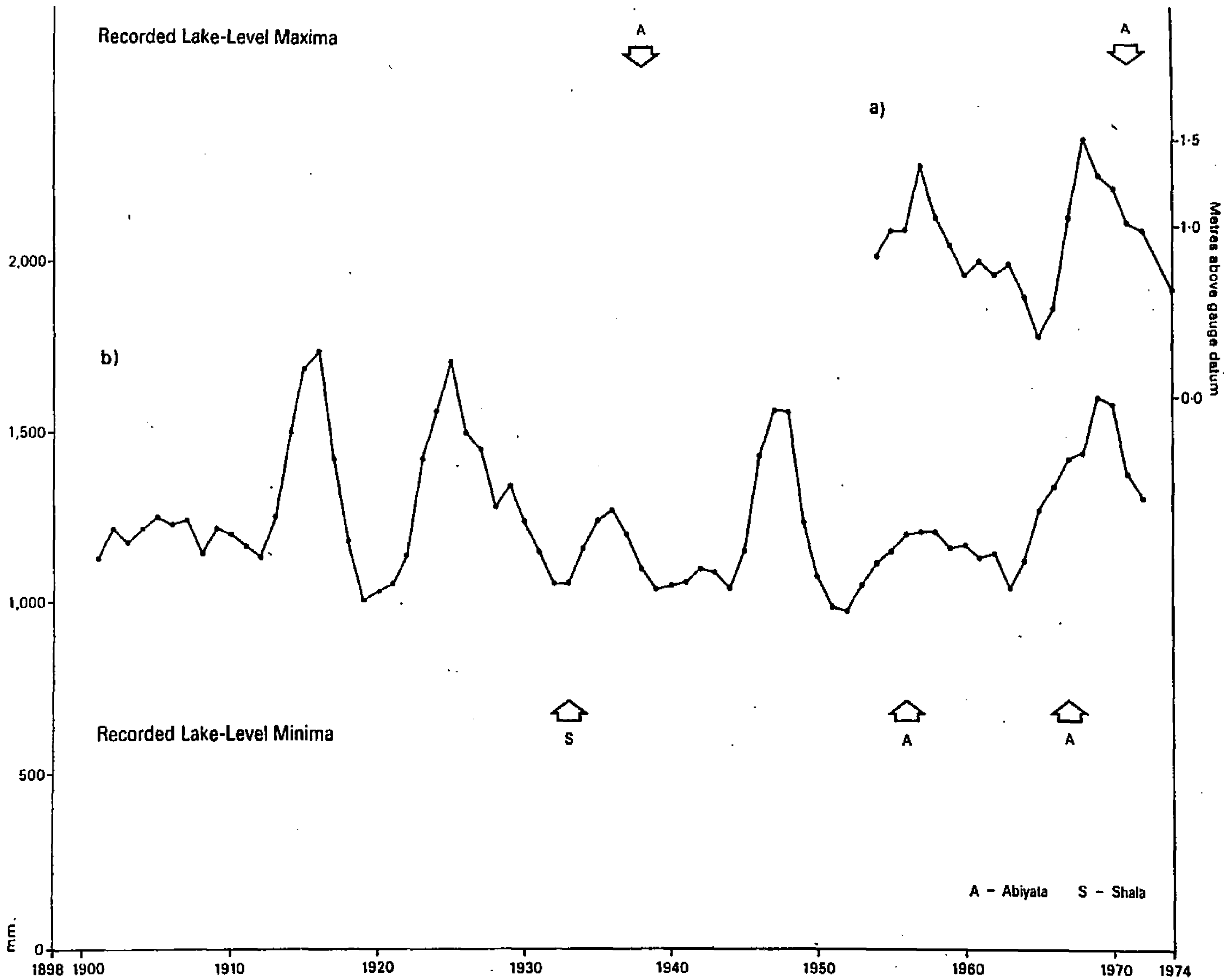
IV. General characteristics of the Holocene lake shorelines

a. Identification

The lake shoreline in the Ziway-Shala Basin are not as impressive on the ground as those described around palaeolakes Bonneville (Gilbert, 1885, 1890), Lisan (Bowman, 1971) and Abhé (Gasse, 1975). This can be attributed to a number of factors: a smaller fetch than in the first three cases; a more humid climate, which favours burial of the beach sediments by fine slope-wash; the dominance of readily-comminuted pyroclastics; and the relative scarcity of algal tufas and stromatolites. The mapping problems which arise are very similar to those encounted by C. Washbourn-Kamau in the Nakuru-Elmenteita and Naivasha Basins, in a comparable geologic and climatic setting (Washbourn, 1967a,b; Washbourn-Kamau, 1971; Kamau, 1977).

The former shorelines are most difficult to recognize along steep fault scarps in resistant rocks, where wave action made little impact, and on gentle piedmont slopes where they have been covered by recent deposits. The striking appearance of the recessional strandlines on aerial photographs results primarily from contrasts in soil moisture rather than topographic

Fig. 5.14: Recorded maxima and minima of Lakes Abiyata and Shala compared with a) the simulated mean lake-level curve for Lake Ziway (McKerchar and Douglas, 1974) and b) 3-year running means of rainfall at Addis Ababa since 1898 (Wood and Lovett, 1974).



Recorded Lake-Level Maxima
A
A
a)
1·5
1·0
0·0
Metres above gauge datum
b)
2,000
1,500
1,000
500
0
mm.
Recorded Lake-Level Minima
S
A
A
A – Abiyata
S – Shala
1898
1900
1910
1920
1930
1940
1950
1960
1970
1974

expression. By 1971, woodland clearance had already effaced many of the features visible in 1956. Because the beach ridges are so subdued, sedimentary exposures are particularly helpful in identifying the limit of the lake in the field.

b. Classification

Following Gilbert (1885), the main shoreline types can be conveniently classified according to the steepness of the topography and the resistance of the underlying rocks (Table 5.4). Exposure to wave action is a third important variable (McLachlan and McLachlan, 1971). During the 1595-1600 m highstand and the retreat from the 1670 m highstand, only types 2 and 3 were developed, presumably because the lake rarely impinged on solid rock. Surprisingly, there do not seem to be any large delta complexes related to the 1670 m lake level. This situation apparently arose because the Meki and Katar Rivers deposited most of their sediment load upstream from the main lake (p.116 and chapter 6). Some part of the Gifato delta may have originated during the 1670 m phase(s), for example the light yellowish-brown silts exposed along the Gorgeza channel at 0738-A1, but the surface morphology of the fossil delta is almost certainly younger than the 1590-1600 m highstand.

c. Orientation with respect to prevailing winds

The Holocene lake shorelines show a strong relationship to the prevailing dry-season (NE) and wet-season (SW) winds, which are channelled along the Rift. In the future it may prove possible to make quantitative comparisons between the alignment of features of given ages and the present-day wind field, but this cannot be attempted until the local weather records kept in Addis Ababa become accessible once more.

The relationship is clearest in the case of beach ridges, which show a marked tendency to align themselves NW/SE at all elevations. For example, at the north end of Ziway and north of Abiyata, the mean orientation of the 1670 m shoreline is S 57° E and S 66° E respectively, at right angles to the long axis of the basin (N 31° E). This direction is not in fact perpendicular

Table 5.4: <u>Classification of Holocene Lake Shorelines</u>

1) <u>Steep slopes</u>

a) <u>Resistant rocks</u>

Rhyolites and strongly-welded ignimbrites

- little modification
  - parts of the N. and S. Shala caldera walls, E. Ziway, O-itu, NE. and E. Langano.

b) <u>Less resistant rocks</u>

Unwelded or weakly-welded rhyolitic and basaltic tuffs

- high cliffs, rock-cut platforms and stacks, gravel terraces
  - S. Langano, parts of E., W. and SW. Shala caldera walls, Adamitulu tuff cone, outlying volcanic centres N. of Alutu, Fiké, Lencha Tiko, Lencha Gudo, Haroresa, Mudi Kedo.

2) <u>Intermediate slopes</u> (greater than 1:30)*

Unconsolidated or weakly-consolidated lake deposits, pumice, colluvium and alluvium

- wave-built terraces or wave-cut platforms backed by low cliffs; small Gilbert-type deltas
  - W., NW. and SE. Ziway, SW. and S. Alutu.

3) <u>Gentle slopes</u> (less than 1:30)

Unconsolidated or weakly-consolidated lake deposits, pumice, colluvium and alluvium

- low constructional ridges, spits and tombolos
  - NE. Ziway (lower slopes of Mt. Bora), W. Bulbula Plain (Dakadima and Gara And), N. Abiyata, S. Abiyata (S. tip and E. of Fiké), NW. Langano, SW. Langano (Mirrga graben), SW. Shala (W. of Chitu Haro)

* This limit is approximate, since it varies with exposure.

to the theoretical line of maximum fetch (99 km, N 42° E), but it corresponds fairly closely to the longest stretches of water uninterrupted by islands or shoals. The effective maximum fetch in both basins of the 1670m lake, together with predicted wave characteristics at different wind speeds, is given in Table 8.2.

The orientation of the 1595-1600 m beach ridges is very similar to the older 1670 m shorelines, except for a few areas like the southern tip of Abiyata where the topography was radically different at the lower level.

The dominance of the strong northeasterly winter winds over the weaker southwesterlies is shown by the greater width of the 1670 m wave-cut bench on the northeast side of islands such as Mudi Kedo, the Lencha Mountains and Haroresa. The lake has breached many of the tuff rings and cones on their windward side (Table 5.5); a phenomenon first observed in a marine context by Charles Darwin during his visit to the Galapagos Islands (Darwin, 1845, ch. XVII). However, these volcanic cones may originally have been somewhat asymmetrical, due to the effect of the prevailing wind at the time of eruption. A net southwestwards drift of beach materials is indicated by the abrupt way in which the recessionals northwest of Abiyata are terminated on the northeast side of buried faults.

Table 5.5: Direction of breaching of volcanic centres within the limits of the 1670 m lake

Centre	Breached to:	Grid ref.
Galila	NE	DU 730758
Horseshoe-shaped island near Chefe Jila	NNW	DU 820680
Adamitulu	NE	DU 680680
Fiké	N	DU 443354
Mechefera	NW	DU 355155
Giké	N	DU 365195

d. Relation to the drainage network

Grove et al. (1975, p.185) noted that ephemeral gullies on Mt. Bora, Mt. Alutu and the slopes NW of Abiyata tended to peter out at the 1670 m shoreline. Washbourn-Kamau (1971) had previously suggested that similar gullies in the Nakuru-Elmenteita Basin were cut by increased runoff when the lake stood at its highest level. Her hypothesis finds support in the Afar, where fossil Holocene deltas are common at the mouths of now inactive wadis (Gasse, 1975; Gasse and Street, 1978a). However, the new 1:50,000 map series of the Ziway-Shala Basin reveals that the relationship of the drainage net to the 1670 m shoreline is much less clearcut. It is very common for gullies to dry up at the foot of fault scarps wherever they encounter permeable deposits. On the Koshé sheet (area 0838D no. 3), for example, many parallel streams descending the eastern slope of the Dugda horst peter out between 1800 and 1740 m. The picture is further complicated because drainage incision is known to have been influenced by changes in base level (chapter 6 CIII).

It is therefore concluded that the drainage pattern does not provide acceptable evidence for increased local runoff during the 1670 m highstand(s), although such an increase appears likely on other grounds (chapter 9).

e. Isostatic deformation

The American geologist G.K. Gilbert was the first to recognize that the earth's crust is isostatically depressed below large waterbodies. In the best-known case, that of Lake Bonneville (Gilbert, 1890, pl. 46; Crittenden, 1963a,b), the Late Pleistocene Bonneville shoreline is now upwarped by as much as 64 m in the centre of the basin, corresponding to a maximum water depth of 335 m. This represents about 75% isostatic compensation of the former water load.

The maximum expectable deformation is given by the equation $w = h/\rho'$, where h is the depth of the water column and ρ' is the effective density of the $/\rho'$, mantle rocks being displaced (Crittenden, 1963b). (The specific gravity of lake water is assumed to be 1.00.) If the density of the substratum is taken as 3.1 g/cm^3 (Searle and Gouin, 1972), the potential deformation of the 1670 m shoreline in the Shala Basin becomes 36 m. Even allowing for the limited precision of altimeter measurements it ought to be possible to detect upwarping of this magnitude. As Figure 5.2 shows, however, the 1670 m shoreline can be regarded as horizontal within the limits of measurement, except for localized and possibly nonsignificant displacements along faults in the Fiké and West Ziway areas.

Several factors may contribute to the observed lack of deformation (Crittenden, 1963b). Firstly, the above computation ignores the rigidity of the crust, which tends to distribute the effects of the water load over a wider area and reduce the central depression (Andrews, 1975, ch. 7). Secondly, the expected deflection would be smaller if displacement of mantle rock occurred at depths below 75 km, where its effective density is greater. Thirdly, the fluctuations in lake level may have been too rapid for equilibrium to be achieved. Crittenden (1963a,b) calculated a relaxation time (the time taken for the crustal deflection to decrease to 1/e of its original value) of 4000 - 10,000 years for the Bonneville area, which makes the attainment of equilibrium in the Ziway-Shala Basin very improbable (Fig. 8.1). Finally, even if isostatic depression did occur, recovery may still be incomplete.

5C EVIDENCE FOR PLEISTOCENE LAKE LEVELS

I. Lowstands

Bathymetric maps of the lakes reveal little about past lowstands. The sole exception is a winding depression on the bed of Lake Ziway up to 2 km wide and 8.95 m deep, which figures on the Italconsult map (1970). It runs from the Katar and Meki Deltas across to the island of Galila, and appears to be an extension of the Bulbula gorge.

The age of this submerged channel is impossible to establish. It may well predate the lava flow which dams the lake, and even Galila itself; but a great age seems unlikely, since the trough has not yet been infilled by sediment. Lloyd (1977) has suggested that it dates back to the "terminal Pleistocene" lowstand of the lakes (chapter 8B).

II. Highstands

a. Introduction

The earliest strandline remnants in the basin are the rhyolite gravels of probable Middle Pleistocene age on the flanks of Gademotta caldera. These were first discovered by Nilsson (1940, p.39). The new evidence gathered by Southern Methodist University expeditions has been summarized in chapter 3C. Lloyd (1977) has recently identified lake benches at 1740 m and 1720 m south of the Katar River, but I have not visited these in the field. He believes that they are recessionals of an 1850 m lake in which the Gademotta gravels (0.18 – 0.15 m.y.) and the subaqueous tuffs in the Shala caldera wall (0.24 – 0.20 m.y.) were deposited. As yet the evidence for this ancient lake is too fragmentary to be evaluated.

Around the base of Gademotta, stretching north to the Meki River, there are outcrops of lower, and presumably more recent, lacustrine sands and beach gravels on which brown calcimorphic soils (F.A.O. orthic luvisols) have developed (Plate 37). A pre-Holocene age for these deposits is suggested on the following grounds:

1) Their lack of strandline morphology.
2) Their height above the Holocene outlet.
3) Their degree of soil development.

b. Soil stratigraphy

Pedologic evidence is central to the identification of Pleistocene deposits. The key area is West Ziway (Figs. 5.9, 5.10), where soils developed on the dated Holocene terraces I to IV of Laury and Albritton (1975) can be compared with those on the adjacent terraces V and VI, outside the limits of the Holocene lake.

Surface soils on well-drained Holocene lake deposits, which have only had about 5000 years to form, are skeletal or weakly developed, contrasting markedly with the much more strongly differentiated soils on Pleistocene sediments. The following soil profile was recorded in an auger hole on terrace IV north of Rama Gebriel (Fig. 5.9). It can be regarded as representative of pumiceous Holocene lake deposits (see Makin et al., 1976). The horizon designations and classification follow the new F.A.O. system (1974).

Soil Profile 1, Vitric andosol (Tv), (site 0838-D12)

This profile is located ½ km east of the West Ziway faultscarp, on level ground at about 1665 m under *Acacia* scrub and grass. It is well drained.

A	0–20 cm	Dark brown (10YR 3/3), slightly gritty loam; medium subrounded blocky; slightly plastic and sticky wet; noncalcareous, pH 8.5.
Bw1	20–60 cm	Dark brown (10YR 3/3), sandy clay loam; crumb structure; slightly sticky and moderately plastic wet; noncalcareous, pH 8.5.
Bw2	60–100 cm	Dark brown (10YR 3/3), sandy loam; crumb structure; non-sticky and slightly plastic wet; noncalcareous, pH 8.5.
Ck1	100–120 cm	Olive brown (2.5Y 4/3), loamy sand; crumb structure; nonsticky and nonplastic wet; calcareous, pH 8.5.
Ck2	120–180 cm	Dark greyish brown (2.5Y 4/2), gritty sand; crumb structure; nonsticky and nonplastic wet; calcareous, pH 8.5.

In this soil, B horizon formation is very slight. Carbonate is only present below 100 cm in disseminated form. Probably the parent material (bedded pumice sand) was calcareous, but leaching of $CaCO_3$ is indicated by a slight maximum from 100 to 140 cm. All horizon boundaries are gradational, and the influence of parent material texture is very marked.

Compared the above with the following description of a soil of the pre-Holocene terrace V, just west of Ziway Town:

Soil Profile 2, Orthic luvisol (Lo)[1], site 0738-B1

This profile is exposed in a small gully cutting into terrace V. It formed on well drained, level ground, at about 1688 m. The present vegetation

1. This soil is classified as an orthic luvisol rather than a calcic luvisol because the calcic horizon lies below 1.25 m.

is *Acacia* scrub. LSA and MSA(?) flakes and cores were found on the slopes of the gully, with MSA artifacts *in situ* in the soil. All horizon boundaries are gradational.

Au1	0-30 cm	Black (7.5YR 2/1), slightly gritty silty clay; strong, fine to medium, subangular blocky; non-sticky to slightly sticky and moderately plastic wet; moderately abundant roots; moderately calcareous, with carbonate films along root channels, pH 8.5.
Au2	30-65 cm	Black (7.5YR 2/1), slightly gritty silty clay; weak to moderate, subrounded blocky; moderately sticky and moderately plastic wet; moderately calcareous, with carbonate films along root channels, pH 8.5.
Bu1	65-115 cm	Brown (7.5 YR 4/3), slightly gritty silty clay; strong, fine to medium, subrounded blocky; moderately sticky and moderately plastic wet; noncalcareous, pH 8.5.
Bt1	115-190 cm	Dark brown (7.5YR 4/4), clay; strong, medium subangular blocky; moderately sticky and plastic wet; dark brown cutans strongly developed on ped faces and root channels; noncalcareous, pH 8.0.
Bt2	190-240 cm	Yellowish brown (10YR 5/4), heavy silt loam; strong, medium to coarse subrounded blocky; slightly sticky and slightly to moderately plastic wet; dark brown cutans moderately developed on ped faces and root channels; noncalcareous, pH 8.5.
Bw	240-290 cm	Dark brown (7.5 YR 4/4) loam; weak, subrounded blocky; slightly sticky and slightly to moderately plastic wet; noncalcareous, pH 8.5.
Cck1	290-330 cm	Dark brown (7.5YR 4/4) silt loam; moderate sub-rounded blocky; nonstocky, slightly to moderately plastic wet; strongly calcareous, with 1-2% 1-3 cm $CaCO_3$ nodules and carbonate films along veins and root channels, pH 8.5.
Cck2	330-370 cm	Dark yellowish brown (10YR 4/4) silt loam; weak, subrounded blocky; nonsticky, slightly to moderately plastic; strongly calcareous, with 10-50% large branching $CaCO_3$ concretions, some smaller (0.5 cm) nodules and carbonate films along root channels, pH 8.5.

This is a typical brown calcimorphic soil with a well-differentiated textural and structural argillic (Bt) horizon and a calcrete layer below 3.3 m. Similar soils outcrop widely along terrace V, although here they have often suffered severe surface erosion; and on both sides of the overflow channel north of the Meki (Figs. 5.3 and 5.9; Plate 5). At site 0838-D7, for example, a truncated calcrete profile 1.50 m thick is preserved on the tread of terrace V at the top of the West Ziway fault scarp (Plates 37-38); in sharp contrast with profile 1 only 500 m to the east (0838-D12). This big change in soil type across the IV/V terrace boundary conflicts with the view put forward by Gèze (1974) that the fault is a Holocene feature.

It is probably unwise to attempt to differentiate the Pleistocene deposits any further using soil evidence alone. Laury and Albritton (1975, pp.1009-1010) have argued convincingly that brown calcimorphic soils are the typical end-product of soil-forming processes acting on volcaniclastic sediments during moister phases in the history of the Rift, when the slopes were stabilized by vegetation. The Gademotta Formation contains at least eight such stacked-up palaeosols (Fig. 3.1), the youngest of which are morphologically very similar to the most mature soils on terraces V and VI, although they contain MSA rather than LSA artifacts. Furthermore, even the surface brown soils are often polygenetic. At site 0838-D1, for example, two argillic horizons are separated by a volcanic ash. The surface soil contains LSA blades and scrapers whereas the buried soil yields MSA artifacts, indicating that it is probably older than 15,000 years (Clark and Williams, in press). Two distinct maxima of illuviated clay are also present at 0838-D8 (Plate 5).

The youngest brown calcimorphic soil characteristically contains LSA artifacts. It has been dated by Southern Methodist University at the Wayso archaeological site (Fig. 5.9, locality 3). Samples of charcoal found in stratigraphic context with LSA artifacts provide a maximum age for the soil, which is developed in pumice and volcanic ash (the Abernosa pumice member?). The results obtained were 11,510 $\pm$ 110 BP (SMU-72) and 10,333 $\pm$ 90 BP (SMU-86) (Haynes and Haas, 1974; Laury and Albritton, 1975). This indicates that the latest phase of brown calcimorphic soil formation corresponded to the early-mid Holocene lacustral phase (chapter 8).

c. Late Pleistocene (?) lake levels.

In this time-bracket we can include both the lake deposits underlying terrace V, and the calcreted pre-Holocene gravels at the southern end of the overflow channel (section B1a); on the grounds that they are intermediate in age and elevation between the Gademotta Formation and the Holocene terraces.

The terrace V deposits north of Gademotta are essentially shallow-water sands with occasional diatomite lenses (Fig. 5.15). The carbonate present is organized into calcrete horizons or large cannon-ball concretions. No mollusca are preserved. The diatomite in section 0838-D7 contains a littoral, brackish-water diatom flora (Appendix 4), which implies a closed and fluctuating lake. Ripple-marked sands are present at 0838-D6, D7 and D10 (Plate 38), indicating proximity to the shore, which must however have lain to the west of the present West Ziway fault.

Tectonic displacements make it nigh on impossible to reconstruct the contemporary palaeogeography. Along the West Ziway fault, the outer edge of terrace V ranges in height from 1682 m at 0838-D5 to 1694 m at 0838-D7, where the throw increases to at least 18 m. Although the relationship of this lake to the Meki-Dubeta col is obscure, the diatom flora tends to rule out the idea of an outflow.

South of Gademotta, the surface of terrace V stands at 1680 m (site 0738-B1), and is underlain by rounded rhyolite beach gravel which may have been reworked from the Gademotta Formation. This shoreline is probably younger, and may correspond to the pre-Holocene outlet level ($\geqslant$ 1673 m) discussed in section B1a).

The preliminary model of lake-level highstands derived from shoreline studies is refined in the next chapter in the light of stratigraphic evidence.

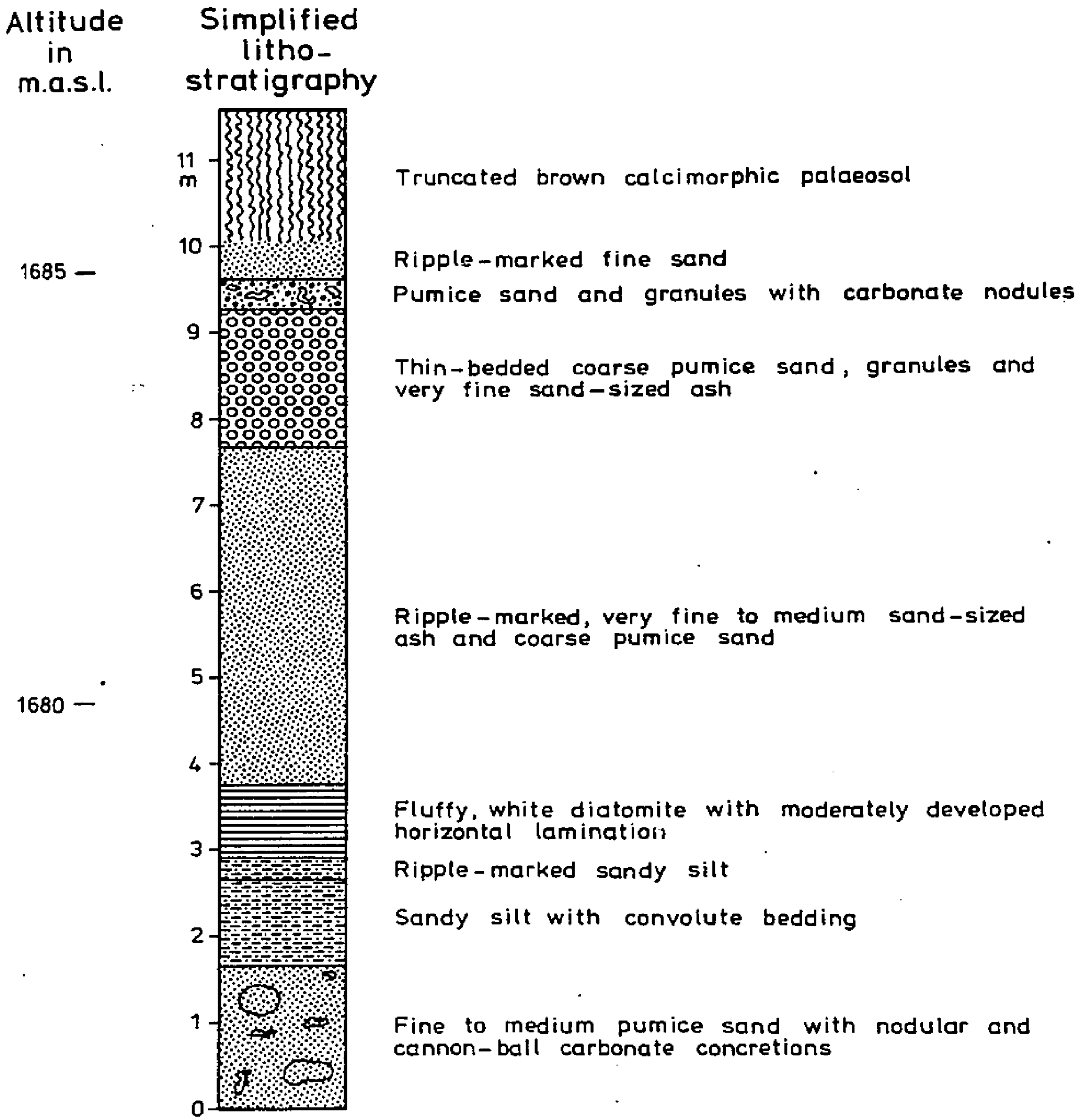


Fig. 5.15: Stratigraphy of section 0838-D7, West Ziway fault scarp (Plates 37-38).

CHAPTER SIX

Stratigraphic Evidence for Former Lake Levels

6A SUMMARY

This chapter is an account of the Late Quaternary sedimentary stratigraphy of three principal areas:

1) The Bulbula Plain
2) The Shala Basin
3) The Meki Valley.

Within each area the sequence of lacustrine and nonlacustrine rock units has been established in a type locality or section. Three formations have been defined, as follows:

1) The Bulbula formation

The type section, in the Deka Wede valley, spans approximately the last 30,000 years. It has furnished a total of ten ^{14}C dates. Another important reference section is the Abiyata Core, with five ^{14}C dates. At the type locality, the lacustrine deposits are diatomaceous marls and calcareous sands; grading laterally into thick diatomites which outcrop in the Kurkura and Kertefa areas. The unifying feature is the presence of the nonlacustrine Abernosa pumice member, a thick accumulation of air-fall pyroclastics of terminal Pleistocene age, throughout the area. The lacustrine members have been correlated on the basis of relative position, thickness, diatom assemblages and radiocarbon age.

2) The Ajewa formation

This formation has been named after a type area in the northeast embayment of Lake Shala. The 3.6 km-long Ajewa section spans the complete height range of the palaeolake. It has yielded ten ^{14}C dates. The oldest lacustrine beds are lower Late Pleistocene or Middle Pleistocene in age. The stratigraphy in this area is very complex and not all the problems of

interpretation have been adequately resolved.

3) The Meki formation

The Meki formation consists largely of alluvial and fluviolacustrine deposits. It has not yet been dated. The sequence records the effect of fluctuations in base level on the Meki River. During the 1670 m highstands the river aggraded its valley to the level of the outlet.

Lateral correlations between these three formations (Table 6.9) allow us to interpret the distribution of sedimentary environments through time (chapter 7). They also help to clarify the sequence of lake-level fluctuations derived from geomorphic investigations (chapter 8).

6B INTRODUCTION

The previous chapter presented a simple model of the temporal variation in lake levels, based largely on geomorphic evidence. This requires substantial modification in the light of stratigraphic information and new radiocarbon dates. For example, it is now clear that there were several Late Pleistocene highstands, and that the lake reached the 1670 m level at least twice.

The unconsolidated lacustrine sediments and interbedded tuffs which outcrop on the present Rift floor were designated the Galla Group by Lloyd (United Nations, 1973). The marked changes in facies within these beds mean that correlations between different parts of the basin have to be based on radiocarbon dating and diatom stratigraphy rather than on lithologic criteria. The Galla Group is therefore informally divided here into several mappable formations with partly overlapping time-ranges. These are listed in Table 6.1.

The three newly named formations have been further subdivided, but because of the lateral variability in lithology only one widely traceable unit, the Abernosa pumice member of the Bulbula formation, has been individually named. Otherwise the most important units have been numbered sequentially within key sections (e.g. Deka Wede 1-4, Kurkura 1-6).

Table 6.1: Sedimentary Formations within the Galla Group

Formation	Type section/locality	Character	Named Units
BULBULA	Bulbula Plain Type section: Deka Wede valley (section 0738-B20, profiles 10C and 11; DU 659611)	Lacustrine, alluvial, colluvial and pyroclastic sediments	Abernosa pumice member (Lacustrine members numbered in key sections)
AJEWA	Northeast Shala Type locality: Ajewa district (section 0738-D2; DU 610265)	Lacustrine, colluvial and alluvial sediments	Roricha fish bed (all laterally traceable units in type locality numbered)
MEKI	Meki Valley Type locality: Entrance of overflow channel (Italconsult boreholes 1 and 5; DV 748056)	Alluvial, delta-plain, colluvial and lacustrine sediments	None (Units numbered in type locality)
GADEMOTTA	Gademotta Ridge Type locality: Eastward-facing slope (archaeological sites ETH 72-5 to 72-8) Reference: Laury and Albritton (1975)	Colluvial, laharic and pyroclastic sediments with palaeosols	None (Units numbered)

The exposures of lake beds in the Bulbula Plain are the best dated and most complete. The type section of the Bulbula formation, 0738-B20 (Gasse and Street, 1978a), provides a detailed sequence of lake-level fluctuations which sheds light on the more equivocal stratigraphy in the Shala Basin (chapter 6D). The main advantage of the northeast Shala type area is that the exposures of the Ajewa formation span the complete height range of the palaeolake, and cover a substantially longer time-period than 0738-B20. Other interesting, but poorly dated, deposits in the Meki Valley (the Meki formation) reveal the way in which the inflowing rivers responded to changes in base level.

In this chapter each of these three formations is described in detail. The lacustrine members are then correlated, using the available ^{14}C dates and the diatom biostratigraphy established by Gasse and Descourtieux (Gèze, 1975; Descourtieux, 1977; Gasse and Descourtieux, in press), in order to build up a detailed sequence of lake-level fluctuations. The lacustral phases (highstands) inferred in this way are, strictly speaking, time-transgressive and should therefore be regarded geologic-climate units[1] in the same way as glacials and interglacials (Morrison, 1966). Following the convention established by Gasse (1975, 1977a), they are given Roman numerals in order to distinguish them from rock-stratigraphic units (Arabic numbers) and diatom biozones (letters). The Bulbula, Abiyata and Ajewa sequences have been numbered separately because none of the sections contains a complete record.

6C THE BULBULA FORMATION

I. Introduction

A great thickness of Late Quaternary sediments underlies the Bulbula Plain, and has been exposed by the Bulbula River and its tributaries as they cut down to keep pace with the falling level of Lake Abiyata. These sediments are informally referred to as the Bulbula formation. Outcrops

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1. American Commission on Stratigraphic Nomenclature (1961, art. 39).

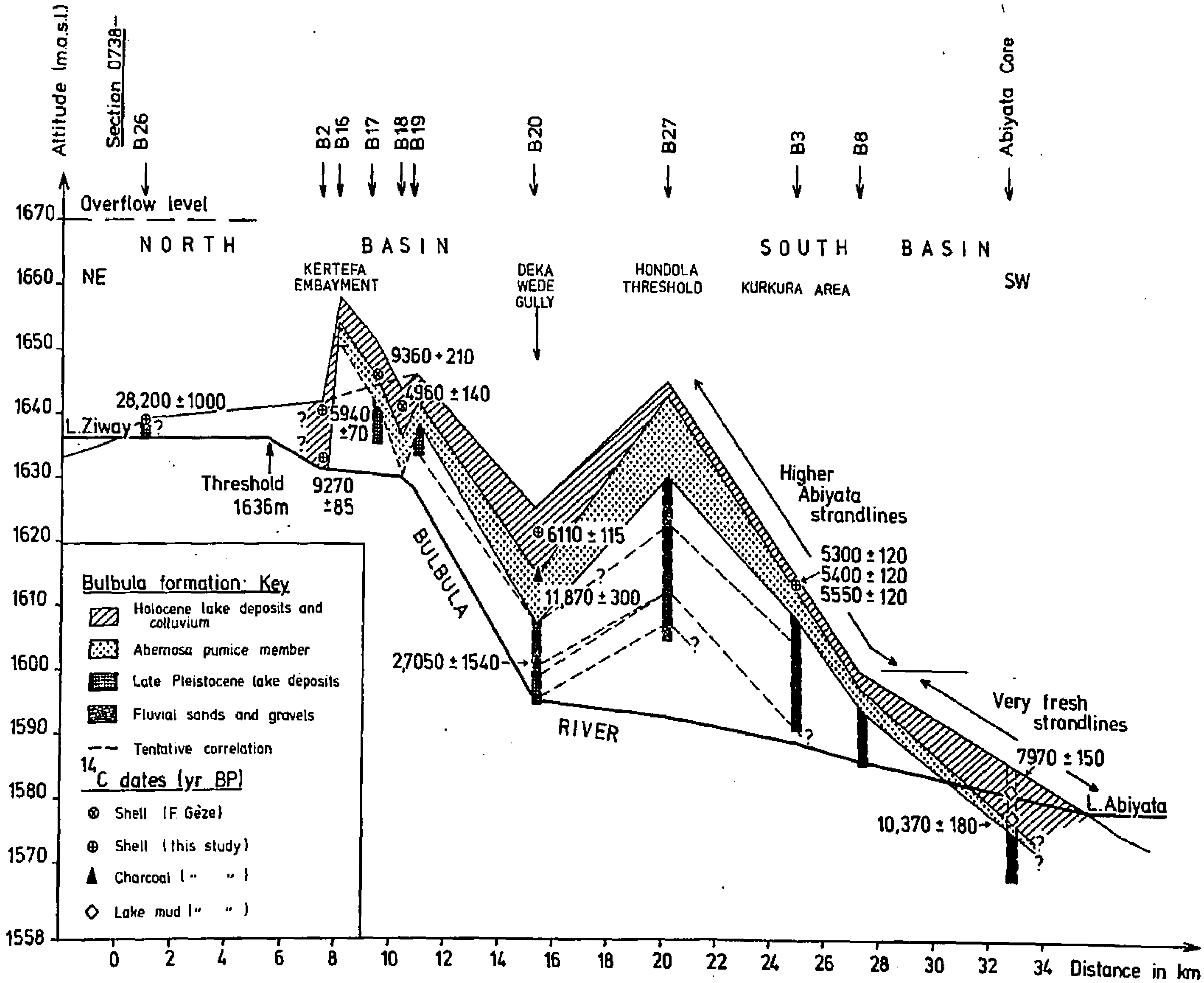
are most numerous where ephemeral gullies enter the gorge from Mt. Alutu (Fig. 5.5, Plate 39).

Figure 6.1 summarizes the stratigraphy of the Bulbula Gorge. Since the river runs closer to the mountain-foot at B16 than at B27, the exposures accordingly reach higher elevations there. But the apparent dip of the sediments to N and S away from the Hondola threshold is real, and results from the geological structure of the plain.

The western piedmont of Alutu is a great fan of rhyolitic pyroclastics and alluvium which intertongues westwards and southwards with lacustrine deposits. As far as one can judge, the volcaniclastic sediments are thickest to the WSW of the mountain, around B27 (Figs. 5.5 and 6.1). Their accumulation probably started before deposition of the oldest exposed lake sediments, which also rise in elevation southwards up to the Hondola threshold and then slope down towards the South Basin. These are primary dips, and not due to uplift of the Hondola sill, as can be seen from the configuration of the terminal Pleistocene and Holocene deposits. During the terminal Pleistocene a fan of pyroclastics was laid down on the dried-up lake floor to leeward (SW) of Alutu. This will be referred to as the Abernosa pumice member, after the large Government ranch of that name. It reaches a maximum thickness of 13 m at B27, dwindling to 4.5 m at B17 and 3.8 m at B3. In its later stages, the fan was gullied by the drainage from Alutu. Overlying it are younger lake beds, draped over a highly irregular topography.

All the lacustrine deposits in the Bulbula formation contain a large component of volcanic glass, reworked from the Alutu pyroclastics. Their other principal constituent is diatomite. High-carbonate sediments occur locally in the Deka Wede embayment.

Fig. 6.1: Schematic cross section along the Bulbula River showing major stratigraphic units. Only key ^{14}C dates are shown.



Altitude (m.a.s.l.)
Section 0738
B26
B2
B16
B17
B18
B19
B20
B27
B3
B8
Abiyata Core
Overflow level
NORTH BASIN
SOUTH BASIN
NE
SW
KERTEFA EMBAYMENT
DEKA WEDE GULLY
HONDOLA THRESHOLD
KURKURA AREA
9360 ± 210
4960 ± 140
28,200 ± 1000
L. Ziway
5940 ±70
Threshold 1636m
9270 ±85
BULBULA RIVER
6110 ± 115
11,870 ± 300
27050 ± 1540
Higher Abiyata strandlines
5300 ± 120
5400 ± 120
5550 ± 120
Very fresh strandlines
7970 ± 150
L. Abiyata
10,370 ± 180
Bulbula formation: Key
Holocene lake deposits and colluvium
Abernosa pumice member
Late Pleistocene lake deposits
Fluvial sands and gravels
Tentative correlation
14C dates (yr BP)
Shell (F. Gèze)
Shell (this study)
Charcoal (" ")
Lake mud (" ")
1670
1660
1650
1640
1630
1620
1610
1600
1590
1580
1570
1558
0
2
4
6
8
10
12
14
16
18
20
22
24
26
28
30
32
34
Distance in km

II. The Deka Wede marls (section 0738-B20)

a. The stratotype of the Bulbula formation: profiles 10C and 11

i. General setting

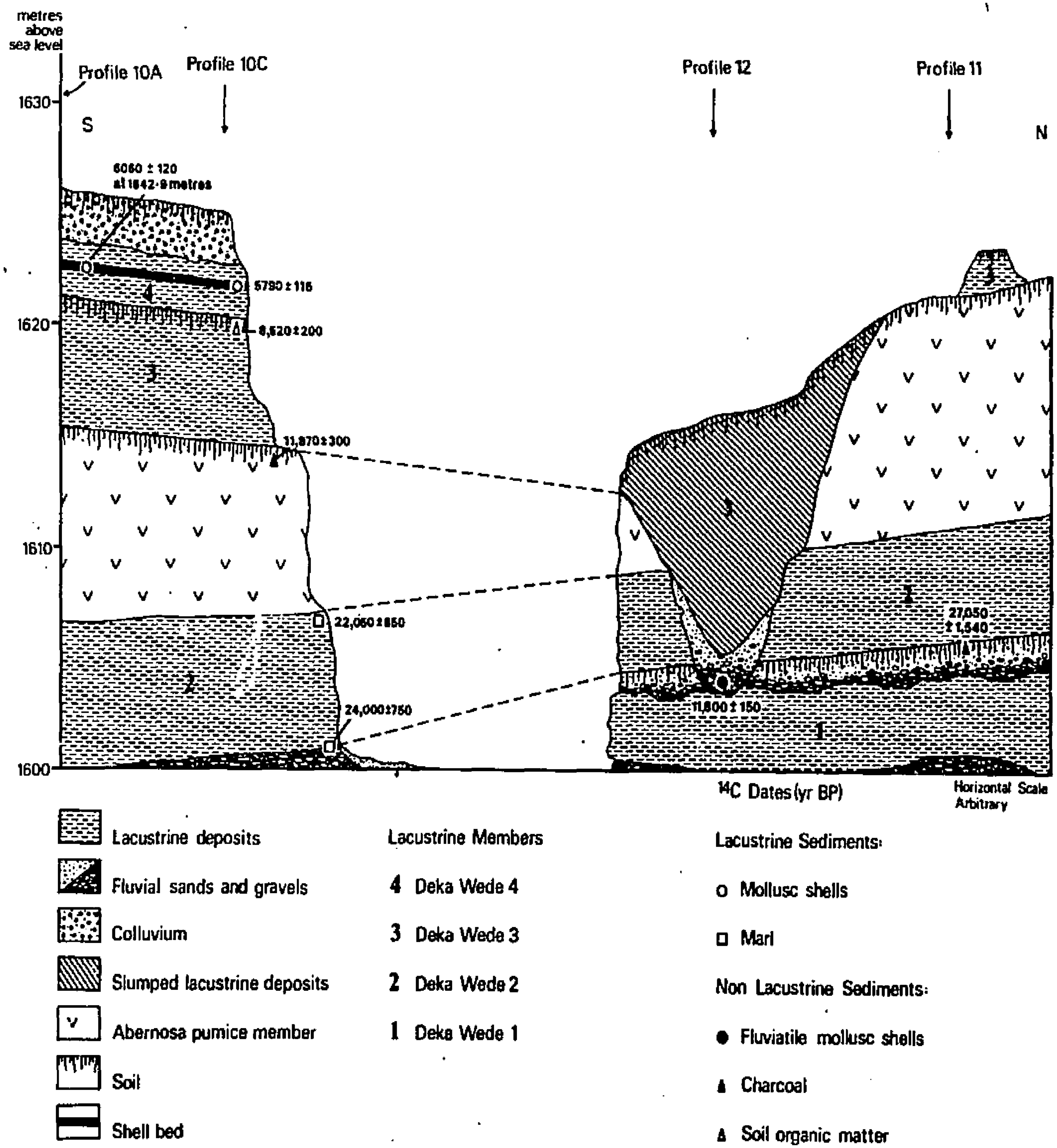
The type section was selected in the largest of the modern Bulbula tributaries, the Deka Wede, at profiles 10C and 11 (Fig. 5.5). Because lateral shifting of the drainage from Alutu is inhibited by lava flows from the mountain, repeated incision has taken place along the line of the present stream during episodes of low lake level. Each time the lake rose, its deposits were draped over the sides of the flooded gully system, with effects which can be clearly seen in Fig. 6.2. On reaching the 1670 m level the shoreline became much straighter (Fig. 5.5).

Increasingly older sediments are exposed as the Deka Wede deepens towards the Bulbula. The type section was chosen where the sequence is most complete and least affected by post-depositional slumping (Figs. 6.2, 6.3). A total thickness of 30.0 m of sediments is exposed between 1600 and 1625 m (42 - 67 m.S.D.). The lowest deposits dip southwards, so that the basal 5.7 m are only exposed in the north wall of the gully (profile 11; Fig. 6.2, Plate 40).

Four principal lacustrine members have been defined at this locality (Deka Wede 1-4). Each represents a major lacustral phase, numbered Bulbula I-IV (Gasse and Street, 1978a; Street, in press, a). These members have been subdivided using the evidence of minor disconformities, coupled with the grain size and carbonate curves (Fig. 6.3). Despite locally deep water in the flooded embayment, the sedimentary facies and diatom floras always betray the proximity of the shore.

Interbedded with the lacustrine sediments there are pyroclastics, stream gravels, colluvium and soils. Two of the palaeosols have yielded artifacts and dispersed charcoal in stratigraphic context (Plate 41). The valley was evidently a favoured site for Palaeolithic encampments during lowstands of the lake. Interpretation of the nonlacustrine sediments has been greatly aided by

Fig. 6.2: Schematic cross section of the Deka Wede gully at the Bulbula formation type site. (Plates 40-41, 57).



metres above sea level
1630
1620
1610
1600
Profile 10A
Profile 10C
Profile 12
Profile 11
S
N
6060 ± 120 at 1642·9 metres
5780 ± 116
8,620 ± 200
11,870 ± 300
22,060 ± 850
24,000 ± 750
27,050 ± 1,540
11,800 ± 150
4
3
2
1
^{14}C Dates (yr BP)
Horizontal Scale Arbitrary
Lacustrine deposits
Fluvial sands and gravels
Colluvium
Slumped lacustrine deposits
Abernosa pumice member
Soil
Shell bed
Lacustrine Members
4 Deka Wede 4
3 Deka Wede 3
2 Deka Wede 2
1 Deka Wede 1
Lacustrine Sediments:
o Mollusc shells
□ Marl
Non Lacustrine Sediments:
● Fluviatile mollusc shells
▲ Charcoal
Δ Soil organic matter

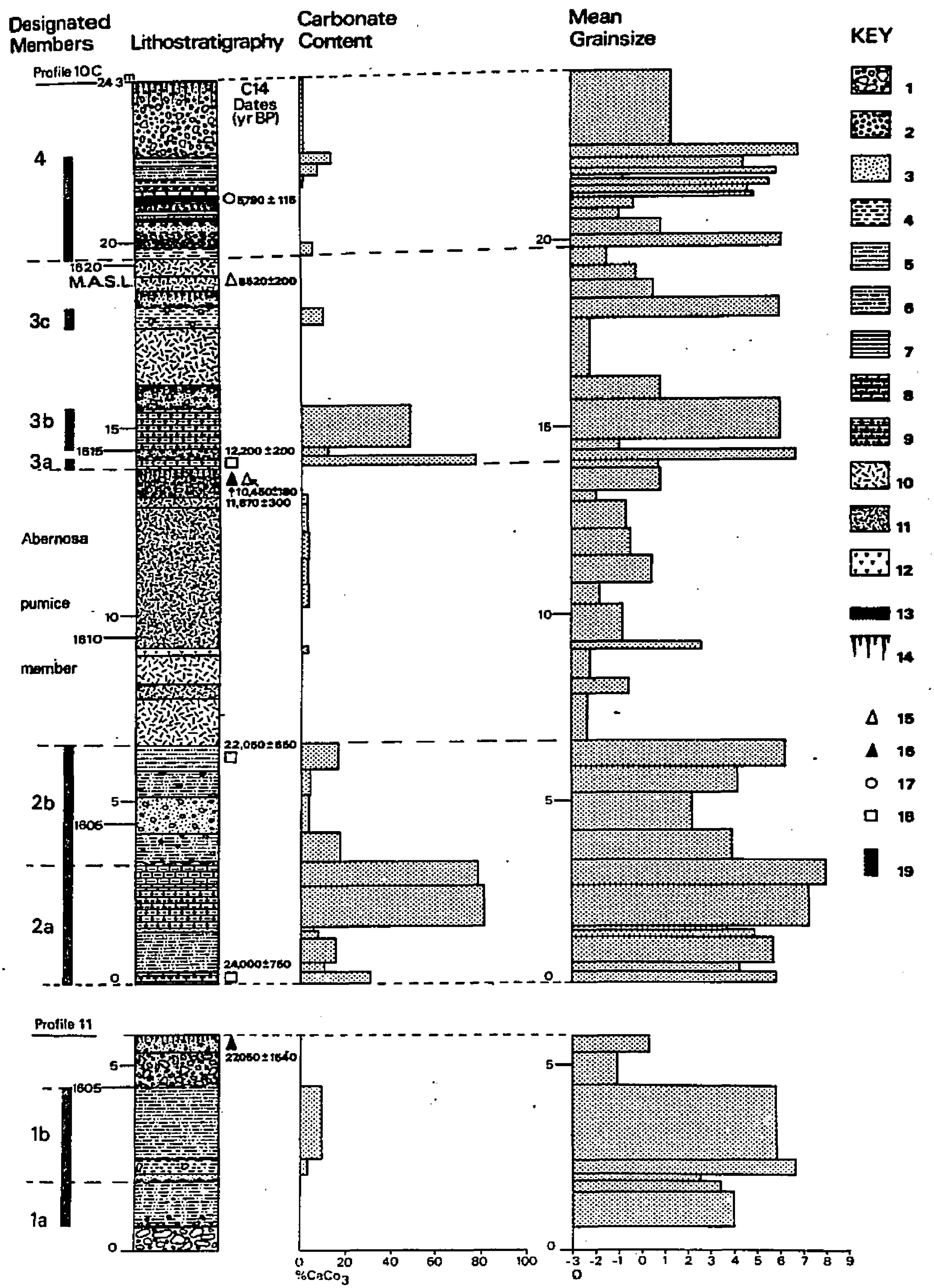
Fig. 6.3: Stratigraphy of the type section of the Bulbula formation (0738-B20/10C-11).

Key

1. Pebbles and cobbles
2. Granules
3. Sand
4. Silt
5. Clayey silt
6. Silty sand
7. Diatomite
8. Clayey marl
9. Silty marl
10. Primary pumice deposit
11. Primary pumice deposit (partly weathered)
12. Primary ash deposit
13. Shell bed
14. Soil

^{14}C dates:

15. Soil organic matter
16. Charcoal
17. Shell
18. Marl
19. Lacustrine units



Designated Members
Lithostratigraphy
Carbonate Content
Mean Grainsize
KEY
Profile 10C
24.3m
C14 Dates (yr BP)
4
5790 ± 115
20
1620 M.A.S.L.
8620±200
3c
3b
15
1615
12,200 ±200
3a
10,450±180
11,870±300
Abernosa
pumice
10
1610
member
22,050±650
2b
5
1606
2a
24,000±750
0
Profile 11
22050±1640
1605
1b
1a
%CaCo3
0 20 40 60 80 100
-3 -2 -1 0 1 2 3 4 5 6 7 8 9
1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19

discussions in the field with M.A.J. Williams, E.F. Lloyd and R. Gillespie.

The dating framework consists of ten radiocarbon ages on shell, marl, charcoal and soil humus (Fig. 6.3 and Appendix 2b). These are stratigraphically consistent except for a knot of dates at the base of member Deka Wede 3a. The oldest sediments are > 27,050 ^{14}C yr. old. In view of the continuity of the sequence in profile 10C, and the reasonable relationship between the lowest charcoal date and the two overlying marl dates, the broad validity of the ^{14}C chronology will be accepted. This is summarized in Table 6.2.

ii. Lithofacies

Identification of lacustrine facies was based on the following criteria: the presence of lacustrine organisms, pronounced sorting by grain size and density, light colour, and a characteristic association of structures such as fine lamination and small-wave ripples (Table 7.1). The lake beds at this site consist essentially of silt-sized calcite and diatom frustules which originated within the lake, diluted by pumice and volcanic ash brought down from the slopes of Alutu. Mollusc shells, ostracods and fish bones also occur in some beds. The origin of shell concentrations is discussed further in chapter 7DIIId.

Figure 6.3 shows that the mean grain-size curve for the noncarbonate fraction of the lake sediments varies in a systematic fashion, which is believed to reflect the water depth at the time of deposition (chapter 7DIb). It ranges from -0.89 Ø (very coarse sand) to +8.51 Ø (coarse clay) (Fig. 6.4 and Appendix 7). The coarsest samples are very rich in pumice, although grains of quartz, feldspar and pyroxene are also visible in thin section. Granulometrically, they are very similar to modern beach sediments (Fig. 2.12d), whereas the finest samples overlap the field occupied by the bottom muds of Lake Shala (Fig. 2.12e). No layer silicates were identified in the clay-sized material, which is largely amorphous to X-rays, apart from traces of quartz.

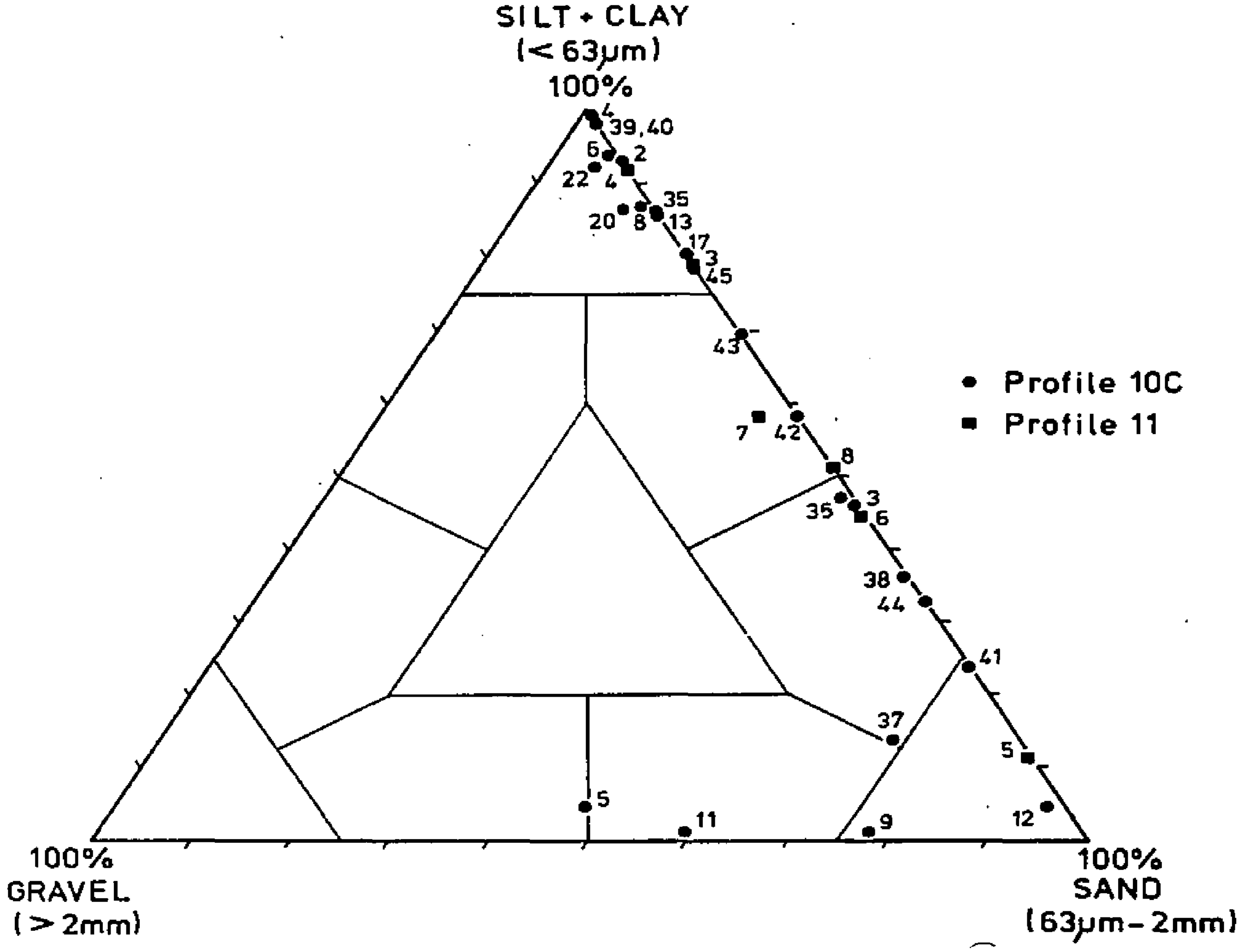
As indicated by Figure 6.3, the total carbonate content of the lake sediments (excluding mollusca) increases rapidly with decreasing grain size

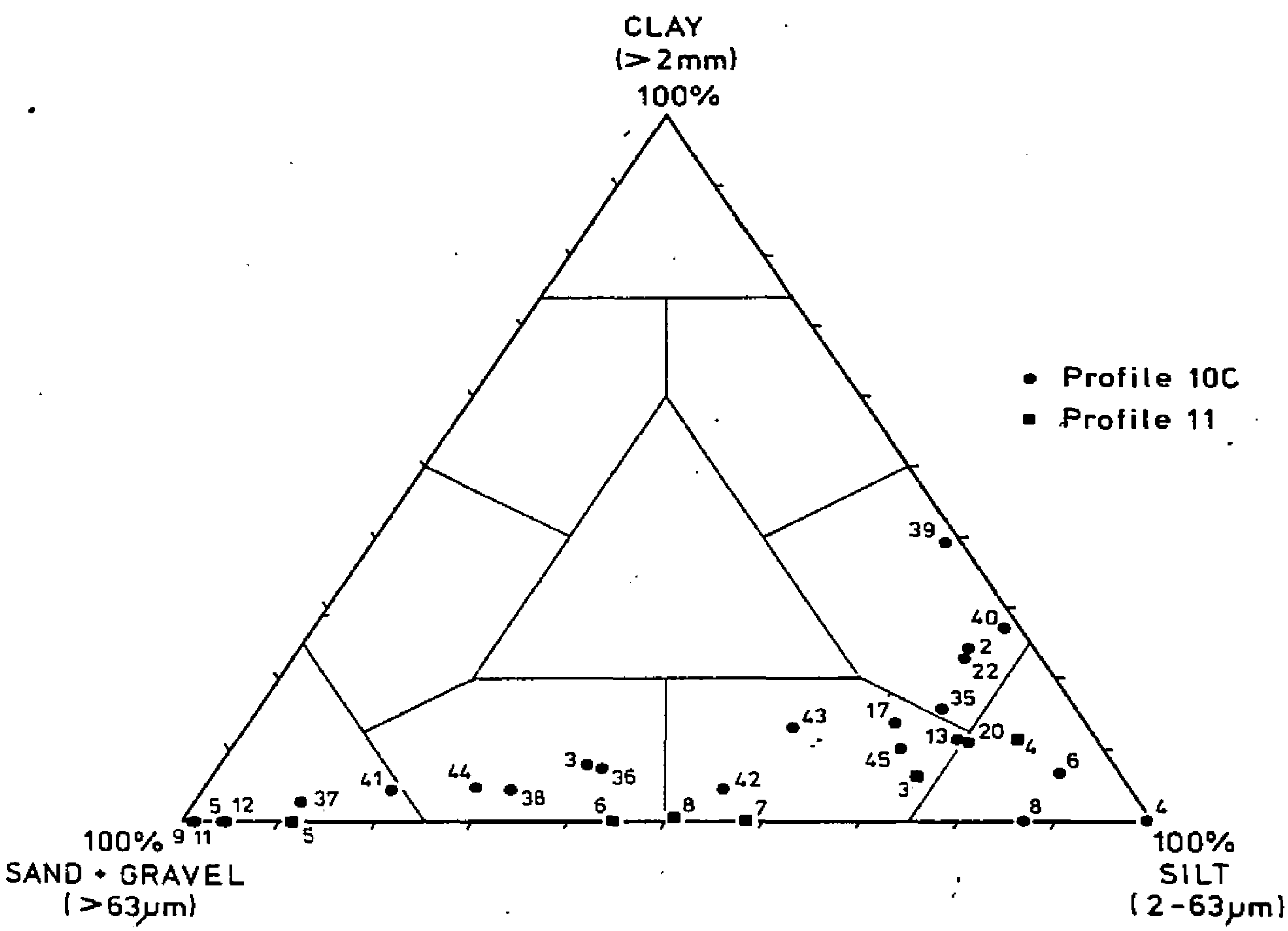
Table 6.2: Stratigraphy of the type section of the Bulbula formation (0738-B20/10C-11)

Simplified lithostratigraphy	Designated members	Inferred lake-level fluctuations	Lacustral phases	^{14}C dates[1] (yr BP)
—— Surface soil —— (Truncated A/C profile developed in brown (10YR 5/3) colluvium)		Recession		
Shelly pumice gravels, waterlaid ash and light grey (2.5Y 6/1) to white (10YR 8/2) calcareous silts	Deka Wede 4	Highstand > 1643 m	BULBULA IV	O 5790 ± 115 (O 6060 ± 120)[2]
—— Buried soil —— (Dark greyish-brown (10YR 4/2), A horizon developed in air-fall pumice)		Recession < 1619 m		▲ 8520 ± 200
Pale brown (10YR 6/3), calcareous silt with scattered shells and ostracods	Deka Wede 3c	Highstand > 1619 m	BULBULA IIIC	
1.55 m air-fall pumice lapilli —— Buried soil —— (Greyish-brown (10YR 5/2), A horizon developed in fluvial sand)		Brief recession < 1616 m		
Light grey (10YR 7/2), diatomaceous marl with scattered shells	Deka Wede 3b	Highstand > 1617 m	BULBULA IIIB	
Light grey (10YR 7/1), fluvial gravel with LSA implements		Brief recession < 1615 m		
Light grey (10YR 7/2), diatomaceous marl	Deka Wede 3a	Highstand > 1615 m	BULBULA IIIA	O 12,200 ± 200
—— Buried soil —— (Greyish-brown (10YR 5/2), A horizon developed in sandy colluvium. Abundant LSA implements, bone refuse and charcoal) Light grey (10YR 6/1 to 7/2), air-fall pumice and ash beds with pale brown (10YR 6/3) weathering horizons	Abernosa pumice member	Recession < 1603 m		Δ 10,450 ± 180 (O 11,600 ± 150)[2] ▲ 11,870 ± 300 "TERMINAL PLEISTOCENE"
Pale yellow (2.5 Y 7/3), calcareous silt overlying pebbly pumice sand Very pale brown (10YR 7/3 to 8/3), diatomaceous marls overlying fine pumice sand and marl	Deka Wede 2b Deka Wede 2a	Highstand > 1607 m Brief recession? Highstand > 1604 m	BULBULA IIB BULBULA IIA	O 22,050 ± 650 O 24,000 ± 750
—— Buried soil —— (Light brownish-grey (2.5 Y 6/2) A horizon developed in fluvial sand and gravel. Abundant obsidian implements, bone refuse and charcoal)		Recession < 1604 m		▲ 27,050 ± 1540
Light grey (2.5Y 7/1), ostracod-bearing, very fine sand Light grey (2.5Y 7/1 to 7/2), pumice sand and silt with scattered shells	Deka Wede 1b Deka Wede 1a	Highstand > 1605 m Brief recession? Highstand > 1602 m	BULBULA IB BULBULA IA	UPPER LATE PLEISTOCENE?
Fluvial cobble- and boulder-gravel with large clasts of welded tuff and obsidian		Lake low < 1600 m		

1. Symbols as for Fig. 6.3.
2. Date interpolated from another profile.

Fig. 6.4: Grain-size distributions of lacustrine sediments from 0738-B20 (Appendix 7a).


SILT + CLAY
(< 63μm)
100%
100%
GRAVEL
(> 2mm)
100%
SAND
(63μm-2mm)
Profile 10C
Profile 11


CLAY
(> 2mm)
100%
100%
SAND + GRAVEL
(>63μm)
100%
SILT
(2-63μm)
Profile 10C
Profile 11

below a mean diameter of about 4 Ø, reaching a maximum of almost 85% at a grain size of 7 – 9 Ø (2 – 8 μm). This suggests that the microcrystalline calcite grains average about 2 – 8 μm in diameter, and have therefore been concentrated at a depth in which wave action was just insufficient to move fine, silt-sized particles.

Examined in thin section, the matrix of the high-carbonate marls is seen to consist of microcrystalline grains of low-Mg calcite (micrite)[1] with 2-5 mole % $MgCO_3$ (Plates 46, 47). No aggregates of algal or other origin appear to be present, although void fillings of coarser (sparry) calcite betray slight weathering of the top of units 2b and 4 (Plates 48, 53). Diatoms are very abundant in some samples, especially in members 3a and 3b, and are responsible for the fine laminations visible in member 2a (Plates 45, 49–51). Ostracod tests (low-Mg calcite) account for only a small proportion of the total carbonate, except in members 1b and 3c (Plate 44). In carbonate-poor samples, such as unit 1a, the majority of the matrix consists of finely ground volcanic glass which appears dark under crossed polars (isotropic) (Plates 42–43). The origin of the marls is discussed in more detail in chapter 7DIIIa.

The nonlacustrine sediments are of four main types. These can be readily differentiated in the field by colour, texture and lithology. Some laboratory data are given in Figure 6.5 and Appendix 7, for comparison with the modern samples shown in Fig. 2.12. These four types are as follows:

1) Coarse gravel containing large cobbles and boulders of resistant rocks derived from Alutu, such as rhyolite, obsidian and welded tuff.

2) Unimodal, but poorly to very poorly sorted, light grey to brown gravels and sands with a wide mixture of pumice and harder lithologies. Pumice pebbles are subangular to subrounded. These resemble present-day gully-floor gravels (Fig. 2.12b) and frequently fill channels cut into older deposits. They may contain clasts of lake marl.

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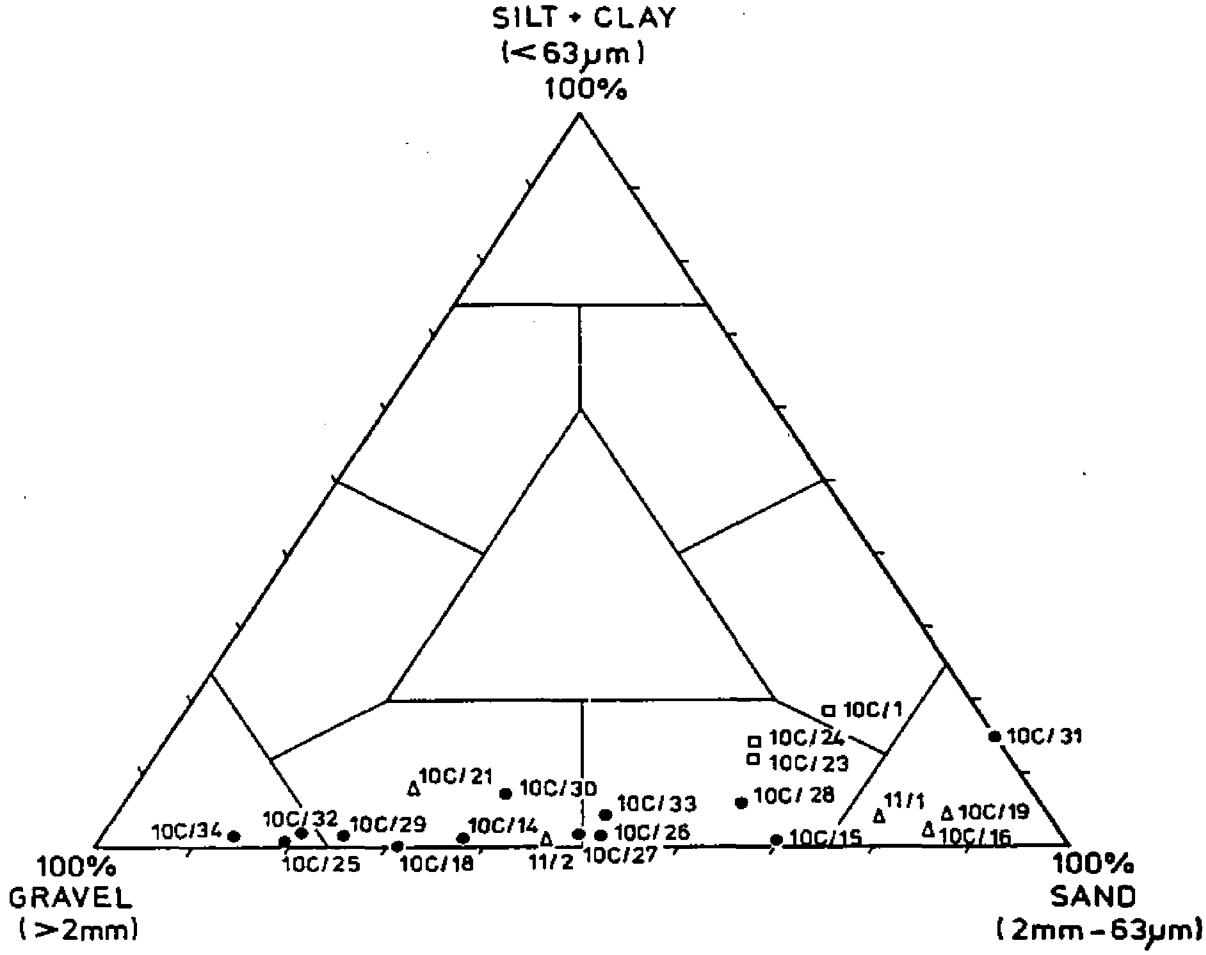
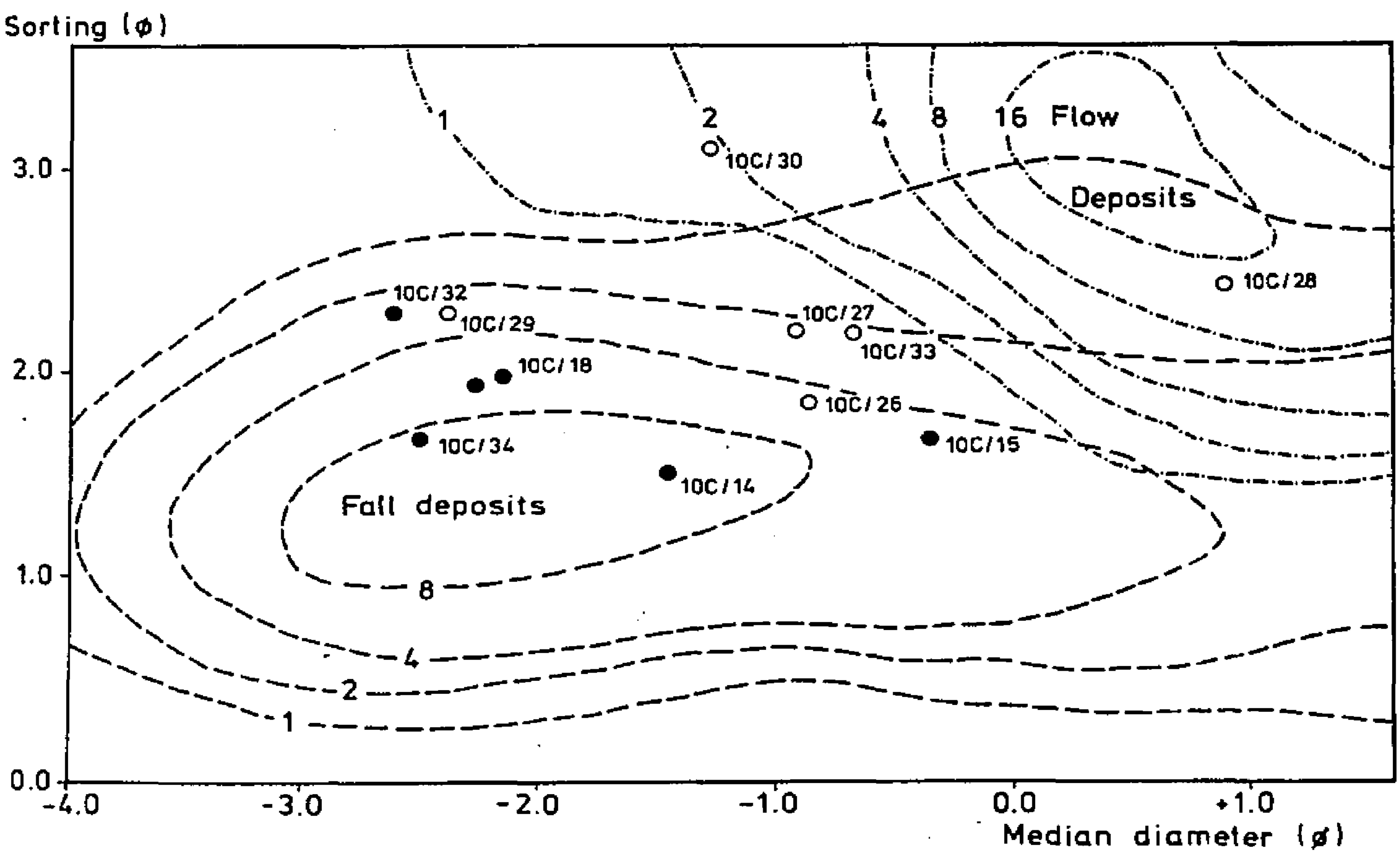
1. Definitions follow Williamson and Picard (1974).

Fig. 6.5: Grain-size distributions of nonlacustrine sediments from 0738-B20 (Appendix 7a).

a) (above) % gravel, sand, and fines
Open squares: colluvium
Open triangles: ephemeral-stream gravels
Black dots: pumice deposits

b) (below) Comparison of Md_{ϕ} and σ_{ϕ} of pumice deposits with characteristic distributions for pyroclastic-flow and -fall deposits (after Walker 1971).
Black dots: least weathered beds
Open circles: weathered beds
Contours in per cent.


SILT + CLAY
(<63μm)
100%
100%
GRAVEL
(>2mm)
100%
SAND
(2mm-63μm)
10C/1
10C/24
10C/23
10C/31
10C/21
10C/30
10C/33
10C/28
11/1
10C/19
10C/16
10C/34
10C/32
10C/29
10C/14
10C/26
10C/15
10C/25
10C/18
11/2
10C/27


Sorting (ϕ)
Median diameter (ϕ)
Flow
Deposits
Fall deposits
10C/30
10C/28
10C/32
10C/29
10C/27
10C/33
10C/18
10C/26
10C/34
10C/15
10C/14
0.0
1.0
2.0
3.0
-4.0
-3.0
-2.0
-1.0
0.0
+1.0
1
2
4
8
16

3) Poorly to very poorly sorted, light grey pumice gravels and gravelly sands with pale brown, calcareous, weathering horizons (e.g. the Abernosa pumice member). Pumice lapilli are generally very angular to subrounded (Plate 55), becoming rounded in weathered material. Lithic clasts are rare except for angular obsidian fragments. Normally graded when unweathered. Some thin bands of grey ash are present. In thin section, the majority of the grains are seen to consist of volcanic glass, although quartz, feldspar and pyroxene are occasionally visible.

4) Brown, pebbly, gritty pumice sands bearing truncated soils. Very poorly sorted. Seams of subangular to subrounded pumice pebbles may be found. Thin clay coatings are present on sand grains (Plate 54).

Reviewing all this evidence, 1) and 2) are clearly fluvial deposits derived from Alutu, whereas type 4) is interpreted as local slope material resembling unit UM5 in the Ajewa formation (late Holocene colluvium with landsnails). The Abernosa pumice member (type 3) presents more problems. Mr. Lloyd, of the UNDP Geothermal Project, has suggested that it consists of primary ejecta which have undergone a certain amount of weathering and possibly some reworking by surface wash or mud-flow activity. Most characteristics of the least weathered beds (samples 14-15, 18, 25, 32 in Fig. 6.5b) are more compatible with an air-fall rather than an ash-flow origin (Walker, 1971; Lirer et al., 1973; Sparks et al., 1973). These include a complete absence of welding, weakly-developed shower-bedding, and significant sorting by grain size (σ 1.48 - 1.98 Ø) and density. This sorting is expressed in the rarity of dense, lithic clasts. Normal grading of pumice is perhaps unusual in air-fall tephra but is unlikely to be found in ash-flow deposits (Sparks et al., 1973).

iii) Stratigraphic sequence

A broad outline of the succession in 0738-B20 has already been published (Gasse and Street, 1978a, b; Street, in press, a). But much more detail can now be added in the light of new ^{14}C dates (Street, Delibrias and Gillespie, in prep.) and comparisons with other sections, which confirm the pattern of minor fluctuations shown in Table 6.2.

The oldest sediment exposed is a coarse gravel containing cobbles derived from Alutu. It is overlain by members Deka Wede 1 and 2. Member 1a begins with a thin pebbly silt, which passes up into ripple-marked sandy silt containing thin seams of coarse sand and very sparse small planorbid shells. The overlying member 1b has an erosional contact at the base. Slightly current-bedded sand rapidly gives way to compact, very fine sand showing slight horizontal laminations (2.4 - 4.35 m) (Plates 42-45). A sample from 3.4 m contained a littoral, brackish-water diatom flora (Appendix 4) and abundant ostracods (Plates 44-45). As the sedimentary facies indicate, water depths were not great.

The strongly channelled upper surface of member 1b is unconformably overlain by 1.35 m of ephemeral-stream deposits. A very weakly developed palaeosol in the upper 20 cm of fluvial sand yielded a large quantity of obsidian tools and débitage, together with weathered bone refuse and dispersed charcoal. A small sample of charcoal was dated at 27,050 ± 1540 BP (SUA-588). The bone fragments were unidentifiable. However, the artifacts from this soil, and from the base of member 2a at P13, are of great interest because they appear to have been made by a primitive LSA flaking technique (H. Kurashina, pers. comm. 1975).

Unfortunately my collection was lost during the revolution in 1976, before it could be properly studied. If confirmed, the presence of such an industry would support an age well within the limits of ^{14}C (Clark and Williams, in press), as opposed to the lower Late Pleistocene age suggested by Professor K.W. Butzer (pers. comm. 1976).

Resting directly on the soil is the basal marl of member Deka Wede 2a. This has been dated at 24,000 ± 750 BP (GIF-3988). There is little sign of any recrystallization in this bed, so that the radiocarbon determination should be fairly reliable. It passes up into ripple-marked or plane-bedded fine sand and silt, followed by a compact marl which is finely laminated from 1.5 - 2.6 m and undoubtedly records the greatest water depth. This contains a freshwater diatom assemblage consisting of planktonic <u>Melosira</u> spp. with a variety of epiphytic and benthic taxa (Appendix 4). The upper surface of the marl was slightly truncated before deposition of the overlying, very poorly sorted, pebbly sands of member 2b. These contain occasional Corbicula shells and clearly mark a return to shallow and fluctuating conditions. At the top of member 2b, a thin calcareous silt dated 22,050 ± 650 BP (GIF-3987) represents a brief deep-water episode before the final regression. Slight weathering of this silt seems to indicate a period of exposure (Plate 48).

The lake deposits are conformably overlain by the Abernosa pumice member, which is 7.45 m thick at this site. At the top a weakly developed soil has formed in very poorly sorted, gravelly colluvium (samples 23 and 24, Fig. 6.5a; Plates 41, 54). This site on the flanks of a major gully system (Fig. 6.2) was evidently very suitable for human habitation. In the upper 20 cm of soil we found large numbers of LSA flakes, blades and scrapers, mingled with bone fragments and small pieces of charcoal (Appendix 6). The food refuse included the remains of a baboon, monkey, medium and large bovids, and disturbingly, a fragment of human cranium! (Plate 56).

The palaeosol has been dated at 11,870 ± 300 BP (SUA-494) on charcoal and 10,450 ± 180 BP (GIF-3986) on soil humus. A measured age of 12,200 ± 200 BP (GIF-3985) was also obtained on the overlying lake marl of member 3a. This cluster of dates is clearly inconsistent. The age adopted in chapter 8 for the Bulbula IIIA transgression is based on an appraisal of the relative reliability of these various materials.

Member 3a contains a freshwater diatom assemblage consisting of planktonic Melosira spp. accompanied by a rich epiphytic and benthic flora (Appendix 4). It is separated from member 3b by a slight disconformity and a layer of fluvial gravel with LSA implements. The presence of unbroken colonies of Melosira in member 3a indicate very calm conditions of sedimentation (Plate 49).

Member 3b is a very similar marl with scattered shells. Its flora closely resembles member 3a but is slightly richer in epiphytic and benthic elements (Appendix 4; Plate 50). Coupled with the lower carbonate content and marginally coarser grain-size, this could suggest somewhat shallower, higher-energy conditions, but more samples are needed. The upper surface of 3b is once again erosional. It is overlain by fluvial sand and then by a thick bed of normally graded, very angular to subangular, pumice lapilli. This represents a single air fall, so that only a short interval may have elapsed between deposition of members 3b and 3c.

Member 3c is a rather compact calcareous silt containing ostracod fragments, scattered shells and pumice pebbles. It indicates shallower conditions than either 3a or 3b. The upper surface has been strongly eroded and shows signs of weathering.

During the subsequent regression a thin layer of reworked pumice sand and two air-fall pumice units were laid down. A buried A horizon in the lower fall deposit gave a ^{14}C date of 8520 $\pm$ 200 BP (GIF-3984) on soil humus.

The lake once again submerged the site. Thin-bedded pumice gravels, waterlaid ashes, shell concentrates and calcareous silts (Plate 52) were laid down under rapidly fluctuating conditions. A particularly rich shell band with fishbones has been dated at 5790 $\pm$ 115 BP. It contained a typical freshwater lacustrine fauna with a significant benthic element (Melanoides, Bellamya, Caelatura and Corbicula). Member 4 is somewhat weathered (Plate 53), and has been truncated by erosion, but the sediments become finer showing that water depths were again increasing.

The section is capped by brown colluvium bearing a modern soil.

b. Profiles 10A, 12 and 13

Neighbouring profiles in the Deka Wede valley (Fig. 5.5) help to clarify the stratigraphy of the type section. At P13, the base of member 2a consists of up to 1.85 m of pebbly,pumice sands, occupying channels cut into member 1. The sands show trough cross-bedding and ripple marks, with a fluviolacustrine mollusc fauna (Corbicula, unionids, Bulinus, Valvata, small planorbids and Melanoides). Long-distance transport is ruled out, because the Corbicula valves are still articulated, though very crumbly. Two completely unworn obsidian pieces - a large blade and a distal blade fragment, both of LSA affinities - were found just below the top of the sands. A fluviodeltaic origin, during the transgressive phase of member Deka Wede 2a, seems indicated.

A rather similar situation occurs at the base of member 3 in profile 12 (Fig. 6.2, Plate 57). A deep channel, incised through member Deka Wede 2 and the overlying Abernosa pumice member, has been infilled first by strongly cross-stratified, pebbly, pumice sand and then by contorted lake beds of member 3. The sand contains uncommonly large articulated valves of Caelatura aegyptiaca dated at 11,800 ± 150 BP (SUA-504). Their size suggests a riverine form. Probably they lived in the bed of a perennial stream, close to its entry into the lake.

Member Deka Wede 4 can be followed upslope from P10C through P10B to P10A, where rich shell bands occur in bedded, water-worn pumice gravel at 1643 m. The fauna includes large numbers of Bellamya and articulated unionids, indicating that the shells were not transported far. The principal shell bed dips at 23° into the terminal Pleistocene gully system (N 50° E), and its measured ^{14}C age is 6060 ± 120 BP (SUA-499). The P10A deposits are interpreted as beach gravels, and it seems probable that the shells dated 5790 BP in P10C were transported downslope into deeper water, especially as no articulated bivalves were present.

c. The head of the Deka Wede gully

Deposits like those in P10A also occur close to the uppermost shoreline (Fig. 5.5). At P1, the edge of the palaeolake is marked by a silica-cemented gravel terrace at 1672 m, underlain by cross-bedded deltaic sands (Plate 25). This terrace probably dates from the latest highstand which reached overflow level. A little further downslope, at P14, gravelly beach deposits dip lakeward at $\leqslant 7.5^{\circ}$. An extremely shelly grit at 1651 m contains articulated Corbicula and unionid valves, Bellamya, and very large, delicate Lymnaea natalensis shells up to 20 mm in size, which could not have been transported any distance. The presence of these large, fragile pulmonates, similar to those now living in the Lower Awash marshes between Dubti and Assaïta (D.S. Brown, in litt. 15/10/75) probably indicates the close proximity of marginal swampland.

Still further downvalley, at P4-9, two principal waterlaid units are present, separated by air-fall pumice layers and a buried soil. These can be correlated with members Deka Wede 3 and 4 in the type section.

In the rest of this chapter the evidence from elsewhere in the basin is described and compared with the key Deka Wede sequence, which is summarized in Table 6.2.

III. The Kurkura diatomites: sections 0738-B3 and 0738-B27

These two sections are situated on the northern edge of the South Basin, where water depths were increasing towards Abiyata (Fig. 6.1). B27 outcrops in the deepest part of the Bulbula gorge, and takes us further back into the Lake Pleistocene than B3, although the latter has been studied in more detail following earlier work by Gèze (1974, 1975). The most interesting feature of both sections is the impressive development of Late Pleistocene diatomites within the Bulbula formation. In contrast, the Holocene is very poorly represented, for reasons which will be discussed in chapter 7DIIId. Because the facies are so different from the Deka Wede area, correlations between the two are based on the position of the Abernosa pumice member and on the grain size and relative thickness of the lacustrine units.

a. Section 0738-B3

Section B3 outcrops where several cattle tracks converge on the Bulbula River from the west, about 3 km north of Bulbula village (Fig. 5.1, pt. B). The general setting is illustrated in Grove et al. (1975, plate VIIb). At river level, the section begins with a shallow-water pumice sand (Fig. 6.6). This is overlain by 4.5 m of calcareous, diatom-bearing silts and very fine sands, and then a massive bed of almost structureless $CaCO_3$-free diatomite (Plate 58). At 10.40 m the diatomite abruptly gives way again to pumiceous shallow-water sands and coarse silts. A return to deeper conditions followed, marked by deposition of a distinctive "pink-orange" prismatic diatomite (10YR 8/3, dry) interrupted by multiple, dark-coloured, water-laid ashes. At 14.0 m the impending retreat of the lake made itself felt. Calcareous, sandy silts and fine gravel were deposited. Final desiccation is recorded by rapidly alternating, pebbly sands and ostracod-rich silts, which seem to represent

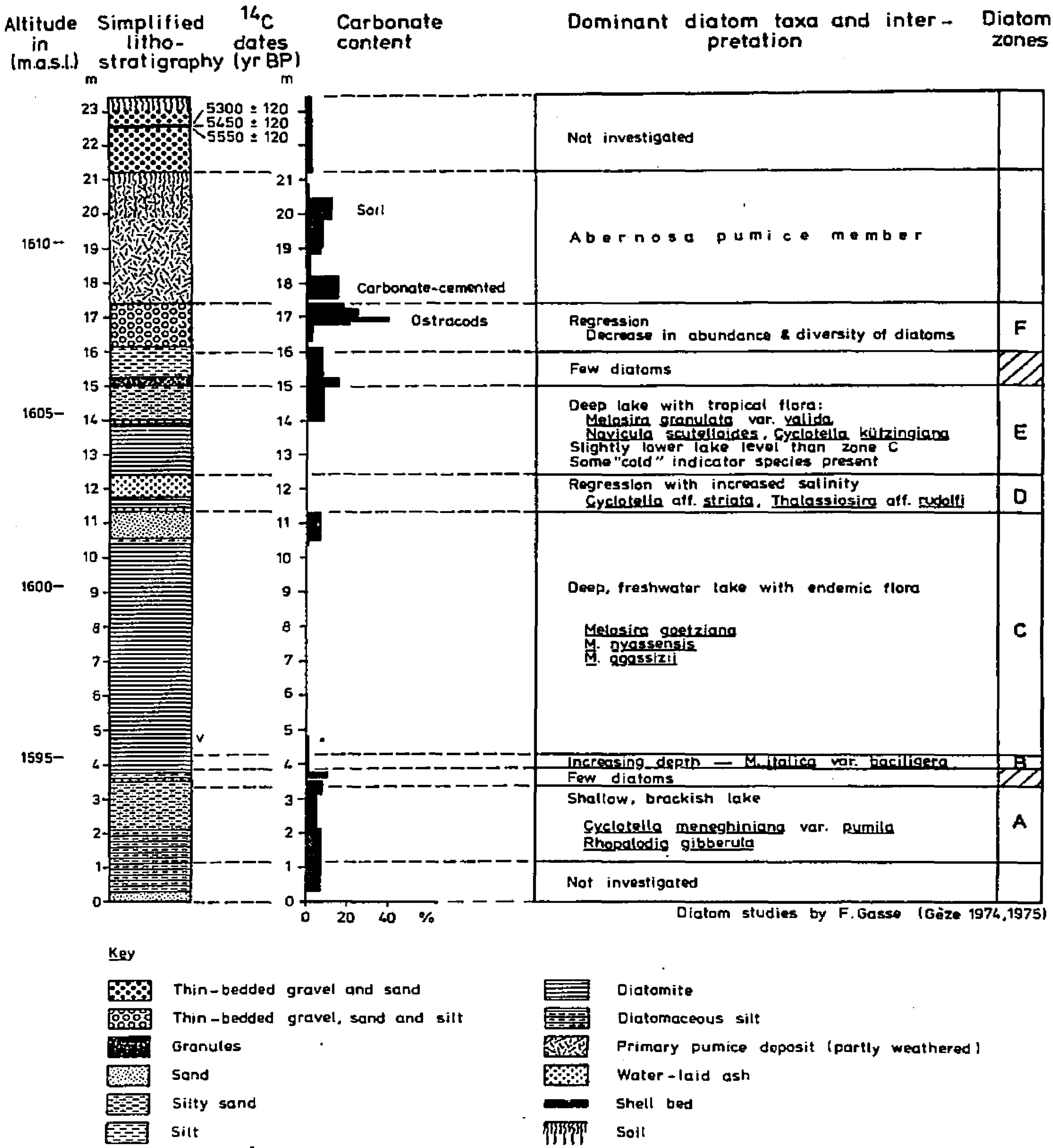


Fig. 6.6: Stratigraphy of Kurkura section 0738-B3.

an oscillating beach-ridge and lagoon environment like the present north shore of Abiyata (chapter 2E IV).

Overlying the lake sediments we find 3.8 m of weathered air-fall pumice beds with a weakly developed palaeosol at the top. These can be correlated with the Abernosa pumice member. The section is completed by 2.15 m of horizontally-bedded beach grit and sand which were not described by Gèze (1975). A shell band is present at 22.50 m (55.3 m.S.D.). This was sampled in 1970 by Grove and Goudie (Grove et al., 1975, table III, no. 12). A second sample collected from the same spot in 1975 gave the following ^{14}C results:

Table 6.3: Radiocarbon dates from section 0738-B3 in the Kurkura area

Uncorrected date (BP)	Lab. no.	Species
5300 ± 120	GIF-4014	Melanoides tuberculata
5450 ± 120	GIF-4012	Bellamya unicolor
5550 ± 120	GIF-4013	Bulinus "truncatus"

These dates allow firm correlation of the upper lake beds with member Deka Wede 4 in section B20/10C. The close accord between measurements on different species is reassuring, in view of the doubts expressed by Haynes and Haas (1974, p.376). The fauna is typically lacustrine, and small numbers of Bellamya confirm the dominance of freshwater conditions.

In Figure 6.6 the stratigraphy of section 0738-B3 is compared with the diatom biozones established by F. Gasse (Gèze, 1975; Gasse, 1975, pp.368-369). This correlation is based on the most prominent ash bands in the sequence (Gèze, 1974, fig. 64).

There is a strong relationship between the sedimentary facies and the diatom floras. Two major deep-water episodes, represented by the purest diatomites, were preceded and followed by periods of shallower, brackish conditions with fluctuating salinity. There is always a certain lag, however, between the change from a deep to a shallow-water sediment type and the corresponding passage from planktonic to littoral floras. Probably this can be explained by partial reworking of diatom frustules by wave action.

In the past, the age of the undated Kurkura diatomites has been rather controversial. Gèze (1975) mistakenly correlated them with the Holocene Kertefa diatomites in the vicinity of sections 0738-B17 and B18 (chapter 6CV). On this basis, he published an outline of the diatom stratigraphy in 0738-B3 as proof of two Holocene highstands. It is now evident from the ^{14}C dates at the top of B3 and from the great thickness of overburden in B27 that the Kurkura diatomites must be Late Pleistocene in age (Gasse and Street, 1978a). This discovery was anticipated by Dr. Gasse in her thesis, when she noted the great similarity between their diatom flora and the early part of the Abhé III lacustral episode in the Afar (30,000-25,000 BP) (Gasse, 1975, p.369) (chapter 10).

More precise dating of the diatom sequence is only possible through correlations between 0738-B3, B27 and B20 (Table 6.4, and section c, below).

b. Section 0738-B27

The sediments in B27 outcrop at higher elevations than their equivalents in B3 (Fig. 6.1). In this area the precipitous sides of the Bulbula gorge are 52.5 m high (Plate 9), and the measured section is exposed along the sides of a cattle track which winds precariously down to the river. The lowest 12.2 m are unfortunately hidden by talus. Future exploration by boat in search of datable materials is likely to be profitable.

At its base, the B27 section begins with coarse, fluvial gravels (Fig. 6.7). The pebbles and cobbles are angular to subrounded. Pumice and obsidian clasts dominate, reaching 0.2 m in size. The likeliest source is Mt. Alutu. These deposits are much coarser than either present gully-floor or Bulbula River gravels, and most nearly resemble those underlying member Deka Wede 1a, at a very similar elevation (Fig. 6.3).

Overlying the irregular surface of the basal gravels are 22.7 m of lacustrine beds, in which we can recognize many of the features of the B3 diatomites. Of these, the most distinctive are the pattern of carbonate

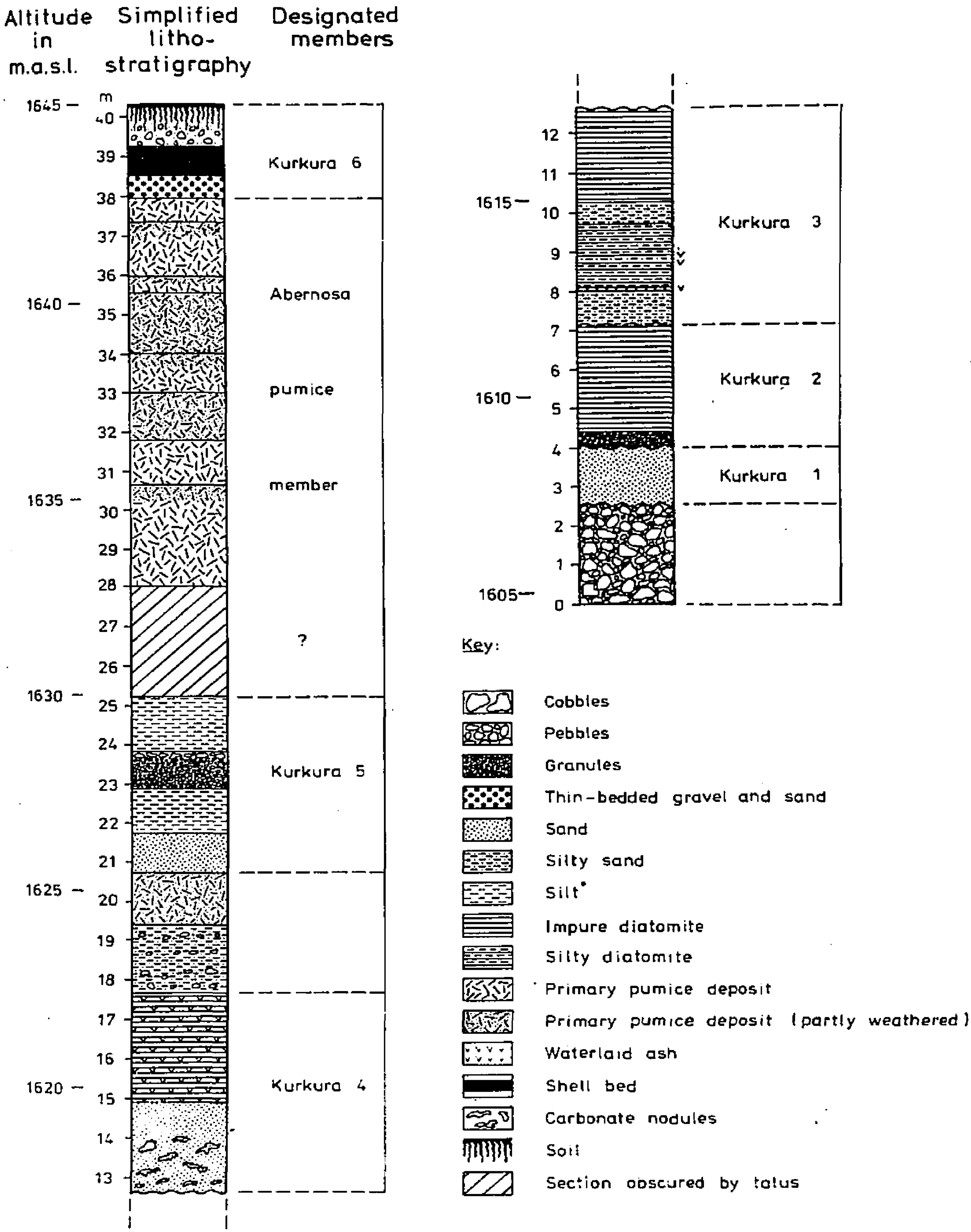


Fig. 6.7: Stratigraphy of Kurkura section 0738-B27.

variations and the "pink-orange", prismatic diatomite with multiple thin ash bands at 15.05 - 17.65 m. Overall, the diatomites are less pure than in B3, which is to be expected in a shallower environment.

The lower lake sediments are preserved by a cover of 13 m of coarse lapilli and ash, representing a minimum of 13 individual eruptions. This is the greatest thickness reached by the Abernosa pumice member. It consists of typical, shower-bedded pumice with a few lithic fragments. Many of the beds fine upwards and show brownish colouration at the top, indicating weathering. Powdery films of $CaCO_3$ are common on the clasts, especially towards the summit.

The section is capped by 2.35 m of pumice shingle, including a 75 cm-thick shell bed with articulated unionids. These thin lag gravels are the sole evidence for Holocene lacustrine conditions at this site, but do suggest a Late Pleistocene age for the diatomites below.

c. The Kurkura diatomite sequence

Minor disconformities, or signs of fluvial reworking, divide the diatomite sequence in B3 and B27 into five members (Kurkura 1-5). The Holocene shelly grayels at the top of both sections form member Kurkura 6 (Table 6.4). The Abernosa pumice member, which is present in both sections, provides the most important link with the Deka Wede sequence.

Members 1 and 2 are not represented in B3. It is assumed here that they are lateral equivalents of member Deka Wede 1 (Fig. 6.3). This part of the Bulbula type section lies some 5 m lower in elevation, so that evidence for retreat of the lake is more equivocal. But the signs of shallower conditions between members Deka Wede 1a and 1b may correspond to the slight erosional unconformity between members Kurkura 1 and 2 in B27.

Members 3 and 4 are the purest diatomites, and both show carbonate minima. They are assumed to be lateral equivalents of members Deka Wede 2a and 2b. The ash bands in the "pink-orange" diatomite (member Kurkura 4) and at the base of member Kurkura 3 reinforce the correlation shown in Table 6.4. Member 5 in B27 is likely to be represented by deeper-water

Table 6:4: Provisional correlation between the members of the Bulbula formation in the Deka Wede and Kurkura areas

Deka Wede area (0738-B20) Members	Kurkura area (0738-B27) Members	Kurkura area (0738-B3) Lateral equivalents (m)	Kurkura area (0738-B3) Diatom zones	Lake-level sequence inferred from Kurkura area	Lacustral phases
Deka Wede 4	Kurkura 6	23.35-21.20	?	Highstand >1645 m	BULBULA IV
Deka Wede 3				?	BULBULA III
Abernosa pumice member	Abernosa pumice member	Abernosa pumice member	Abernosa pumice member	Lowstand <1608 m	
		17.40-16.15	F	Recession	? — — ?
	Kurkura 5	16.15 - 14.00	E	Minor rise >1630 m	BULBULA IIC
? ? Deka Wede 2b	Kurkura 4	14.00-10.40	D	Highstand >>1622 m	BULBULA IIB
Deka Wede 2a	Kurkura 3	10.40 - 0.00	C B A	Highstand >>1617 m	BULBULA IIA
				< 1591 m	
Deka Wede 1b	Kurkura 2			Minor rise > 1612 m	BULBULA IB
				<1609 m	
Deka Wede 1a	Kurkura 1			Minor rise > 1609 m	BULBULA IA
				Lowstand <1607 m	

Key

- Airfall pumice
- Fluvial gravels
- Diatoms very scarce
- Depositional hiatus
- Erosional unconformity

facies in B3, which is situated more than 20 m lower. A provisional correlation with the fine silt at 15.25 - 16.15 m in B3 is suggested. The deposits of diatom zone F indicate oscillation of the shoreline across the B3 site, and are unlikely to have a lacustrine counterpart at B27.

d. Summary

In sections B3 and B27 we have evidence for two principal Late Pleistocene deep-water episodes which probably resulted in overflow. These are believed to be equivalent to lacustral phases Bulbula IIA and IIB (Table 6.2). They were preceded and followed by smaller oscillations of the lake (Table 6.4). A single Holocene fluctuation (Bulbula IV) is recorded at both sites.

IV. The Southern Bulbula Plain

a. The Abiyata Core and the Bulbula Village Section (0738-B8)

i. Stratigraphy

The most detailed record of the Late Quaternary in the Abiyata Basin comes from the 1974 Ministry of Mines' core, now under process of study. 162 m of lacustrine and fluvial sediments were penetrated by the borehole, but only the uppermost 25 m will be discussed here. The great value of this core for our purposes is that the Holocene section has yielded five ^{14}C dates on organic matter, which confirm the less reliable shell dates obtained elsewhere.

The drilling site was located on the northern edge of Abiyata, about 0.67 km north of the November 1972 shoreline (Fig. 5.1, point C), but within the limit of the 1595-1600 m lake. Its estimated elevation is 1585 m. The continuous core, which was intended for salt and diatomite prospection, was 10 cm in diameter from 0 to 120 m and 5 cm from 120 to 162 m. Recovery was excellent - about 95%. However some segments were very disturbed, even though the hole was drilled dry. Shallow-water sands were particularly deformed during extrusion. We were granted permission to take samples of dried core material at 25 cm intervals.

Figure 6.8 shows the first results from the upper 25 m. The diatom and phytolith data are shown by courtesy of Mlle. Claudine Descourtieux. The diatom-content curve appears to be a particularly good indicator of water depth at this site. So far her interpretations substantially confirm those from the type section; but as work is still in progress, what follows should be regarded as a provisional summary.

The section of core in Figure 6.8 begins with greyish pumice sands (10YR-7.5Y 5/1 to 8/2) which extend from 25 to 19.4 m. They vary from gritty coarse sand, with rhyolite pebbles up to 4 cm long in places, to very fine sand and silt. Clasts of lake sediment are present in some beds. Both carbonate and organic matter are low but variable in this part of the core (generally < 7% and < 2.5% respectively). The sands seem to record a long interval during which only fluvial or shallow-water lacustrine sediments were deposited. Indeed, in the whole interval from 56 m to 19.4 m, finer-grained lake beds are present only between 38 and 43 m. It seems certain that the deposition of this great thickness of sediments must have been accompanied by subsidence of the basin floor.

Between 19.4 and 14.2 m, there is an alternation from white diatomite (2.5Y 8/2) or diatomaceous silt to coarse grey pumice sand (2.5Y 4/1 to 7/1). The organic and carbonate content are correspondingly variable (0 - 3.3% and 1 - 14% respectively). Unfortunately, this segment was seriously disturbed during extrusion. But it seems reasonable to conclude that lake-level had begun to rise in an oscillatory fashion.

Two beds of impure,white to pale yellow diatomite (7.5Y-2.5Y 8/1 to 8/3) occur between 14.2 and 10.1 m, separated by silty diatomite and fine greyish-brown sand (2.5 Y 5/2). These beds can be tentatively correlated with members Kurkura 3 and 4-5. Both show carbonate minima (0%) and organic-carbon maxima (2.8 - 2.9%). They appear to represent two major deep-water phases with an intervening partial regression which is not recorded by the diatom floras.

The sediments then become more sandy from 10.1 to 5.58 m, suggesting a period of lower lake level. They consist mainly of pale brown (2.5Y 6/2) to light grey (2.5Y 7/2) medium to coarse pumice sands, although a brief return to brown clayey silt with sand lenses (2.5Y 5/3) occurs at 8.3 - 8.8 m. A ^{14}C sample from 8.25 - 8.55 m has been dated at 10,370 ± 180 BP (GIF-3969).

Then suddenly, at 5.58 - 5.35 m, we encounter a slightly clayey, dark greyish-brown sand (2.5Y 4/2) containing small root fragments and abundant phytoliths. This appears to be a buried A horizon. It is dated 9950 ± 170 BP (GIF-3968). The high $\delta^{13}C$ value of the organic matter (-12.5‰) suggests that it is derived from sedges such as papyrus; tropical grasses; chenopods; or plants which draw heavily on dissolved bicarbonate for photosynthesis, such as submerged aquatics or phytoplankton (Smith and Epstein, 1971; Stuiver, 1975; Thompson, 1976). Values in this range are typical of eutrophic lakes like George which support dense blooms of blue-green algae (Viner, 1977).

The palaeosol is overlain by thinly bedded lake deposits (5.35 - 2.33 m), consisting of grey, diatomaceous clays or silty clays (2.5Y 5/1) with numerous bands of light-grey (5Y 6/1) to pale yellow (2.5Y 8/3), silt- or sand-sized material. Some of the coarse bands appear to represent direct ash falls into the lake, but conditions do appear to have fluctuated considerably. In the Bulbula Village section (0738-B8), only 5.8 km to the NE and 15 m higher, the base of the upper lake beds also consists of an alternation of prismatic, calcareous silts and waterlaid ashes. The clays in the core contain 7 - 19% $CaCO_3$ and 1.6 - 6.6% organic carbon, partly in the form of macroscopic plant fragments. Diatom frustules are also moderately abundant ($\leqslant 10^9$ valves/g dry sediment) (Fig. 6.8), although masked by organic matter. Three slightly inconsistent ^{14}C ages were obtained for this core segment: 9550 ± 170 BP (GIF-3967) at 4.90 m; 9810 ± 170 BP (GIF-3966) at 4.00 m and 7970 ± 150 BP (GIF-3965) at 2.60 m. However the lower two do not differ significantly from GIF-3968. The $\delta^{13}C$ values range from -18.2 to -23.3‰, which suggests a different origin for the organic matter from GIF-3968; being typical of normal terrestrial C3 plants and of floating aquatics or phytoplankton which utilize atmospheric CO_2 (Smith and Epstein, 1971; Stuiver, 1975).

The uppermost 2.33 m of the core consist of light grey pumice silt and sand (2.5Y 7/2), bearing a weakly developed modern soil. Rootlets are common in the uppermost 1.05 m. On first sight these sediments appear to record the final retreat of the lake. However, the diatom stratigraphy below suggests a more complex history.

ii. Diatom zones

The diatom studies by Mlle. Descourtieux (Descourtieux, 1977; Gasse and Descourtieux, in press) provide evidence for an undifferentiated Late Pleistocene deep-lake phase, represented by the white diatomites, and a tripartite Holocene highstand (Fig. 6.8). During the Late Pleistocene maximum (zone A_2) the dominant species were Cyclotella ocellata and Melosira granulata var. valida, which indicate a very dilute, rather oligotrophic, lake of low alkalinity. The high abundance of the cosmopolitan species Cyclotella ocellata, which is not found in appreciable numbers in any modern African lake (Richardson and Richardson, 1972), coupled with a small peak of nordic and alpine diatoms, may suggest that water temperatures were somewhat cooler than at present.

The Holocene lacustral phase began with an abortive transgression, Abiyata II, just before 10,370 BP. The lake waters freshened considerably but still remained somewhat brackish, judging from the persistence of Rhopalodia gibberula in zone B. Lake level then fell again, allowing the development of wetland vegetation. The sparse microflora in the palaeosol at 5.4 m implies very shallow, alkaline water and high or variable salinity. This is underlined by the presence of subaerial species such as Hantzschia amphioxys (4%). The most probable interpretation of the isotopic, diatom and phytolith evidence is that this was a periodically flooded Sporobolus grassland with alkaline pools, like those found today around Langano and Abiyata (Bolton, 1969; Makin et al., 1976).

Between 9950 BP and 9810 BP, a sudden rise in level took place (Abiyata III). The lake waters rapidly became more dilute. Shortly before 9810 $\pm$ 170 BP a planktonic diatom flora became established (zone C_3). Stephanodiscus astraea and its varieties are characteristic of deep, oligohaline lakes with moderate alkalinities and low dissolved silica contents (0.1 – 10 mg/l) in their surface waters (Chapter 8BIIi). Fluctuations in the abundance of Cyclotella ocellata and Fragilaria brevistriata indicate minor oscillations in lake level, and small numbers of temperate species are still present.

The Abiyata III highstand lasted until shortly after 7970 $\pm$ 150 BP. The reappearance in zone C_4 of species common in East African soda lakes such as Cyclotella meneghiniana, Thalassiosira rudolfi and Nitzschia spp. marks a return to shallower, alkaline, conditions. Many of these taxa can take up a planktonic existence in the dense blooms of blue-green algae which develop in shallow, eutrophic lakes (Hecky and Kilham, 1973). The regression is most marked between 1.25 and 1.55 m, where diatoms are scarce. It is possible that the site dried out at that time. Brackish-water, littoral species then returned as lake level rose once more.

From 0.75-0 m the diatom flora records a final deep-lake phase (Abiyata IV). This seems surprising because the sediments are relatively sandy. Reworking is possible but unlikely, because a typical transgressive sequence of diatom zones is present (D_1 - D_3). The subsequent regression and the rise to 1595-1600 m have left no trace in the sedimentary record, for reasons which remain obscure.

Comparison with modern samples suggests that the lake waters during phase Abiyata IV were oligohaline, less alkaline than during phase III, and probably richer in silica (>10 mg/l). It was also slightly more oligotrophic. No temperate diatom species are present.

iii. Sedimentation rates

If it is assumed that the ^{14}C chronology of the core is reliable, with the exception of the 9550 BP date, which may have been contaminated by a small amount of younger carbon, then it becomes necessary to explain the large fluctuations in the rate of sedimentation experienced at the core site. Deposition appears to have been relatively slow during deep lake phases: only 0.76 m/10^3 yr between 9810 $\pm$ 170 and 7970 $\pm$ 150 BP as compared with 7.14 m/10^3 yr from 10,370 $\pm$ 180 to 9950 $\pm$ 170 BP and 10.0 m/10^3 yr from 9950 to 9810 BP, for example. The Kurkura diatomites are also much thinner here than in 0738-B3 or B27. Either sediment inputs and diatom productivity were lower at the centre of the lake than near the margins, or else the sediment was somehow prevented from accumulating, perhaps by current winnowing.

There also seems to be a hiatus, perhaps an erosional unconformity, between 10.0 and 8.8 m: an impression which is confirmed by comparison with the Bulbula Village section (0738-B8). In B8 the pale yellow, impure Kurkura diatomites at the base of the section are overlain by a fluvial silt with staining along root channels. This is followed by a regressive lacustrine sequence similar to 0738-B3, consisting of alternating beds of blocky silt, gritty silt with ostracods[1] and waterlaid ash. The succeeding Abernosa pumice beds are still 4.3 m thick in Bulbula Village (Fig. 6.1), although they seem to have been reworked in the core.

It is easier to explain the high rate of deposition between 10,370 ± 180 and 9950 ± 170 BP, when coarse sediment may have been brought in by streams from the exposed Bulbula Plain.

The period 9950 ± 170 to 9810 ± 170 BP is particularly interesting. It has already been suggested in chapter 2DVII that the onset of overflow from Abiyata should be accompanied by a rapid dilution such as occurred at this time. Until the surface of Lake Shala rose above the Gorgeza col, the level of Abiyata would remain more or less constant at about 1592 m, while receiving a greatly enhanced input from Lakes Ziway and Langano. This might explain the very rapid and variable sedimentation at the core site, which lies close to the Bulbula and Horocallo inflows. It may also account for the slightly chaotic radiocarbon dates.

iv) Summary

The Abiyata core provides evidence for three, stratigraphically distinct, Holocene lake-level maxima (Abiyata II-IV), the first of which was of smaller amplitude than the others. The Late Pleistocene diatomites at this site are unusually thin, and only one major deep-lake phase can be demonstrated (Abiyata I). Diatom studies of the core confirm that the Holocene lake-level

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1. Note added in proof:
A sample of ostracods from this section has now been dated 24,590 ± 500 BP (see p.351).

fluctuations deduced from the Deka Wede type section resulted in significant variations in salinity in the Abiyata Basin; depending on whether it was communicating with Lake Shala or not. The two sequences are correlated in Table 6.9 at the end of the chapter.

b. The Horocallo channel: section 0738-B7

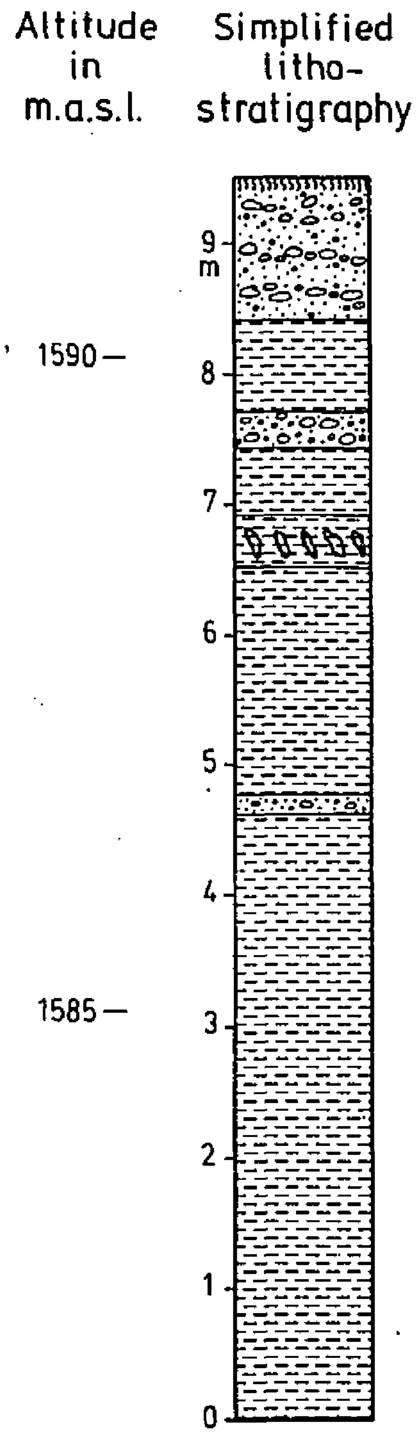
This section is quite different in character from those previously described along the Bulbula River, and is included here because it exposes shallow-water deposits of the late Holocene 1595-1600 m lake. It lies at the northern end of the Horocallo spit separating Abiyata and Langano (Fig. 5.1; Plate 34). Whether any earlier sediments are included is uncertain; but a high degree of reworking is to be expected in this exposed location.

The deposits consist essentially of a thick sequence of ostracod-bearing, fine silts with originally planar laminations, becoming more sandy and pebbly towards the top. Tilapia bones and scales (up to 1 cm across) are increasingly common above 4.80 m: the bones are unusually light-coloured and fresh in appearance. No shells are present. Certain sedimentary structures record past deformation by liquefaction, perhaps as a result of earthquake activity or of rapid depositional loading of the nose of the spit. They include convolute laminations and brecciated silt clasts "floating" in a coarser matrix (Born, 1972). The upper, sandy beds are particularly severely affected (Fig. 6:9).

So far, unfortunately, organic remains have not been found in sufficient quantities to enable these beds to be dated.

V. The northeastern Bulbula Plain

Under this heading are included two groups of sections on the southern margins of the North Basin. The first, 0738-B16 to B19, outcrops in the sheltered Kertefa embayment at the foot of Alutu. The second consists of two shallow exposures beside the Bulbula near its outlet from Lake Ziway: 0738-B2 and B26. In several of these high-level sections the Lake Pleistocene is represented by shallow-water or beach facies, which outcrop beneath the Abernosa pumice member. In the Kertefa embayment, there are



Description	Munsell colour
Gritty, silty sand with lenses of pumice pebbles and rolled fish bones and scales	2.5Y 5/2 to 6/1
Silt with convolute bedding and lenses of coarse pumice sand. Fish bones present	2.5Y 6/2
Coarse sand with pebbles and granules. Shows convolute bedding	2.5Y 5/1
Silt with convolute bedding	2.5Y 6/2
Silt with pseudobedding and lenses of sand and granules	2.5Y 6/3
Silt, coarsening upwards. Laminated bedding with occasional disturbances resembling bomb sags. Scattered, rolled fish bones	2.5Y 7/1
Gritty, fine to medium pumice sand with small pebbles	2.5Y 6/2
Finely laminated silt with occasional disturbances of the bedding resembling bomb sags	2.5Y 6/2

Fig. 6.9: Stratigraphy of Horocallo section (0738-B7).

very thick, diatomaceous Holocene deposits of quite different character from the Late Pleistocene Kurkura diatomites: it is here that Gèze obtained his radiocarbon dates (Gèze, 1975), which we can use as a key to the chronology of the Kertefa area.

a. The Kertefa diatomites (sections 0738-B16 to B19)

In this area, Holocene lake sediments reach their most spectacular development (Plates 59-61). The exposures occur along deep, subparallel gullies, which enter the Bulbula River from Alutu (Fig. 5.5; Plate 39). The northernmost, B16, heads among rhyolite flows and domes, whereas the two other large streams further south, B17 and B19, derive most of their runoff from pumice-covered slopes. Moreover, the measured section B16 occupies a very sheltered site at the foot of the mountain, in contrast to the other localities which were situated further from the shore. Section B19/1, across the river, lies close to the southeast tip of the Dakadima spit, as delimited by the 1660 m contour on the 1:50,000 map.

Frederic Gèze (1974, 1975) was the first to publish an outline of the stratigraphy in this area. His "Abelosa Section" appears to correspond to B17, and the site of his upper shell sample to B18. However many of Gèze's detailed interpretations require re-examination.

The Holocene Kertefa diatomites attain their greatest thickness immediately to the east of sites B17 and B19/3, where they underlie a prominent constructional terrace at $\geqslant$ 1660 m. This seems to be equivalent to Laury and Albritton's terrace IV in the West Ziway area (Fig. 5.9). Towards the river, the lake sediments have been truncated by varying degrees of erosion.

Figure 6.10 is an attempt to correlate the principal exposures in this embayment. The profiles have been lined up at the base of the Holocene deposits. The essential features of the stratigraphy are as follows:

Water-laid pumice gravels of probable Late Pleistocene age are present at the base of B17 and B19/1, at elevations of 1635-1641 m. Probably the other sites lay in deeper water. In B17 these gravels are well-bedded but

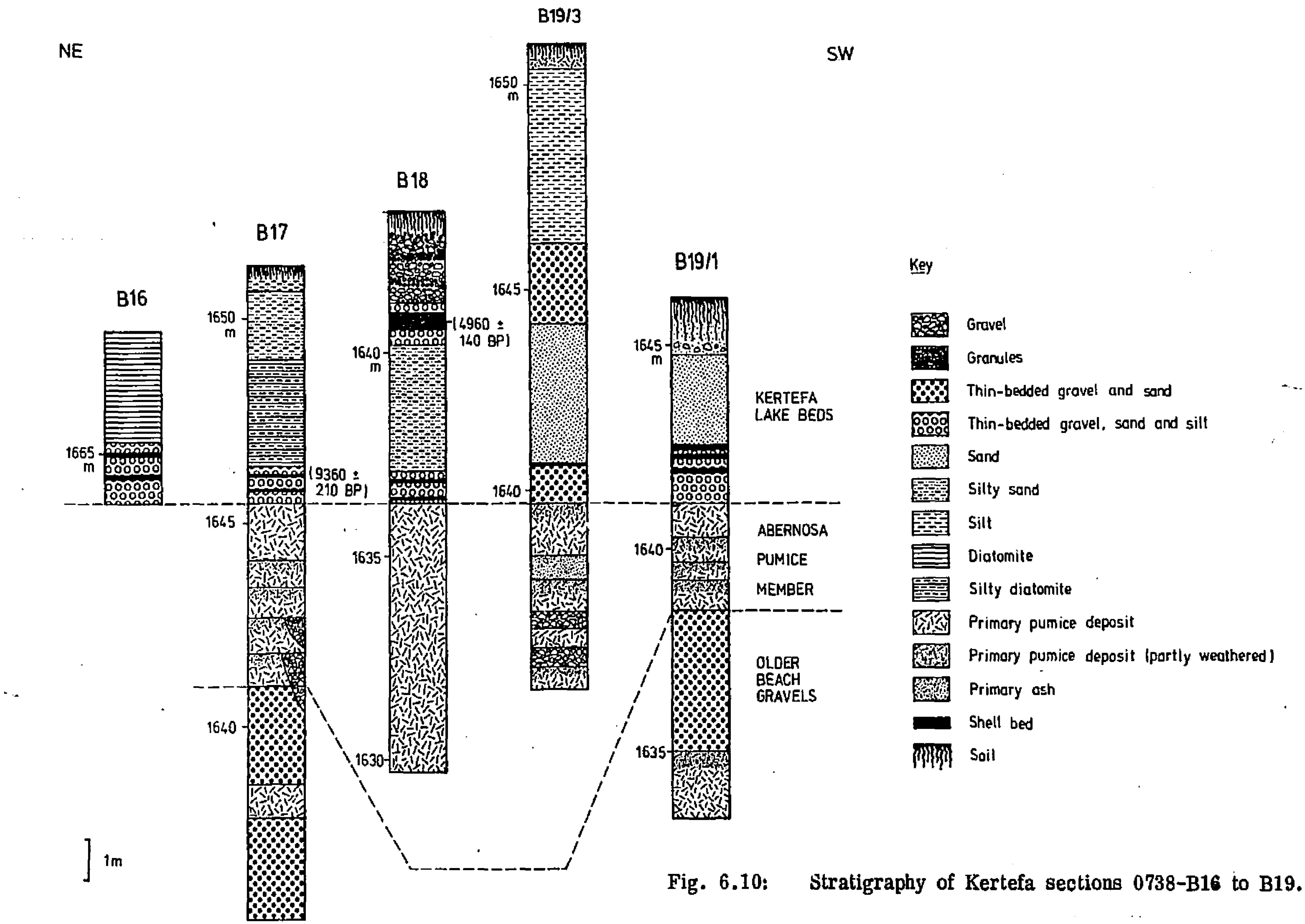


Fig. 6.10: Stratigraphy of Kertefa sections 0738-B16 to B19.

unfossiliferous, and resemble beach shingle. At B19/1, foreset bedding is locally present, and the deposits contain tiny shells of some interest. These include juveniles of several lacustrine or riverine taxa such as Lymnaea and Bulinus, a land snail (Cecilioides sp.) and numerous valves of the tiny bivalve Pisidium subtruncatum Malm (J.G.J. Kuiper, in litt. 28/3/76). This is a north European and North American species whose only foothold in Africa is now the coastal region of Algeria (Kuiper, 1966). It occurs subfossil in the deposits of Mega Chad and in the "Palaeolithic lake beds" of the Faiyum in Egypt (Gardner, 1932; Kuiper, 1968), but this is the first record from eastern Africa.

Overlying these older water-laid deposits are bedded pyroclastics, ranging from 2.65 to more than 6.6 m in thickness (Plate 62). These can be correlated with the Abernosa pumice member. Although Gèze (1975, fig. 3) interprets them as shallow-water lacustrine sediments, an air-fall origin seems more likely, for the reasons given when discussing the Deka Wede section (section IIaii, above). The original character of many beds has however been obliterated by weathering (Plate 63). In B17, contemporary fluvial activity is indicated by shallow sand and gravel-filled channels which cut through the tephra into the underlying gravels.

The base of the upper, Holocene, lake deposits lies at 1636–1645 m. It is immediately recognizable by the rapid alternation of three sediment types: shell concentrates, often with a brecciated base; fine pumice gravel; and well-sorted silts or very fine sands. The latter are very frequently calcareous. An oscillating shoreline can be envisaged. Like the regressive sequence in 0738-B3, the fluctuating north shore of Abiyata serves as a useful model, but since the lake waters were becoming increasingly dilute it was possible for mollusca to flourish. Fish remains, including a Barbus spine, were found in the upper shell band in B17. This is believed to be the level dated by Gèze (1975) at 9360 $\pm$ 210 BP.

In all the sections we then find a thick sequence of fine-grained delta-slope deposits (Born, 1972) (Plates 59–61). These are the Kertefa diatomites. At maximum they reach 10.1 m in thickness upstream from the B17 site, and

9.8 m at B19/3, where they are divided into two units by 2 m of pebble- and cobble-gravel. There is a lateral gradation from fluffy, calcareous diatomites with occasional sand laminae in B16 to diatomaceous silts in B17 and pumiceous silts or finest sands in B18 and B19. Low-angle foreset bedding is present in B16 and B19/3 (Plate 61). Fine, horizontal lamination occurs in B17 (Plate 60). The diatomaceous silts in B17 contain a freshwater diatom flora dominated by Fragilaria brevistriata, accompanied by numerous epiphytic species. A number of planktonic taxa such as Cyclotella ocellata and Stephanodiscus spp. are also present (Appendix 4).

It is interesting to relate the observed variations in facies to the palaeogeography of the embayment. The Kertefa diatomites become more detrital and less biogenic southwards as the source-area geology changes. In addition, the upper fine-grained unit in section B17 becomes increasingly coarse towards the shore. But where the shoreline was rocky, as at B16, fluffy diatomites appear to be a truly marginal facies, vanishing beneath a cover of recent fan gravels at the foot of the massif (1661 m) (Fig. 5.5).

At B18 the Kertefa diatomites are truncated by a shell bed dated at 4960 ± 140 BP (Gèze, 1975). The shell-bearing sand is overlain by 2.25 m of water-laid gravels, crowned by a modern soil. Its mollusc fauna is typically lacustrine, with a large benthic element. We find such freshwater indicators as Bellamya unicolor and Caelatura aegyptiaca, which is found with both valves articulated, testifying to limited transport. Ostracods from this bed include Cyprideis sp., Ilyocypris sp., Metacypris sp. and Darwinula sp. (R.H. Bate, in litt. 3/3/76). The last three taxa tend to be restricted to freshwater environments: in the U.K., Darwinula and Ilyocypris inhabit streams or ditches, which in the case of the Fens may become temporarily brackish from high tides. Darwinula stevensoni was identified living in Lake Ziway by Lowndes (1932). Cyprideis is a salt-tolerant genus, but the specimens present in this sample are noded, suggesting a freshwater environment, since species living in saline waters are smooth-shelled (R.H. Bate, in litt. 29/3/76).

b. The Boloko section (0738-B2)

This shallow section is of interest because it provides clear confirmation of the occurrence of two major Holocene deep-lake phases. It is located on the east bank of the Bulbula, opposite the main stock watering-point on the south side of Adamitulu (Fig. 5.9; Plate 39). The stratigraphy is shown in Figure 6.11.

At the base we find slightly calcareous, lacustrine silts and sands containing thin shell bands. The lower shell layer has been dated at 9270 ± 85 BP (Q-1358). These are overlain by 3.7 m of very poorly sorted, colluvial deposits derived in part from reworking of air-fall pumice. The uppermost sediments are calcareous, lacustrine silts with a single rich shell bed at 8.9 m. This has yielded a radiocarbon date of 5940 ± 70 BP (Q-1359). The robust benthic mollusca (*Bellamya* and *Melanoides*) in this concentration could hardly contrast more strongly with the frail specimens found in the lower and middle shell beds, which consist primarily of small bivalves (*Pisidium* spp.) and pulmonates, notably *Biomphalaria barthi*. Besides *Pisidium moitesserianum*, a palaearctic species which is widely distributed in the lake deposits, the two lower Holocene shell beds contain two additional palaearctic taxa - *P. milium* and *P. nitidum* - associated with the tropical *P. kenianum* Preston and *P. streetae* sp.n. (J.G.J. Kuiper, *in litt*. 28/3/76).

This is the first find of *P. milium* and *nitidum* in eastern Africa. They both have interesting distributions. *P. milium* is a north European species now found living in association with *P. subtruncatum* in Algeria but nowhere else in Africa (Kuiper, 1966). It occurs subfossil in deposits of early Holocene age in Tibesti (Sparks and Grove, 1961; Böttcher *et al.*, 1972), which were laid down at a time when cooler conditions were prevailing in the Chad Basin (Servant *et al.*, 1976). *P.* cf. *nitidum* has so far been identified only in the Palaeolithic and Neolithic lake beds of the Faiyum in Egypt (Gardner, 1932; Kuiper, 1966).

An ostracod, *Cyprideis* sp., was found in the upper and middle shell beds. The tests are noded, suggesting a freshwater species (R.H. Bate, *in litt.* 29/3/76).

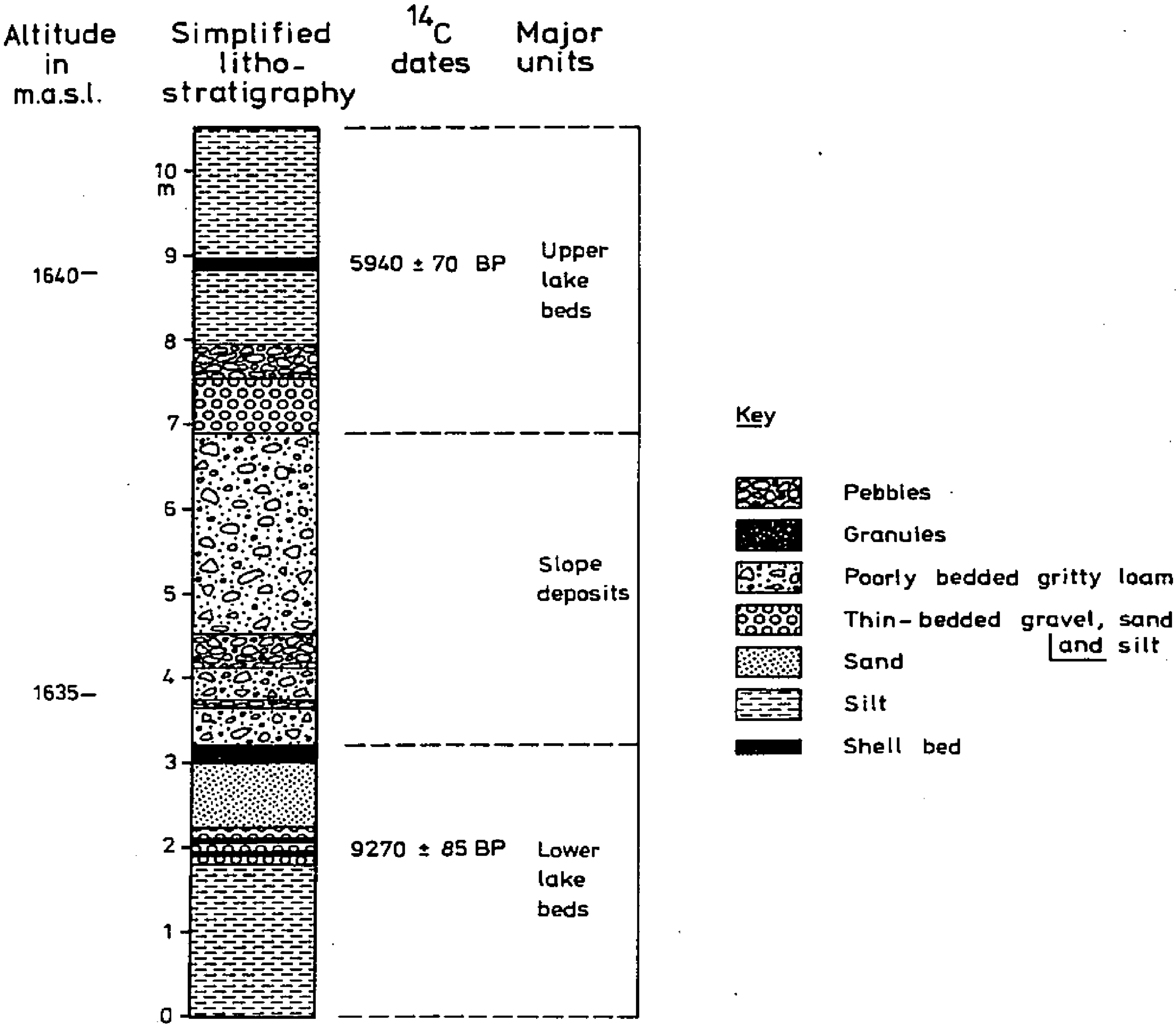


Fig. 6.11: Stratigraphy of the Boloko section (0738-B2).

This section confirms that a major regression of the lake took place between 9270 ± 85 and 5940 ± 70 BP. The lower and upper lake beds can be correlated with members Deka Wede 3b and 4 in the Bulbula type section. The early and mid-Holocene faunas both indicate freshwater conditions, but the absence of large benthic species from the lower lake beds is remarkable, and may reflect slightly different ecological conditions (chapters 7D IIId and 8B).

c. The Gerbi section (0738-B26)

This very shallow exposure, just beside the outlet from Lake Ziway, provided the first conclusive proof of high lake levels prior to 20,000 BP. Laury and Albritton (1975) dated Corbicula shells from the "older basinal sediments" at 26,780 ± 440 BP (SMU-61). Some confusion surrounds this site, which appears to be incorrectly located on their map (ibid., fig. 3). The exposure referred to is almost certainly the one at DU 707731. Here, 3.15 m of light brownish-grey, well-sorted, ostracod-bearing, pumice sands contain abundant shells. Apart from the odd Bulinus "truncatus", these are almost exclusively very robust valves of Corbicula sp. It is possible that the shells were reworked by the Holocene lake, because they show signs of having been torn loose from a weakly $CaCO_3$-cemented shell bed.

Madame Delibrias has kindly attempted a second ^{14}C date on a large sample of Corbicula collected in 1976. She cleaned the shells very carefully and removed the outer fraction with acid. The result was 28,200 ± 1000 BP (GIF-4010). Obviously, shell dates > 20,000 BP have to be treated with caution (Thurber, 1972), but the age obtained falls well short of the 2σ dating limit at Gif-sur Yvette (40,000 BP). In view of the robustness of the valves, and the lack of any recrystallization (Appendix 2c), a finite age for the shells will be accepted. We can deduce that lake level reached at least 1639 m at or before 28,000 years ago. This highstand may conceivably correspond to Bulbula I.

d. Summary

The sediments described in section V were laid down on the southern margins of the North Basin, at altitudes of 1630-1661 m. They therefore only provide evidence for the lake-level fluctuations which were large enough to unite Ziway to the three southern lakes. A single, Lake Pleistocene highstand (tentatively correlated with Bulbula I), and two distinct Holocene highstands (equivalent to Bulbula IIIB and IV), are recorded by sections 0738-B26 and B2. In the Kertefa embayment (sections B16 to B19), the Holocene is represented by thick marginal diatomites or diatomaceous silts (the "Kertefa" facies) which outcrop between 1636 and 1661 m. No clear stratigraphic distinction between the two Holocene deep-lake phases can be made in this area.

6D THE AJEWA FORMATION

I. Introduction

All the sections described so far have been typical of the Bulbula Plain, where the sediments of the Ziway-Shala palaeolakes are draped over a gently sloping, largely depositional topography. Around Lake Shala, however, the lake beds have been preserved in sheltered pockets around the steep and rocky inner walls of the caldera. Correlation from bay to bay is often difficult. But this steepness has certain advantages, since the Shala sections preserve a detailed record of the fluctuations of the lake from its highest levels to modern base level. Lowstands are often represented by unconformities which can be traced for hundreds of metres down-section. Furthermore, the Ajewa formation is rich in datable materials, including shells, limestone, fish bones and charcoal. A total of 11 radiometric ages have been obtained to date.

The type section, 0738-D2, is like the Deka Wede section, relatively complete, but very much harder to interpret. Its stratigraphic record contains several notable ambiguities, which have been resolved in part through comparisons with the adjacent gully 0738-D13; although the apparent lack of lower Holocene lake deposits has not yet been satisfactorily explained. Fine exposures of Late Quaternary sediments also occur in the Deka Halela and Rukesa embayments (Plate 35) but have not yet been dated.

II. The Ajewa type section: 0738-D2

a. General description (Figs. 6.12a, 6.13)

Even at the outset, the Ajewa section posed a considerable stratigraphic challenge, since the sediments are plastered over the irregular slopes of the inner caldera wall and are never more than 9.7 m thick. They outcrop along 3.6 km of natural section, created by a winding gully which was initiated before 1956 (Fig. 6.12a). This is completely dry for most of the year. During the rains, water runs rapidly off the steep and rocky flanks of the Galé horst (1714 m) and descends to the lake over fourteen principal waterfalls where more resistant beds punctuate the sediments. Some basic survey data for this section are given in Table 6.5.

The 1974-75 survey was based on a levelling line tied in to the elevation of the lake surface on 14/10/74. This is referred to as Shala datum (S.D.) (Table 6.5). Profile elevations were interpolated between bench marks set up along the course of the gully. Above 3100 m, the valley becomes choked with shrubs and creepers which make surveying difficult. Because lateral tracing of contacts was hindered by the vegetation, the stratigraphy of this stretch could not be checked as thoroughly as further downstream, but the broad outline is believed to be reliable.

As a first step in correlation, the section has been divided up into four stretches within which internal stratigraphic relationships can be established. These are as follows:-

Downstream (D)	0 - 970 m
Downstream Middle (DM)	970 - 1750 m
Upstream Middle (UM)	1750 - 3260 m
Upstream (U)	3260 - 3637 m

Correlations between these four subdivisions pose much greater problems, because of facies changes or gaps in the succession. An attempt has been made to combine all the evidence in Table 6.8. This is the most complete sequence which can be reconstructed on the basis of present data.

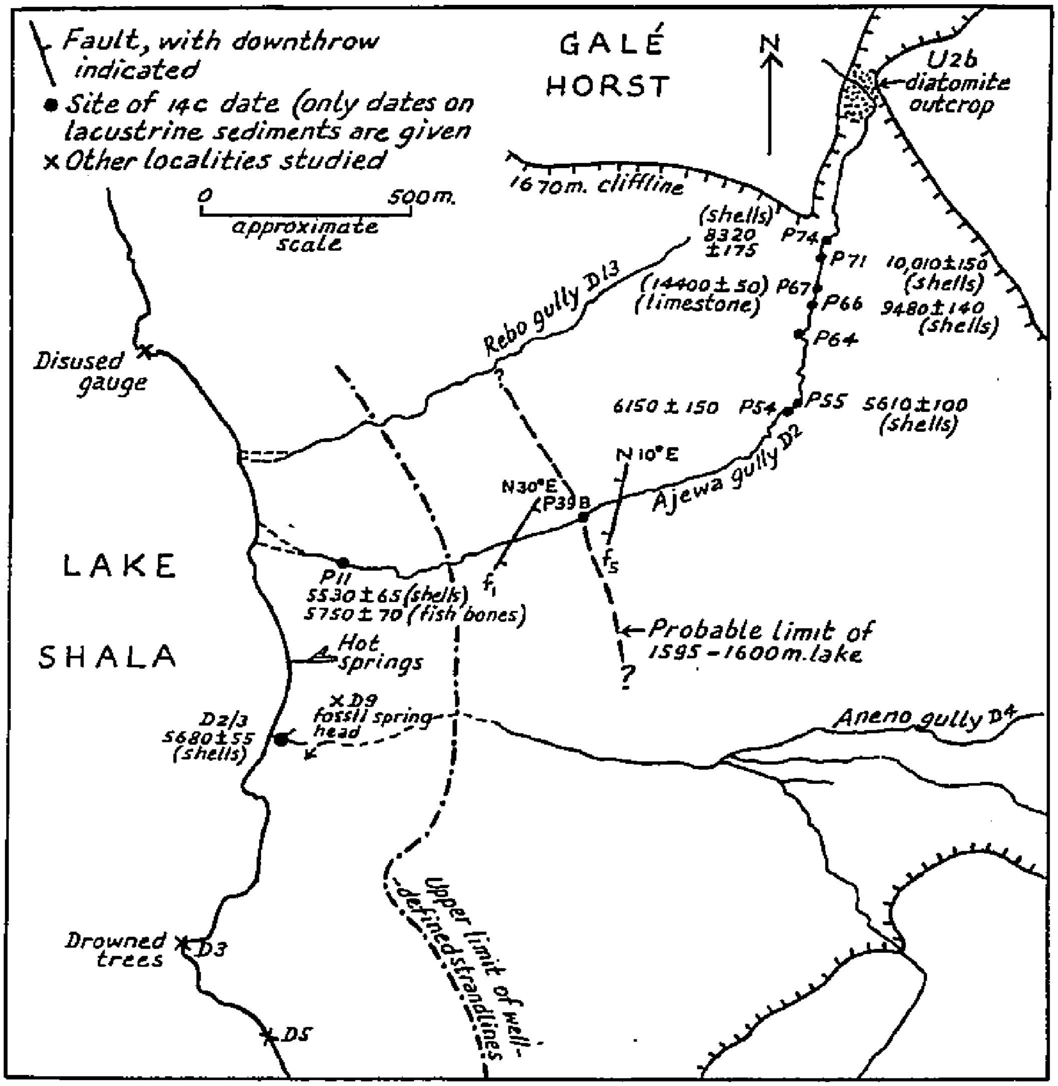


Fig. 6.12a: Map of the northeast embayment of Lake Shala, showing locations of ^{14}C dates.

Fig. 6.12b (overleaf): Known distribution of the Roricha fish bed around the shores of Lake Shala.

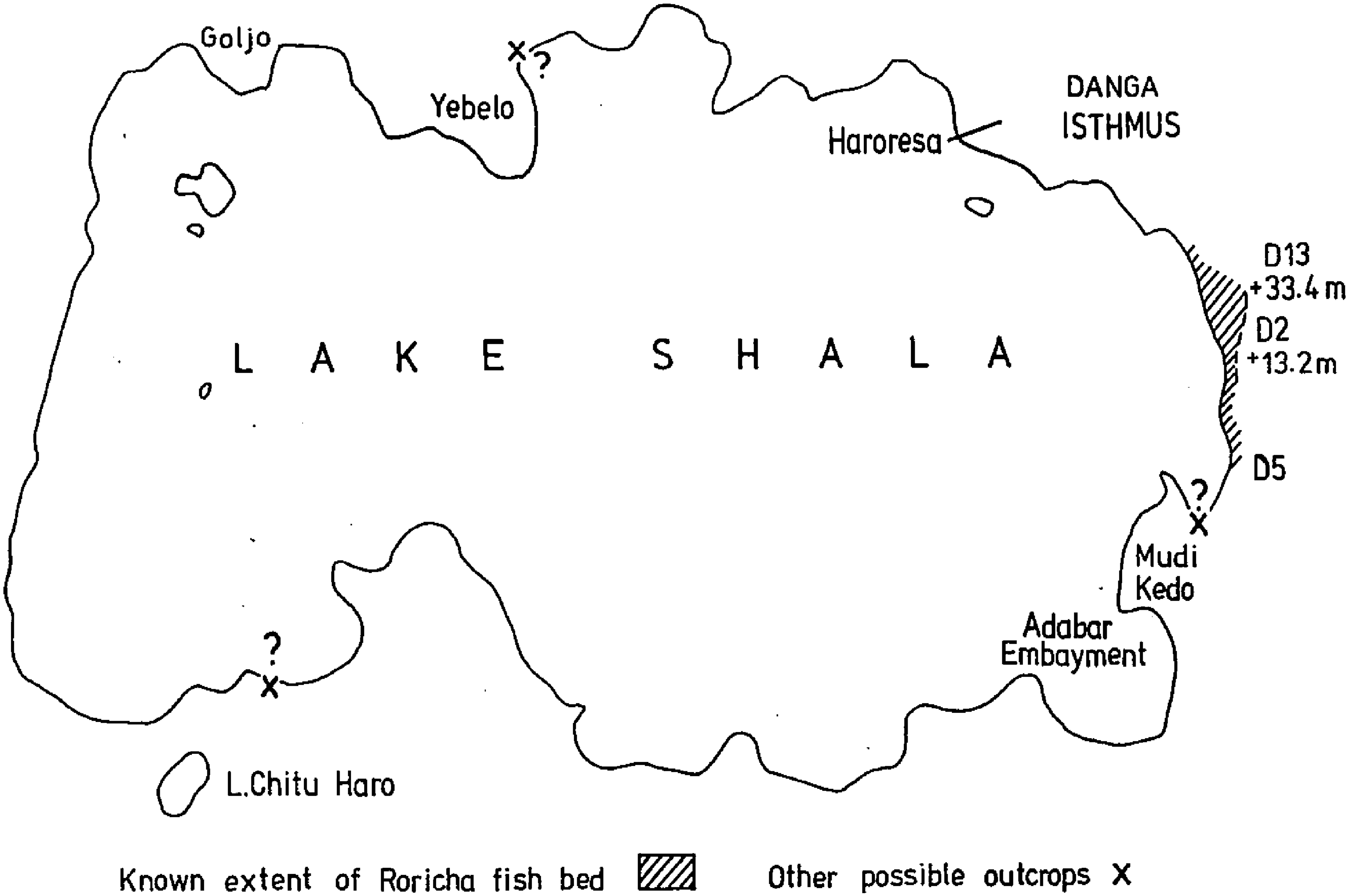

Goljo
Yebelo
DANGA
ISTHMUS
Haroresa
D13
+33.4 m
D2
+13.2 m
D5
LAKE SHALA
Mudi
Kedo
Adabar
Embayment
L.Chitu Haro
Known extent of Roricha fish bed
Other possible outcrops X

Table 6.5: Summary of survey data from the Ajewa section, (0738-D2), Northeast Shala

Thalweg length (to foot of Galé escarpment)	3637 m
Altitude range (1558 m to 1677.3 m)	119.3 m
Overall slope of thalweg	3.28% (1:30)
Number of major waterfalls	14
Highest waterfall	4.2 m
Maximum depth of exposure (P87)	8.9 m
Maximum recorded thickness of deposits (P67)	9.7 m
Maximum local dip of sediments (unit DM1)	19.5°
Number of faults cutting right through sediments	5
Details of largest faults:	
f_1 cuts bedrock - dip	60°
- strike	N 30° E
- throw	1.45 m (down to SE)
f_5 cuts sediments - dip	vertical
- strike	N 10° E
- throw	$>$ 5.3 m, $<$ 10.7 m } down to W
Number of radiocarbon dates	10
Survey datum (Lake Shala level of 14/10/74)	1558.0 m
Number of bench marks along levelling line	44
Number of measured profiles	96
Average spacing of profiles	37.9 m

b. Stratigraphic units

In section 0738-D2, unlike those discussed previously, all the main lithostratigraphic units, lacustrine or otherwise, have been numbered. This is necessitated by the complexity of the stratigraphy. The units defined informally in Table 6.6 are lenticular sedimentary bodies which can be traced laterally for considerable distances; which have distinctive sedimentological characteristics; and which are frequently bounded above and below by erosional contacts or by palaeosols. I recognize that these boundaries are time-transgressive. Some units such as U2 and DM3 have been further subdivided at levels of significant facies change.

Each part of the gully will be described individually before the complete sequence is reconstructed:

D: 0 - 970 m (Table 6.6a)

The sediments in this stretch, close to the lake, rest on a knobbly substratum of caldera volcanics. The oldest rock exposed is a green pumice lapilli-stone which outcrops at the road crossing. This bed has apparently been cemented by lateral movement of SiO_2-rich water towards the springs, as blocks of indurated diatomite still adhere to its upper surface at P11. Very similar, nonsilicified pumice can be found at P24-26. Between 19/3/75 and 5/1/76, erosion revealed 0.3 m of finely laminated diatomite (D1) overlying the lapilli-stone at P9-10.

The upper surface of unit D1 is highly irregular: at P9-11 it is overlain by greenish gravels and sands (unit D2), which occupy a channel cut into the diatomite (Plate 64). The large clasts are angular to subrounded. They consist predominantly of ignimbrite, pumice and lithified diatomite. The sedimentary structures present imply a fluvial origin (Plate 65) (Reineck and Singh, 1973; Picard and High, 1973, pp. 171-175). According to Picard and High, avalanche-front cross-bedding is indicative of high-velocity currents and relatively deep flow. At the top of the gravels, an organic sand with crumb structure suggests limited soil development (Plate 65).

Unit D3 consists of up to 2.6 m of deep-water lacustrine silts which

Table 6.6: Summary of stratigraphy, section 0738–D2

Table 6.6a: Downstream reach (D): 0 – 970 m

Unit	Description	Inferred Origin	Maximum thickness (m)
D6	Greyish (2.5Y – 5Y 5/1 to 6/3), ostracod-bearing pumice sands and silts with microdelta cross-stratification	Gilbert-type delta	2.8
D5	Bedded gravelly sand	Regressive, beach	0.9
D4	Pale brown (2.5Y 6/2 to 6/3), well-sorted, plane-bedded or ripple-marked fine to medium sands and silts with fish bones and ostracods	Shallow-water lacustrine	2.5
D3	Pale brown (2.5Y 6/3), well-sorted, plane-laminated silts with occasional ostracods and a marker horizon of articulated fish skeletons and shells (the Roricha fish bed) $\leqslant$ 0.37 m above base.	Deep-water lacustrine (temporary drop in level recorded by fish bed?)	2.6
D2	"Greenish" (5Y – 2.5Y 5/2 to 5/1) sand and imbricated gravel with channel and avalanche-front cross-stratification. Contains fragments of diatomite D1. Fines upwards into gravelly organic sand.	Alluvial (channel fill)	2.25
D1	Finely laminated, "greenish" (7.5Y 8/1) diatomite	Deep-water lacustrine	0.3
	Shower-bedded, "green", air-fall pumice lapilli-stone	Caldera volcanics	

KEY:

Erosional unconformity

Soil

Table 6.6b: Downstream Middle reach (DM): 970 - 1750 m

Unit	Description	Inferred origin	Maximum thickness (m)
DM6	Cross-bedded pumice pebble-gravel or plane-bedded sugar sand	Beach	2.0
DM5	Pale brown (10YR - 2.5Y 6/2 to 6/3), slightly sandy, gritty silt with pumice pebbles	Alluvial?	1.3
DM4	Pale brown (10YR - 2.5Y 5/4 to 7/2), gritty loams with calcrete layers. Contains derived (?)freshwater mollusca and a mixture of rolled and unrolled artifacts	Colluvial/ alluvial ?	5.2
DM3b	Pale yellow (2.5Y 8/3), plane-laminated, diatomaceous mudstone overlying grey pumice gravel with diatomite clasts and white diatomite	Lacustrine (varying depths) Brief regression marked by gravel	2.75
DM3a	Light grey (2.5Y - 5Y 5/1 to 7/2), sands and pebbles with microdelta cross-stratification	Gilbert-type delta	2.3
DM2	"Greenish-grey" (5Y - 7.5Y 6/3 to 7/2), gritty silts and sands with small calcareous nodules	Colluvial?	1.0
DM1	Finely laminated, "greenish" (7.5Y 8/1) diatomite	Deepwater lacustrine	5.9 (est.)
	Shower-bedded, "green", air-fall pumice lapilli	Caldera volcanics	1.3
	"Green", glassy ignimbrite flow		

Table 6.6c: Upstream Middle reach (UM): 1750 - 3260 m

Unit	Description	Inferred origin	Maximum thickness (m
UM5	Yellow-brown (10YR 5/3 to 5/4), gritty, sandy loam with land snails and occasional reworked freshwater shells	Colluvial	2.2
UM4	Pale brown (10YR 6/2 to 6/3), well-sorted, structureless silts and fine sands with basal concentrations of shells, ostracods and rolled fish bones	Deep-water lacustrine (temporary drop in level recorded by shell bed?)	5.0
UM3	Greyish-brown (10YR - 2.5 Y 6/2 to 6/3), well-sorted silts with vertical root (?) holes, diffuse branching carbonate concentrations, scattered fish bones and some freshwater shells; overlying gritty brown sand and pebble gravel	Lake-margin swamp. ↑ Alluvial (ephemeral channels)	2.5
UM2	Brownish (10YR 5/2 to 6/4), gritty loam with calcrete layers	Colluvial/ alluvial?	4.0
UM1	Tufa-cemented pebble- and cobble-gravel with shells, overlying reworked, very pale brown (10YR 8/3) diatomaceous silts	Alluvial (ephemeral channels). Lacustrine	3.4
	"Greenish" welded tuff	Caldera volcanics	

Table 6.6d: Upstream reach (U): 3260 – 3637 m

Unit	Description	Inferred origin	Maximum thickness (m)
U3	Yellow-brown (10YR 5/3 to 5/4), gritty sandy loam with land snails and occasional reworked freshwater shells	Colluvial	1.4
U2c	Well-rounded, grey pumice pebbles	Beach	1.1
U2b	Fluffy, white (2.5Y > 8/1) diatomite with root casts	Lacustrine (gently shelving marginal embayment)	4.2 (est.)
U2a	Grey (10YR 4/2 to 5/2), well-sorted, plane-bedded silts and sands with lenses of waterworn pumice pebbles	Shallow-water lacustrine	1.4
U1	Yellow-brown, gritty, sandy loam, calcreted in part (not well exposed)	Colluvial	6.0 (est.)
	Ignimbrite, pumice or agglomerate	Caldera volcanics	

rest conformably on the gravels below (Plate 64). These are best preserved at P7-11. They closely resemble the lacustrine Galana Boi Formation at Lake Turkana (Vondra and Bowen, 1978). The silts contain 0.3 - 8.0% organic carbon and 3 - 13% $CaCO_3$. The organic content decreases steadily towards the surface, presumably due to oxidation. In places rather diffuse diatomaceous laminations are present. When wet, the silts have a buttery consistency, but on drying, distinctive prismatic structures about 10 - 30 cm in diameter develop.

The Roricha fish bed

About 21 - 27 cm above the base of unit D3, there is a thin, water-laid ash. A shelly marker horizon 10 cm above the ash has so far yielded 21 intact fish skeletons and many fragments (Plate 66). This has been dated at 5750 ± 70 BP (Q-1361) on fish-bone collagen, and 5530 ± 65 BP (Q-1369) on mixed shell, at P11. A third measurement of 5680 ± 55 BP (Q-1557) was obtained nearby at site 0738-D2-3 (Fig.6.12), where the shells rest almost directly on bedrock. Five more complete fish skeletons were recovered at the same stratigraphic level further south in the Roricha embayment, where the shell band outcrops on the shore (site 0738-D5). From now on, it will be called the Roricha fish bed.

The large fish species identified are as follows:

Tilapia cf. *nilotica*
Barbus cf. *intermedius*
Barbus sp.
Synodontis sp. (catfish)
otolith of *Lates* sp.? (Nile perch) (see Appendix 5).

In addition, we found numerous serrated dorsal *Barbus* spines which are quite unlike the spine possessed by any *Barbus* of that size living in the region today. Their identify remains elusive. The closest match appears to be with one of the "European" groups, *B. callensis*, which is now found in Morocco (K.E. Banister in Grove *et. al.*, 1975, p. 196). The fish skeletons were buried by a layer of reverse-graded, water-worn pumice about 2 cm thick, which contains small, fragile mollusca at the base. These are relatively sparse compared with many shell concentrations in the basin. The fauna consists mainly of Melanoides, Bulinus, Valvata and Pisidium spp. This assemblage is dominated by aquatic species which live on weed, or in the bottom muds under very low energy conditions. But a few individuals of the land snail Cerastua sp. were present (B. Verdcourt, in litt. 1/12/75). This suggests reworking of slope materials by stream or wave action.

What caused the mass mortality and burial of the fish? Various hypotheses can be suggested:

1) Seismic activity (Parker, 1971).
2) Volcanic discharge of toxic substances into the lake, for example a sudden fall of pumice (Wilcox, 1959; Beadle, 1974, pp. 178-180).
3) Deoxygenation produced by sudden overturn of the lake waters (Bishop, 1969; Beadle, 1974, ch. 6).
4) A rapid fall in lake level, leading to the deaths of fish trapped in shallow pools (Beadle, 1974, ch. 19).
5) A single catastrophic flood (cf. McLeroy and Anderson, 1966).

It is clear from the ^{14}C dates that we are dealing with a single and widespread event. The Roricha fish bed outcrops all along the eastern shore of Shala from 0738-D5 to D11, and is probably equivalent to the prominent shell layer in D12. A very similar bed can be found just east of Yebelo (Fig. 6.12b). On the northeast shore, it ranges in height from 33.4 m.S.D. to below 1975 lake level. However, intact fish are restricted to a limited stretch from D2 to D5, up to 14 m.S.D. Elsewhere, disarticulated fish bones occur scattered throughout a layer of normally graded, shelly, pumice grit and sand up to 30 cm thick rather than beneath it.

The evidence relating to the origin of the Roricha fish bed can be summarized as follows:-

1) Seismic activity.

Although earth tremors are occasionally recorded in this part of the Rift, this hypothesis is very difficult to test. It is necessary to assume that the large quantities of pumice and mollusca in units D3 and UM4 (see discussion below) were also transported from the lake margins by earthquake-generated turbidity flows.

2) Volcanic activity

a) The pumice layer in unit D3 is reverse-graded. This is typical of coarse pumice deposits which have settled in water, because of the increase in buoyancy with clast size, in contrast to sand- and silt-sized ash, which does not float (McLeroy and Anderson, 1966).

b) Shala lies very close to the Chabbi volcanic centre, which has undoubtedly been active during the Holocene.

c) However, no shells or fish remains are associated with the thick layer of water-laid pumice and ash in unit DM3b.

3) Breakdown of stratification

a) Many cases of massive fish mortality through asphyxiation have been reported during overturn of deep, stratified lakes in Africa and Indonesia, when stagnant H_2S-charged water is brought to the surface (Beadle, 1974, ch. 6; Eccles, 1974). Similar disasters also occur in the shallower waters of productive lakes like George and Chilwa if the highly reducing and oxygen-consuming mud is stirred to the surface by the wind (Beadle, 1974, pp. 69-70, 280).

b) At present, the bottom waters of Lake Shala are fairly low in oxygen (1.0 - 2.3 mg/l). Although not completely anoxic, this is close to the lower limit for fish in inland tropical waters, which averages about 2 mg/l (Balon and Coche, 1974). The muds are anaerobic and give off toxic H_2S if disturbed (Vatova, 1941).

c) Mixing with oxygen-depleted bottom water could be triggered by very windy conditions, or by an increase in surface water density resulting from cooling or evaporative concentration (Eccles, 1974; Hutchinson, 1975, ch. 7).

d) A rise in the level of Lake Shala is likely to be initiated by overflow of much fresher water from Abiyata. This influx might induce stable stratification, especially if the lake was very saline to begin with (Wilson, 1967; Hutchinson, 1975, p. 481; Assaf and Nissenbaum, 1977). At present the specific gravity of the surface water is around 1.013, compared with 0.9970 for pure water at 25°C (Omer-Cooper, 1930; Beadle, 1974, fig. 6.6). Permanent stratification would allow more rapid colonization of the surface waters by fish and mollusca than if the entire water mass had to be diluted. Drier conditions, or the cessation of the freshwater inflow, could cause overturn and result in the sudden death of the fauna by asphyxiation (Hecky, 1978).

4) Desiccation

a) Fish kills have often been reported during the drying-up of fluctuating brackish lakes such as Chew Bahir (Grove and Goudie, 1971b) and Chilwa (Beadle, 1974, ch. 19). Mortality appears to result from a complex interaction of elevated levels of alkalinity and decreased oxygen concentrations (McLachlan et al., 1972). A similar event at Lake Shala would require a very rapid fall in level and the presence of pools in which fish could be trapped. This possibility cannot be ruled out, given the rocky topography.

b) Dr. Francoise Gasse has examined a sample of the diatoms from the silts immediately underlying the subfossil fish at site 0738-D5 (Appendix 4). The flora is largely littoral and indicates quite brackish conditions. Of the species present, Rhopalodia gibberula, Anomoeoneis sphaerophora and their varieties are quite common in Abiyata today (Gasse, 1975). It looks as if conditions lay at the saline and alkaline end of the range for aquatic macrophytes, mollusca and large freshwater fishes.

c) A drop in lake level would fit in with the observation that the thin pumice bed contains transported terrestrial shells. The pumice and glass shards lack the blue-grey vitreous lustre typical of undisturbed waterlaid tephra, but instead resemble the dull, iron-stained pumice gravel associated with shell concentrations in many of the Bulbula Plain sections (0738-B16 to B19, for example). Since the Shala sediments are rich in pumice, a thin layer could easily be produced by wave reworking or slumping into deep water during a recession (Donovan and Archer, 1975). Alternatively, it was noted in Lake Chilwa that a progressive reduction in water level from 1963 to 1967 caused the molluscs to follow the receding water and leave their normal littoral habitats. This resulted in a redistribution and accumulation of shells at the lake centre (A.J. McLachlan, in litt. 25/7/78).

d) However, a simple consideration of evaporation rates (chapter 2B) suggests that even if its inflows were completely cut off due to prolonged drought, the level of Lake Shala would be unlikely to drop faster than 2 m/y without volcanic or tectonic intervention. It is therefore difficult to image how the fish could be effectively trapped.

5) Catastrophic flooding

a) McLeroy and Anderson (1966) have found reverse-graded pumice bands averaging 1.5 cm thick in laminated Oligocene diatomites in the U.S.A. These layers contain wood fragments. The authors suggest that the pumice was carried in by winter or spring floods and floated out onto the lake. If the Roricha fish bed was formed as a result of a single, torrential storm, than a) the fish must have been killed at the same time, perhaps by the increase in turbidity, and b) the mollusca must have been transported out from the shore zone by the flood waters. This would account for the presence of Cerastua.

b) Against this hypothesis we have the widespread distribution of the fish bed, the very limited catchment of the tributary gullies, and the wave-sorted appearance of the pumice gravel above 14 m S.D. It also seems improbable that only one such incident occurred during the deposition of 2.5 m of silts.

Reviewing all the evidence given above, an abrupt drop in lake level seems the best explanation for the existence of the Roricha fish bed. It appears most probable that the fish were killed by deoxygenation resulting from a breakdown in stratification. According to this model, Lake Shala initially rose to at least 1651 m (from Q-1101), uniting the three southern lakes. A temporary climatic reversal may then have caused an abrupt drop in level, leading to reworking of the marginal deposits, and destroying the density stratification in the lake. This hypothesis derives further support from the evidence relating to unit UM4.

Unit D4. The upper surface of unit D3 is strongly channelled, testifying to a fall in lake level to 1562 m or below (Plate 67). It is overlain at P12-17 by well-sorted, shallow-water lacustrine silts and sands displaying ripple marks, small-scale cross-stratification and lenses of coarse sand or small pebbles. The sediments contain 7 - 14% $CaCO_3$, mainly in the form of ostracod tests. Melanoides and Bulinus shells are very scarce and may be reworked.

Abundant fish bones were also found in this unit. They occurred in concentrations, but were so crumbly that it was impossible to tell whether they had been deposited as whole fish or merely as the food refuse of fish-eating birds, like the clusters of fish bones found in modern fine-grained sediments along the shores of Abiyata. Whichever was the case, it is likely that large fish were living in the lake at the time. However only cichlids (probably the salt-tolerant Tilapia) are represented (Appendix 5).

Unit D5 (plane-bedded beach gravel and sand) appears to represent the regression of unit D4.

Unit D6 consists of well-sorted, deltaic sands and silts which are only found very close to the lake (Plate 68). They contain 5 – 10% $CaCO_3$ in the form of ostracod tests. Similar small Gilbert-type deltas (Gilbert, 1885) can be seen at the mouths of many of the gullies along this stretch of shore, below 5.8 m.S.D. (chapter 5BIII). Conspicuous convolute bedding probably resulted from localized liquefaction of the sands on the unstable delta fronts as they advanced into the lake (Born, 1972; Reineck and Singh, 1973, pp. 78–79).

DM: 970 – 1750 m (Table 6.6b)

Unit DM1. In this stretch the oldest lake sediment is again a rhythmically laminated greenish diatomite (DM1) (Plate 69). Between P23 and P24 it can be seen to rest on air-fall pumice lapilli which are draped over the irregular cooling surface of a glassy ignimbrite flow (Plate 70). This flow has proved difficult to fit into the stratigraphy of the caldera (P.A. Mohr, in litt. 18/1/78) – my own impression is that it has flowed down the inner walls and is therefore one of the youngest volcanic units. Dr. Frank Fitch, of Birkbeck College, kindly attempted to date a sample by the fission-track method. No tracks were found in any of the large number of zircons prepared, which suggests that the ignimbrite is exceptionally young (< 100,000 years). An attempt is being made to date the flow by $^{40}Ar/^{39}Ar$ analysis, but no results are available as yet.

The green pumice may record the final, explosive foundering of the caldera floor (cf. Cole, 1969; Laury and Albritton, 1975). Its upper 35 cm have been reworked by running water - perhaps as the lake filled up the newly created basin - but there is no sign of erosion to indicate that significant time elapsed before the start of lacustrine deposition. The overlying diatomite (DM1) is draped conformably over pumice slopes of up to 19.5^{o}.

The diatomite itself consists of a minimum of 4800 - 5000 pairs of green and white laminae, with a mean thickness of 0.87 - 0.91 mm per couplet. However the top is always an erosional unconformity. Bulk samples contain from zero to 71% low-Mg calcite, which is partly segregated into small, spherical grey nodules.

Could the rhythmites represent seasonal layering, as in glacial varves? On the face of it, this seems very probable, for the following reasons:

1) The couplets are of the right order of thickness to be annual layers (Table 7.2).

2) The Rift floor climate is highly seasonal, with two wet/cold and two dry/warm periods during the year. It is not certain, however, that the small rains would manifest themselves limnologically as a distinct event from the main rains, and indeed the bimodality of the rainfall pattern may have been less pronounced in the past (Richardson and Richardson, 1972).

3) Similar rhythmites (alternating grey, translucent and white, powdery laminae) are found in the Abhé Core, where they date from the Abhé II highstand (Gasse, 1975, pp. 140-146). Dr. Gasse found that the grey layers were dominated by the cosmopolitan/temperate diatom Stephanodiscus hantzschii var. pusilla, and the white ones by the tropical endemic Nitzschia aequalis, suggesting deposition during cold and warm seasons respectively.

Rhythmites are normally deep-water features which survive only where the normal scavenging and burrowing fauna of a lake bottom is excluded and prevented from reworking the surface sediment and where the layering is not disturbed by current action (Turner, 1970). Their preservation is therefore favoured by permanent or seasonal stratification of the water mass (McLachlan and McLachlan, 1971; Ludlam, 1976), which causes oxygen depletion of the bottom water. An exception to this general rule is Lake Turkana, where clastic laminations are preserved despite oxygenated bottom-water conditions, for reasons which remain obscure (Yuretich, 1976).

Shala is a very deep lake which may have undergone permanent or near-permanent stratification in the past. This situation is most likely to have occurred during periods of greater dilution following prolonged dry conditions (Wilson, 1967; Assaf and Nissenbaum, 1977), especially if the hot-spring inflow in this corner of the lake was substantially less dense than the lake waters, as at present.

We can reconstruct little of the palaeogeography of this remote episode. The upper limit of the diatomite outcrop is 1602 m in D2/44 and 1595 m in the adjacent section D13. At these altitudes the bed is still laminated, suggesting an overlying water depth of at least 50 - 100 m (estimated from Beadle, 1974, ch. 6). However the outcrop is faulted in at least three places (Figs. 6.12, 6.13) and no correlative marginal deposits can be identified. Although its detrital content does increase slightly towards D13, the diatomite implies conditions under which very little clastic sediment was supplied to the lake: a very different situation from today. Until the diatom stratigraphy of the Upper Pleistocene levels in the Abiyata Core is completed, our knowledge of this highstand and its climatic implications will remain fragmentary.

Unit DM2. The diatomite is unconformably overlain in P29-33 by a curious gritty silt or sand-sized material. This contains small calcareous nodules, and has pumice pebbles and granules scattered throughout. Bedding is apparent only in places. No mollusca of any kind were found. Similar sediment infills a deep channel incised into the diatomite at P27. These gritty deposits remind one very strongly of the younger units DM4 and UM2, and so discussion of their origin will be deferred until pp. 275-78.

Unit DM3. Unit DM3a is a transgressive lacustrine sequence resting unconformably on a very irregular diatomite surface. It begins at the base with sandy diatomite gravel of probable fluvial origin, and then passes up into well-sorted, medium pumice sand (3 - 7% $CaCO_3$) with lenses of rounded pumice pebbles (Plate 71). The sands display large-scale microdelta cross-bedding, often forming two distinct sedimentation units. They recall the deltaic sands of unit D6, but no shells or fish bones are present in this case.

Unit DM3b begins with a pale yellow pumiceous silt, which passes upwards into a bed of very pure, fluffy white diatomite containing up to 12% $CaCO_3$ (Plate 71). This is overlain by a thin layer of pumice gravel with diatomite clasts, followed by a thick bed of pale yellow, noncalcareous, diatomaceous mudstone. This has a characteristic prismatic appearance when weathered (Plate 72), and includes a 2 cm-thick layer of rounded pumice pebbles about 0.8 m above its base. In section 0738-D2, the top of unit DM3b has always been truncated by erosion (Plate 72).

The very pure basal diatomite in unit DM3b is floristically similar to the marginal freshwater diatomites of Holocene age, such as unit U2b (chapter 7DIIa and Appendix 4). It probably represents an initial expansion of the lake over the P44 site, followed by a minor recession which entrained deposition of the pumice gravel. The overlying pale yellow mudstone represents the maximum transgression. It is cut by all five faults which intersect the gully and is displaced as much as 5 - 10 m by f_5. Although it only outcrops between 1584 and 1611 m, its homogeneous, fine-grained character indicates that the shoreline lay much higher. This is borne out by the abundance of planktonic diatoms, especially endemic *Melosira* spp. (Appendix 4).

The mudstone is also exposed in the adjacent section 0738-D13, where the marker band of pumice pebbles increases in thickness to a maximum of 0.65 m. Here it is reverse-graded, passing upwards from grey ash to subrounded to well-rounded, medium to large pumice pebbles, and clearly represents a single tephra fall into the lake. The increase in thickness of the pumice from D2 to D13 may simply reflect wind drifting of a pumice raft against the NE shore of the lake.

In 0738-D13 a regressive sequence with shells, terminating in a palaeosol, has been preserved at the top of unit DM3b. The mollusc fauna is dominated by pulmonates and pisidia. The latter include *P. moitesserianum*, *P. streetae* sp.n. and *P. margarithae* sp.n. (J.G.J. Kuiper, *in litt.* 23/3/76). The shells are small, fragile, and thinly dispersed through the sediment. Probably this is an *in situ* assemblage which lived on the stems of aquatic plants or in the surrounding mud.

Units DM4 and DM5. At first sight both units appear so heterogeneous as to be almost homogeneous. In pedological terminology, they consist of very poorly sorted, pebbly, gritty silt or sand loams. These are generally pale brown in colour, but range from light grey to yellow-brown. From a distance, individual beds have a massive appearance, averaging 0.25 - 1.0 m in thickness (Plate 73), but at close quarters, thin discontinuous seams of pebbles or cobbles are visible. These consist chiefly of rounded to well-rounded pumice clasts. However lava and pitchstone are locally represented, especially near the base, where the deposits have a more water-sorted appearance. Signs of soil formation occur at several levels, especially at the top of unit DM4, which can be separated from unit DM5 at P39-40 by the presence of a truncated palaeosol, some 70 cm thick. The deposits generally give a slight to violent reaction with dilute HCl. A few levels are quite strongly cemented, for example the lip of the waterfall at P39. In some strata the carbonate is organized into small nodules.

Small scattered shells or shell fragments were found in unit DM4 in P41 and P42. These include *Melanoides*, *Valvata*, pisidia, *Biomphalaria* and smaller planorbids. Unit DM4 also contains rolled and unrolled obsidian artifacts of both MSA and LSA attribution (Appendix 6b). At P39B, unit DM5 contained a lens of large fragments of *Acacia* charcoal. These have been dated at 2510 $\pm$ 100 BP (GIF-4011). Their matrix is a slightly gritty sandy silt which reacts vigorously with HCl. It contains seams of pumice pebbles which pinch out laterally. Slight soil development is indicated by tubular structures (burrows?), and DM5 is generally reminiscent of floodplain deposits (M.A.J. Williams, pers. comm., 1975).

Is unit DM4 also partly alluvial in origin? The lower part shows signs of water sorting like the present gully-floor gravels. Higher up the deposits contain more interstitial fines, although thin stringers of waterworn pumice are still present. The platy or blocky soil-like horizon at the top (P39-40) are very similar to beds in the Upper Member of the Koobi Fora Formation at Lake Turkana which have been interpreted as overbank deposits by

A.K. Behrensmeyer (pers. comm., 1977). The presence of shells at P41-42 is not diagnostic, since all the species present could have lived in small ponds or marshy areas, but could equally well have been reworked from older lacustrine deposits. The association of waterworn (derived) and fresh LSA artifacts would support a colluvial or alluvial origin.

Unit DM4 is therefore tentatively interpreted as a complex of sheetwash deposit, ephemeral-channel gravels, and finer-grained pond or overbank deposits. It is very similar in character to units DM2 (which is even more calcareous) and UM2, which probably have a similar origin.

Unit DM6. A shortlived lacustrine transgression seems to have followed deposition of unit DM5. The evidence takes the form of a patch of plane-bedded gravel and sugar sand at P30-33, and a very localized outcrop of pebble gravel and sand with beach-foreshore cross-bedding (Born, 1972) at P40, too small to be plotted on Figure 6.13. Since the upper part of unit DM5 has been reworked by wave action, the latter must have occurred less than 2510 ± 100 years ago.

UM: 1750 - 3260 m (Table 6.6c)

Unit UM1 is the oldest exposed unit, and is lithologically very variable. It has been dated at one site, P67 (Grove et al., 1975, fig. 6), where it forms an upward-coarsening sequence passing from diatomaceous silts at the base into indurated cobble-gravel. Floristically, the silts are extremely diverse, and the highly fragmented state of the diatom frustules leads Dr. Gasse to suspect reworking (Appendix 4). The overlying conglomeratic limestone consists of waterworn ignimbrite boulders and rounded pumice pebbles, strongly cemented by calcium carbonate and lenses of opaline silica. This limestone layer also contains fairly rich concentrations of freshwater lacustrine shells; the principal taxa being unionids (some articulated), Melanoides, Biomphalaria pfeifferi, Corbicula, Pisidium moitesserianum, Pisidium sp., and abundant small planorbids.

The carbonate/silica matrix of the limestone suggests an algal tufa deposit, especially as this site lies only 2.5 km upstream from the modern lake-shore springs. Travertine and sinter are common around hot springs in the Main Rift; silica deposition is believed to reflect higher underground temperatures than carbonate. Algae are found growing prolifically in spring vents below about 65°C and may play an important role in tufa deposition (United Nations, 1973). However the upper surface of the limestone layer has a smooth laminated appearance reminiscent of many petrocalcic (calcrete) horizons (Goudie, 1973) and has probably been precipitated by more recent ground waters moving laterally over the surface of the plugged horizon below. Thin, laminated tufa coatings can be found further upvalley on large boulders and bedrock outcrops in or below unit UM1 (P68-80B).

A.T. Grove and A.S. Goudie obtained an age of 14,400 ± 750 BP (GaK-3748) on a sample of the inorganic limestone matrix from P67 (Grove et al., 1975). The validity of this date and its implications for the lake-level chronology will be discussed on pp. 288-89.

Further upstream, at P68-70, unit UM1 can be divided into a lower sequence of well-sorted sand, silt and waterworn pumice pebbles, and an upper one of cobble-gravel with tufa overlain by poorly-sorted pebble-gravel of rather mixed lithologies in a gritty silt matrix. The latter contains stromatolite clasts. In terms of environment, UM1 therefore appears to represent the passage from lacustrine conditions to a rocky shoreline with springs, and finally to ephemeral stream channels. The 50 cm-thick bed of rounded pumice pebbles in the lower part is reverse-graded and probably records a tephra fall into standing water.

Unit UM2 strongly resembles DM4, although somewhat more weathered in appearance. The bulk consists of poorly sorted and bedded, gritty, sandy, silt loams with stringers of waterworn pumice pebbles. Low-Mg calcite is commonly diffused throughout the matrix but is sometimes organized into small concretions, or into moderately cemented horizons (19 - 26% $CaCO_3$) which outcrop on the gully floor as waterfalls (P57, P64).

Along parts of the section a truncated palaeosol is present at the top of unit UM2. This is best seen at P63, where a pale brown, gritty sandy silt A horizon with crumb structure, or weakly developed fine angular blocky structure, passes downwards into a more clayey B_k horizon with medium to fine angular blocky structure, and then a weakly cemented C_k horizon. The parent material below is slightly gritty silt with scattered pebbles, but the clayey B_k horizon could be partly sedimentary.

The majority of unit UM2 would seem to be colluvial or alluvial in origin, for the same reasons as DM4. No mollusca of any kind were found. However at P57 well-sorted silts and plane-bedded fine sands with pebble lenses are present at the base of the section. These contain ostracod fragments and occasional diatoms. It is very difficult to tell how well microfossils from lacustrine sediments would survive reworking by surface runoff, but they do provide circumstantial evidence for a Late Pleistocene transgression of the lake to 61 m S.D. A correlation with unit UM1 is probable but cannot be substantiated because of the limited exposure and different facies.

Unit UM3 is a fining-upwards sequence which intervenes at P56–66 between the UM2 soil and the fully lacustrine unit UM4. At P64 it can be seen to grade from gritty sand and pebble-gravel at the base into slightly gritty silts. The uppermost sandy silts are generally structureless, and pale brown to greyish brown. But at P57–58 they show fine, horizontal laminations and contain fish bones and scattered shells (Melanoides, Bulinus, Valvata and small planorbids) but no ostracods. This is a fairly tolerant and not distinctively lacustrine fauna. Commonly the silts are darkened at the top by organic matter, and at P59 a 50 cm-thick dark greyish-brown soil with crumb structure and soft, diffuse carbonate concentrations is present. At P64 this palaeosol displays vertical tube-like structures (root holes or burrows) which have been infilled by lacustrine sediment. Here it also contained a large lens of carbonized Acacia wood, which has been dated at 10,220 ± 140 BP (SUA-500).

The most likely interpretation of unit UM3 is that it represents the lateral and vertical passage from fluvial to swampy, lake-margin conditions (Freytet, 1975). The dark hydromorphic palaeosol recalls the modern soils in marshland and seasonally flooded meadows on the western side of Ziway, which are dark grey, organic loams or clays (Makin et al., 1976; soils U and M). According to the model, the lake rose temporarily to 74 m.S.D. (1632 m) (from P66) so that marginal swampland invaded the central part of UM, and then fell again before deposition of unit UM4.

In contrast, the overlying unit UM4 is unquestionably lacustrine. Between P47 and P72, it consists of well-sorted, structureless silts and fine sands with prolific concentrations of shells, ostracods and rolled fish bones near the base (Fig. 6.13; Appendix 5). However, ostracods are also disseminated throughout the fine fraction, giving $CaCO_3$ contents of up to 26%. At P74-75, where UM4 outcrops round the base of a rock knob, it consists of pebbly, diatomaceous silts to medium sands with microdelta cross-bedding. Here the grain-size distribution closely resembles the deltaic foresets in unit D6.

Unit UM4 is usually capped by a dark brown A horizon buried by younger colluvium (unit UM5). This palaeosol appears to represent the recession of the swampy lake margin across the area, but is less well developed than the one associated with unit UM3.

The age and origin of the shell concentrations in unit UM4 need to be considered in conjunction with the Roricha fish bed. The dating results obtained so far are given in Table 6.7.

Table 6.7: ^{14}C dates from unit UM4, section 0738-D2

Group	Site	Uncorrected date (years BP)	Lab. no.
A	P54	6150 ± 150	SUA-497
	P55 *	5610 ± 100	GaK-3387
B	P66	9480 ± 140	SUA-501
	P67-71 *	9220 ± 190	GaK-3386
	P71	10010 ± 150	SUA-502
	P74 *	8320 ± 175	Q-1114

* Approximate location (from Grove et al., 1975 and Grove and Goudie (unpubl.)
All dates are on Melanoides tuberculata shells.

The ages obtained from UM4 fall into two spatially separated clusters: an upstream, northerly group A, ranging from 8320 to 10,010 BP, and a downstream, southerly group, B, of 5610 and 6150 BP (Figs. 6.12a and 6.13). Several hypotheses can be advanced to explain this rather curious pattern:

1) The stratigraphy shown in Figure 6.13 is seriously in error.

2) Some of the radiocarbon samples were contaminated.

3) Only one highstand occurred, but accumulation of shells was time-transgressive.

4) Mollusca dating from an early Holocene lacustrine highstand were reworked during a catastrophic event around 5700 BP which also gave rise to the Roricha fish bed.

These hypotheses can be evaluated as follows:

1) <u>Stratigraphic error</u>

a) The UM reach was surveyed with great care in 1974 because of the conflicting dates obtained by Grove <u>et al.</u> (1975).

b) The shell bed forms five separate lenses up to 280 m long and 60 cm thick, with a vertical range of as much as 9.45 m (P64-68). In between, there are short gaps where mollusc concentrations were not recorded. The consistent occurrence of shell lenses close to the base, and their characteristic association with a matrix of poorly sorted, pebbly fine sand, overlying a thin, greyish-brown sandy silt with coarse prismatic structure, suggest a common origin, except at P74-75, where the setting is clearly different and reflects shallower conditions.

c) There are no other indications of a disconformity within UM4.

2) <u>Contamination</u>

a) The large discrepancy between groups A and B could most easily be explained by contamination of SUA-497 and GaK-3387 with recent carbon. However X-ray determinations on the shell samples collected in 1974-75 showed that none had undergone measurable recrystallization (Appendix 2c). They were also quite free from encrustation by pedogenic carbonate, which would yield erroneously young ages. Moreover the ostracods in the matrix are beautifully preserved, although they might be expected to succumb first to dissolution.

b) Dr. R. Gillespie, of the Sydney University Dating Laboratory, notes that statistically significant differences exist <u>within</u> each cluster of dates as well as <u>between</u> the two. Although there is no meaningful difference between SUA-501 and GaK-3386, SUA-497 and GaK-3387 are probably, and SUA-501 and -502 significantly, different (<u>in litt</u>. 28/3/77). In group B, Q-1114 is distinct from the other three. Of the remainder, SUA-502 seems the most suspect in the light of the dates from the Abiyata Core, although it is consistent with the age of 10,220 $\pm$ 140 BP obtained from unit UM3; but zero error (contamination by old carbon) cannot be invoked to explain the anomalously old age of a single shell sample. We are therefore left with the possibility of reworking as the most likely explanation for the discrepancies.

3) <u>Time-transgressive deposition</u>

The origin of shell concentrations in general will be discussed in chapter 7DIIId. The pattern of dates in this section does not fit easily into any simple transgression-regression model of lake-level fluctuations.

4) <u>Reworking</u>

a) The model previously adopted for the origin of the Roricha fish bed also fits the evidence from unit UM4. The dates in group A overlap with the three derived from the fish bed itself.

b) According to this hypothesis, the lake rose to at least 1646 m between 10,010 $\pm$ 150 and 8320 $\pm$ 175 BP (phase Ajewa IVA; Table 6.8). Populations of benthic and pulmonate mollusca became established in the inner recesses of the Ajewa embayment. But there is surprisingly little trace of any clastic sediments associated with the Ajewa IVA phase, with the possible exception of the shallow-water deposits in P74. Either material was prevented from accumulating, or else the deposits were reworked during the transgression of the Ajewa IVB lake. It is possible, for example, that strong reactivation of the hot springs during the relatively brief IVA maximum resulted in the flushing of loose sediment down the caldera walls; an explanation which has been advanced for the localized absence of lower Holocene deposits in Lakes Abhé and Kivu (Gasse, 1975; R.E. Hecky, <u>in litt</u>. 16/9/75).

c) Lake level then seems to have fallen to 1561 m or below, since unit D3 rests conformably on fluvial gravels extending down to this elevation. It is conceivable that Q-1114 dates from the early part of this regression, though the possibility that mixing of shells of early and mid-Holocene ages has given a spurious result should be borne in mind.

d) By 6500 ± 120 BP the shoreline again stood at 1651 m in the Mirrga graben (Q-1101). The prismatic sandy silt which underlies the shell concentrations can be attributed to this Ajewa IVB transgression, since it is very similar to the thin mud blankets deposited during the filling of Lakes Chilwa and Kariba (McLachlan et al., 1972; McLachlan, 1974; McLachlan and McLachlan, 1976). These are thought to have been produced by two main processes: the settling of fine suspended silt, and burrowing by expanding populations of mud-dwelling larvae. The lake waters once again freshened sufficiently to allow the colonization of the Shala Basin by large populations of fish and mollusca.

e) A catastrophic drop in level around 5750 ± 70 BP (Q-1361) then caused reworking of the exposed lake floor. (This age on fish-bone protein should be fairly reliable, and corresponds almost exactly to the mean of the four mid-Holocene dates on shell.) The matrix of the shell concentrations in unit UM4 is partly sorted and strongly coarse-skewed (σ_ϕ 1.37, Sk −0.70), testifying to the operations of some transporting or winnowing process, so it seems that the transport of shells into deeper water by wave or turbidity-current action gave rise to a shell pavement overlying the transgressive silt. In the inner recesses of the Ajewa embayment the shell material seems to have been reworked from lower Holocene deposits, whereas in the D and lower UM areas further south it appears to have come from living shells of mid-Holocene age (Fig. 6.12a).

f) By 5200 BP or before the lake had again risen to its outlet level, and the shell concentrations were covered by up to 5 m of silts forming the main bulk of units D3 and UM4.

Unit UM5 can be identified continuously from P64 upstream, where it represents a tongue of colluvium derived from the Galé fault scarp. It is a poorly sorted, massive, gritty silt with angular to subrounded pebbles of ignimbrite and pitchstone scattered throughout. Medium angular blocky structure is usual. Films of clay or white powdery carbonate are common on ped surfaces. Large, well-preserved land snails were found in abundance at P79 and P82, and many smaller species were obtained by sieving the colluvium from P79 (Appendix 3). A fragment of Melanoides tuberculata was also found in this sample, and reworked lake shells are common at P78-80A, but only the terrestrial gastropods are undamaged. Since very little is known about the ecology of Ethiopian land snails, we are not able to make any ecological inferences.

Downstream, at P49-54, unit UM5 is slightly darker in colour and shows signs of water sorting, probably by surface wash. In this part of the gully it does not contain any land snails, although reworked lake shells are abundant.

Where the surface of unit UM5 has not been extensively trampled by the large herds of sheep and goats which graze in this area at the beginning of the dry season, a weakly developed soil can be seen at the top, for example at P64. This shows incipient textural B and C_k horizons, but is generally quite similar to the soils on Ziway terrace IV.

U: 3260 - 3637 m (Table 6.6d)

The stratigraphy of this reach is essentially simple, though poorly exposed. A lenticular body of lake sediments, deposited in a sheltered graben ($\leqslant$ 1675 m) at the base of the Galé fault scarp (Plate 74), is bounded above and below by colluvium (units U1 and U3) closely resembling unit UM5. The lateral relationship between the two reaches is difficult to follow upstream from 3150 to 3380 m, where extensive reworking has taken place. It does however seem very probable that U3, which also contains land snails, is equivalent to UM5.

The lake beds consist of up to 1.4 m of plane-bedded, grey, pumiceous silt with lenses of floated (?) pumice pebbles (unit U2a), conformably overlain by a maximum of about 4.2 m of fluffy diatomite with root casts (unit U2b). The latter contains 1-3% $CaCO_3$ and is very similar to other marginal diatomites of Holocene age (chapter 7DIIa). It is covered by 1.1 m of well-rounded pumice pebbles (unit U2c). Unit U2 only provides evidence for a single lacustrine transgression. The lake seems to have abutted against slopes of volcanic agglomerate and tuff which provided little clastic sediment. The diatom frustules in unit U2b are extremely well preserved, and contain a large proportion of freshwater epiphytic or benthic species (Appendix 4).

c. Lake-level record

We now come to the thorny problem of correlating between the different parts of the Ajewa gully. Table 6.8 is an attempt to reconstruct the complete sequence, using sedimentary criteria and radiometric dates, and allowing for downslope facies changes.

The basal part of the sequence, up to the diatomaceous mudstone DM3b, presents few problems. However, since DM3b is missing at the downstream end, there is no way to demonstrate from the exposures in 0738-D2 that D3 is younger than DM3b. Fortunately both are preserved in the adjacent Rebo section (0738-D13), where an erosional unconformity separates the two. We are also faced with the problem of correlating the lower, lacustrine part of UM1 with the sequence further downstream. Its equivalence to unit DM3b, as shown in Table 6.8, is purely hypothetical, because the facies present are so different, but derives some support from the great similarity between the overlying units DM4 and UM2.

Units D3 and UM4 have many characteristics in common. They are both calcareous, very silty, and rich in faunal remains, especially mollusca. Neither DM3b nor the older laminated diatomite (D1/DM1) contain shells or shell casts, and DM3b is noncalcareous. The ^{14}C dates strongly suggest that D3 is a deep-water equivalent of UM4, and it seems likely that unit U2 is also of the same age.

The lateral passage from UM5 to U3 can be followed on the ground. It is not at all clear how they relate to DM5 further downstream. Unit DM5 is, however, older than a lacustrine transgression (Ajewa V) which reached about 1601 m at P40 (unit DM6). It seems probable that unit D4 is a lateral equivalent of the DM5 shoreline gravels at P40 and that D5 records the subsequent regression. The youngest unit, D6, has no certain counterpart upstream.

Table 6.8: Composite stratigraphy of the type section of the Ajewa formation (0738-D2)

Simplified lithostratigraphy (palaeosols not shown)	Units	Inferred lake-level fluctuations	Lacustral phases	^{14}C dates[1] (yr BP)
Cross-bedded, greyish deltaic sands and silts with ostracods.[2]	D6	Minor rise to ca. 1564 m	AJEWA VI	pre-1933 AD
Bedded, gravelly sand[2]	D5	Recession < 1559 m		
Cross-bedded beach gravel or plane-bedded sugar sand (upslope)[2] Pale brown, shallow-water lacustrine sands and silts with ostracods and fish bones (downslope)[2]	DM6 D4	Minor rise to ca. 1600 m	AJEWA V	LATE HOLOCENE
Yellow-brown colluvial loams with land snails (upslope) Pale brown alluvial silts (downslope)	UM5/U3 DM5	Lowstand < 1569 m		▲ 2510 ± 100
Grey lacustrine silts and white diatomite (upslope) Pale brown lacustrine silts and sands with concentrations of shells, ostracods and fish bones (middle) Pale brown, laminated lacustrine silts including Roricha fish bed (downslope)	U2 ? UM4 D3	Highstand ≫ 1616 m	AJEWA IVB	● 5530 ± 65 ⊖ 5610 ± 100 (● 5660 ± 55)[3] ∇ 5750 ± 70 ● 6150 ± 150
Greenish fluvial sand and imbricated diatomite gravel (downslope)	D2 ?	Recession < 1561 m		
Reworked mollusc shells and fish bones	(UM4)	Highstand > 1646 m	AJEWA IVA	○ 8320 ± 175 ⊖ 9220 ± 190 ● 9480 ± 140 ● 10,010 ± 150
		Brief recession		
Fluvial gravel passing up into greyish-brown swamp deposits with fish bones and freshwater shells (upslope)	UM3	Minor rise to ca. 1632 m	AJEWA III	▲ 10,220 ± 140
Yellow-brown colluvial loams, calcreted in part (upslope) Pale brown, gritty, colluvial/alluvial loams with calcrete layers. Contain LSA artifacts (middle)	U1 ? DM4.?/UM2	Lowstand < 1594 m		"TERMINAL PLEISTOCENE"
Shelly, tufa-cemented gravels overlying reworked diatomaceous silts (upslope) Light yellow diatomaceous mudstone (middle)	UM1 ? DM3b (upper part)	Highstand > 1636 m	AJEWA IIB	⊟ 14,400 ± 750 ? UPPER
Grey fluvial gravel with diatomite clasts (middle)	DM3b	Brief recession		LATE
White diatomite overlying cross-bedded light grey deltaic sands and pebbles (middle)	DM3b (lower part) DM3a	Minor rise to > 1605 m	AJEWA IIA	PLEISTOCENE ?
Greenish-grey, gritty, colluvial silts and sands with small, calcareous nodules (middle)	DM2	Lowstand < 1584 m		
Finely laminated, greenish diatomite	D1/DM1	Highstand ≫ 1602 m	AJEWA I	LOWER LATE PLEISTOCENE ?
Caldera volcanics				MIDDLE PLEISTOCENE 0.20-0.24 m.y. or less

1. Symbols as for Fig. 6.13, with the addition of ∇ fish bones.
2. May be time-equivalent to upper part of UM5/U3.
3. Date interpolated from section 0738-D2-3.

According to the model in Table 6.8, we can distinguish three complex lacustral phases and two shorter episodes. The earliest phase, Ajewa I, represents the first flooding of the caldera. The second, Ajewa II, comprises a double highstand (IIA - IIB), with a brief intervening regression. Ajewa III and IV make up the familiar tripartite Holocene lacustral phase which has left traces all round the basin. The two ^{14}C dates from the adjacent Mirrga graben show that the lake overflowed during the Ajewa IVB episode.

The lake then retreated and its deposits were partly eroded or buried by colluvium. Two smaller transgressions, Ajewa V and VI, have occurred since 2510 BP, reaching at least 1601 and 1564 m respectively. Ajewa VI probably occurred during the last few centuries (chapter 5BIII).

d. <u>Correlations with the Bulbula formation (Table 6.9)</u>

The earliest lacustrine highstand in the Shala Basin, <u>Ajewa I</u>, almost certainly has no equivalent in the Bulbula exposures. It will probably be found to correspond to levels below 56 m in the Abiyata Core.

<u>Ajewa IIA and IIB</u> can be tentatively equated with the Late Pleistocene Bulbula I and II highstands deduced from the Deka Wede section. If this is correct, then unit DM3 should correspond to the Kurkura diatomites (table 6.4). This interpretation is quite different from the views expressed earlier by Grove <u>et al</u>. (1975), who regard the Ajewa II episode as a precursor of the Holocene lacustral phase. The new correlation proposed here is based on several lines of evidence:

1) The <u>Melosira</u>-dominated diatom flora of the yellow mudstone in unit DM3b is strongly reminiscent of biozones C and E in the Kurkura diatomites and zone A_2 in the Abiyata Core (Figs. 6.6, 6.8). This is a flora characteristic of deep, oligohaline lakes.

2) The almost carbonate-free, diatomitic facies of the mudstone is also very similar to the Kurkura diatomites in sections 0738-B3, B27 and the Abiyata Core, where carbonate minima are also found.

3) The pattern of an initial brief highstand, a recession, and a second, longer-lasting, deep-water phase is reminiscent of Bulbula I and II.

Table 6.9: Correlation of principal lacustral phases in the Ziway-Shala Basin

Local lake-level sequences			Combined lake-level sequence	Estimated dates (yr BP)	Elevations (m.a.s.l.)
Bulbula Plain	Abiyata Core	L. Shala			
		Ajewa VI	ZIWAY-SHALA VIII	pre-1933 AD	ca. 1564
					< 1559
		Ajewa V	ZIWAY-SHALA VII	Late Holocene	ca. 1600
				< 2500 < 4800	< 1569
Bulbula IV	Abiyata IV	Ajewa IVB	ZIWAY-SHALA VI		1670
				6500 8500	< 1561
Bulbula IIIC Bulbula IIIB	Abiyata III	Ajewa IVA	ZIWAY-SHALA V		1670
				9800 10,200	< 1579
Bulbula IIIA	Abiyata II	Ajewa III	ZIWAY-SHALA IV		ca. 1632
				11,500 21,000 ?	< 1594
Bulbula IIC Bulbula IIB Bulbula IIA	Abiyata I ?	Ajewa IIB ?	ZIWAY-SHALA IIIC ZIWAY-SHALA IIIB ZIWAY-SHALA IIIA		ca. 1641 ≫ 1622 ≫ 1617
				26,500 ? 28,000?	< 1591
Bulbula IB Bulbula IA	?	Ajewa IIA?	ZIWAY-SHALA IIB ZIWAY-SHALA IIA		> 1638 > 1609
				30,000?	< 1584
?	?	Ajewa I	ZIWAY-SHALA I	Lower Late Pleistocene?	> 1602

4) The absence of faunal remains in unit DM3 contrasts starkly with lake sediments of early and mid-Holocene age.

5) The single radiocarbon date of 14,400 BP on calcareous tufa from unit UM1 should probably be regarded as a minimum age for units DM3 and UM1, given the evidence for pedogenic translocation of carbonate in the overlying unit UM2 and the signs of lateral seepage along the top of the limestone layer. It seems unlikely that a total thickness of 6.5 m of colluvial/alluvial and swamp sediments, including two soils, could form between 14,400 ± 750 and 10,010 ± 150 BP. An age of about 21,000 y for the UM1 tufa seems far more likely on the basis of the chronology from 0738-B20.

The Ajewa III and IV lacustral phases together correspond to Bulbula III and IV. The ^{14}C dates suggest that the Ajewa IVA maximum equates with Bulbula IIIB - IIIC, and Ajewa IVB with Bulbula IV. The shortlived rise to ca. 1632 m before 10,220 ± 140 BP (Ajewa III) may correspond to Bulbula IIIA and to the pronounced freshening of Lake Abiyata in the interval represented by zone B (Fig. 6.8).

The most recent episodes, Ajewa V and VI, have no counterpart in the higher areas of the Bulbula Plain. Ajewa V seems to be equivalent in age to the late Holocene 1595-1600 m shorelines, which can be traced across the Ajewa area (Fig. 5.7). Ajewa VI, as previously mentioned, was apparently restricted to Lake Shala.

Table 6.9 is an attempt to combine the stratigraphic evidence from the central Bulbula Plain, the Abiyata Core and the Ajewa embayment. This is based on Tables 6.2, 6.4 and 6.8, and Figure 6.8. The principal lacustral phases in this composite sequence have been numbered Ziway-Shala I-VIII. The environmental conditions prevailing during each episode are reconstructed in chapter 8B.

6E THE MEKI FORMATION

I. The Italconsult cores

As yet we understand very little about the way in which African rivers responded to climatic changes, except perhaps in the case of the Nile, where it could be argued that we know too much about fluvial erosion and aggradation, and too little about palaeoclimate. In many basins the dynamic response of the rivers was also conditioned by fluctuations in base level. This is the case in the Meki Valley, where the level of Lake Ziway has varied from 1635 m to 1670 m during the Holocene. In doing so, it has changed the pattern of sediment deposition upstream. This section summarizes the available information on the stratigraphy of the Meki Valley. It does not pretend to be more than a reconnaissance report on an interesting and complex area.

In 1970 the Italconsult engineering team drilled a line of five boreholes into the valley fill at the mouth of the overflow channel, and penetrated at least 50 m of silts and sands with interbedded gravels (Italconsult, 1970, vol. 3) (see also chapter 5BI).

In 1975, I heard that these cores had been kept by the Awash Valley Authority in Addis Ababa, and was granted permission to examine them. To my dismay, they were stored in wooden core boxes in an open yard. Only 55.75 m out of the original 170 m of core had survived. Fortunately it was still possible to piece together at least the broad outlines of the stratigraphy (Table 6.10).

The environments of deposition can be interpreted using data on surface sediments in the Meki Valley:

Unit 1 is similar to the present bedload of the Meki River at this point, which consists of grey, moderately well sorted, medium to coarse, pumice sand (chapter 2EIV). But the unit 1 material is generally finer-grained. Unit 2, which consists of well-rounded clasts of resistant, extrusive rocks, is more difficult to match. It is most comparable with the gravel bar at 1671 m which plugs the mouth of the overflow channel. Unit 2, however,

Table 6.10: Summary of stratigraphy in Meki Valley boreholes

Unit	Minimum thickness (m)	Description
3	16	Brown to dark greyish-brown (10YR 4/2 to 5/3), gritty fine sands and silts with occasional clayey partings and lenses of pumice granules or scattered pebbles. Strong dark brown (7.5YR 3/4) to yellow-brown (10YR 5/8) mottling. Some Fe-Mn concretions along partings. Slightly to moderately calcareous in parts. Fine, lenticular bedding.
2	4.5	Pebble- and cobble-gravel, with weakly to strongly cemented matrix of mottled, brownish (2.5Y-10YR 5/2 to 5/3) or pale yellow (2.5Y 7/3) gritty silt and sand. Calcareous in parts.
1	8	Light grey to pale brown (2.5Y-10YR 6/2 to 7/3), fine pumice sand and silt with local olive-yellow (2.5Y 6/6) to dark brown (7.5YR 3/4) mottling or staining along root channels. Lenses of pumice grit, or dark laminated clays occur in places. At least 3 grey ash layers are present. Noncalcareous to very slightly calcareous.

occupies an estimated altitude range of 1641-1650 m in the cores.[1]

Unit 3 is most closely matched by the sandy soils on old levees of the Meki Delta (Fig. 2.9) (Makin *et al.*, 1976, soil B). These are non-calcareous, loamy soils with predominantly brown colours (10YR 5/3). Soil B has several features in common with unit 3: these include depositional banding; mottling; and the presence of manganese nodules. In the upper

1. The height of each core site was estimated from Italconsult (1970, vol. 3).

delta, the present river is flanked by low, natural leveees some 200–500 m wide, from which the land slopes gently away (< 1%) towards an almost level plain floored by basin clays. Soil B occurs on older, undulating levee ridges in the central part of the delta (Fig. 2.9).

It is possible to see in the borehole record an upward shift from fluvial deposition and transport to a predominantly depositional mode like that prevailing on the Meki delta plain today. During the period of transition, coarse bedload was deposited as gravel bars on the valley floor. A reasonable inference is that the early Holocene rise of ca. 34 m in the level of Lake Ziway resulted in an upvalley migration of the zone of deltaic deposition, so that the pre-existing valley was infilled by fine alluvium up to the overflow level. When lake level finally fell, the river entrenched its course some 16 m near the mouth of the spillway, leaving terraces on either side. Formation of the present Meki Delta then commenced.

The reasons for gravel deposition are less easy to understand. Perhaps the change resulted from a dramatic increase in runoff from the Ethiopian Escarpment at the beginning of the Holocene. Alternatively, a case could be made for higher bedload transport under more arid and seasonal conditions during the terminal Pleistocene. Whatever the explanation, it is clear that the Meki was transporting gravels from the escarpment at the time the overflow channel ceased to function, and that its present load consists of much finer, locally derived material.

II. Downstream changes

The above model of events is supported by the evidence from two sections further downstream (0838-D9 and D4; Fig. 5.3). These show that unit 3 of the Meki formation passes downvalley into impure, pumiceous diatomite similar to the Kertefa deposits (site D9), and then into well-sorted, very fine to medium pumice sands which were probably deposited under shallow-water conditions in the open lake (site D4) (see chapter 2EIV). Buried traces of a pre-Holocene delta plain of the Meki exist in the form of brown, alluvial silt loams underlying these lacustrine sands. The diatomites from the flooded valley contain a very diverse littoral and benthic diatom flora associated with abundant Fragilaria brevistriata (Appendix 4). This assemblage is very similar to the marginal Kertefa diatomites, and is indicative of calm, freshwater conditions with abundant plant growth. Fragments of plant cuticle have actually survived in the sediments, which is unusual. This evidence supports the suggestion made earlier that most of the sediment load of the Meki was dumped in the vicinity of the overflow channel during lake-level highstands.

III. Summary

The stratigraphy of the Meki formation has been pieced together from natural exposures and from cores drilled by Italconsult. Shifts in the pattern of fluvial deposition occurred in response to changes in the level of Lake Ziway. During periods when the lake was overflowing, the Meki aggraded its valley to the level of the outlet. Since the last overspill episode it has incised its channel up to 16 m into this alluvial fill, leaving terraces on either side. No suitable materials have so far been found to date the sequence more precisely.

CHAPTER SEVEN

General Aspects of the Ziway-Shala Lake Sediments

7A SUMMARY

This chapter briefly reviews certain systematic aspects of the lacustrine sediments which are relevant to the interpretation of palaeoenvironments. These include the criteria which were used to identify lake sediments; the nature and origin of the most characteristic sediment types; and the distribution of the latter in space and time. Published rates of deposition from other lake basins have been used to estimate the length of time represented by the poorly dated marls and diatomites in the Bulbula formation.

The main conclusions are as follows:

1) Clastic sediments. These consist primarily of pumice and volcanic ash. They are typical of beach and nearshore environments with an adequate supply of sediment, and are located around the highest shorelines and above and below the thick accumulations of diatomite and marl.

2) Diatomites. These are of two main types: relatively clayey sediments deposited in the centre of the lake basin during lake-level maxima (the "Kurkura" facies) and very pure, localized beds which accumulated in protected embayments characterized by dense plant growth and a low clastic sediment supply (the "Kertefa" facies).

3) Carbonate sediments.

a) Marls. Calcitic marls are restricted to the Deka Wede valley. They seem to be biochemical in origin and were probably precipitated by a group of hot springs entering the lake at the foot of Alutu.

b) Calcareous muds and diatomites. Carbonate deposition also occurred more generally in the inner offshore zone under brackish conditions, which is interpreted as the result of mixing of inflowing river waters with the alkaline lake waters. Carbonate precipitation in the pelagic environment is only definitely known to have occurred during the early Holocene highstand, in association with a *Stephanodiscus*-dominated diatom flora. The cause is suspected to be the photosynthetic extraction of dissolved CO_2 by the phytoplankton.

c) Algal stromatolites have only been found in two localities, on exposed and rocky shores where hot springs may have been active. They appear to be of Late Pleistocene age at both sites.

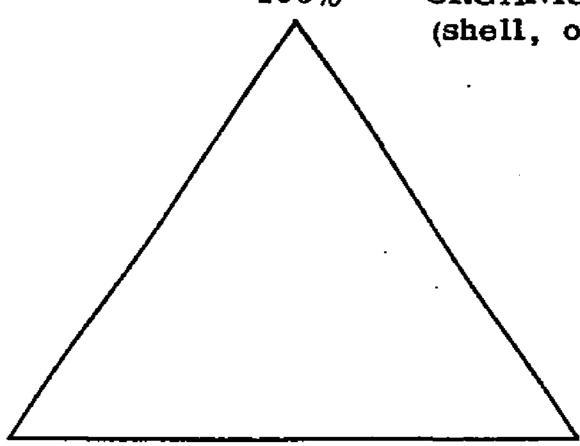
d) Shell concentrations are most common in sediments of early and mid-Holocene age. They are often associated with transgressive and regressive facies. Several categories can be distinguished, based on the faunal elements present and the degree of reworking. Although mollusc assemblages are potentially of value for reconstructing nearshore habitats, biological productivity, dissolved oxygen levels, and (within wide limits) salinity, the origin of shell concentrations in some areas, notably northeast Shala, proved so difficult to establish that only very general ecological inferences have been attempted. The most reliable shell beds for ^{14}C dating purposes appear to be those which contain articulated bivalves and show least sign of reworking.

7B INTRODUCTION

The previous chapter outlined the sequence of lacustrine deposits at individual sites in the Ziway-Shala Basin. This revealed that certain sediment types such as diatomite, marl and clastic deposits have been quantitatively predominant throughout the Late Quaternary, allowing comparisons to be made with the depositional record of other lakes.

The main components of lacustrine sediments in eastern Africa are summarized in the diagram below:

100% ORGANIC COMPONENTS
(shell, opal, gyttja, etc.)



100%
ALLOGENIC COMPONENTS
(detrital, pyroclastic)

100%
AUTHIGENIC COMPONENTS
($CaCO_3$, clays, evaporites, zeolites etc.)

Very dilute lakes such as Victoria or Naivasha are usually characterized by biogenic sedimentation (organic muds and diatomites) (Kendall, 1969; Richardson and Richardson, 1972; Mothersill, 1976); whereas brackish, river-fed lakes like Turkana deposit mainly clastic sediments (Yuretich, 1976; Vondra and Bowen, 1978). Lakes which precipitate large amounts of authigenic minerals other than calcite are found in the more arid areas where groundwater inflows tend to exceed the inputs from precipitation and runoff, and evaporation losses are large. Typical examples are Magadi and Asal (Hay, 1966; Gasse and Stieltjes, 1973; Surdam and Eugster, 1976).

The lakes in the Ziway-Shala Basin have never deposited significant quantities of zeolites or evaporites other than calcite. This suggests that their salinity has rarely risen much higher than at present, and has often been appreciably lower. The pattern of sedimentation during the Late Quaternary shows many parallels with Abhé and the other central Afar lakes (Gasse, 1975, 1977a,b), and to a lesser extent with Lake Turkana (Yuretich, 1976; Vondra and Bowen, 1978). These resemblances have been explored in detail by Gasse and Street (1978a,b) and Gasse, Rognon and Street (in press), and will not be discussed further here. Our main conclusions are summarized in Chapters 8 and 10. This chapter will focus instead on the general problems of identifying lake deposits, and on the origin and distribution of the most important sediment types, in order to draw together the information presented in chapter 6 and provide a background for the interpretation of palaeoenvironments in chapter 8. The time-stratigraphic framework is based on the correlations established in chapter 6 (Tables 6.9 and 8.1).

7C IDENTIFICATION OF LAKE SEDIMENTS

The characteristics of lacustrine sediments in the Ziway-Shala Basin have already been briefly described in chapters 2EIV and 6CII, where comparisons were made with alluvial, aeolian, colluvial and primary pyroclastic deposits. Emphasis has been placed on criteria which are of value in the field rather than on petrographic description. From this viewpoint, the most distinctive properties of the lake sediments are their light colour in unweathered exposures, pronounced sorting by grain size and density, fossil content and sedimentary structures (Table 7.1). Because of the scarcity of published work on modern lacustrine facies in eastern Africa, these criteria have been verified using general literature reviews (Reeves, 1968, chapters 5, 6; Picard and High, 1972; Reineck and Singh, 1973), case studies from other areas (Born, 1972) and personal observations in Ethiopia, Kenya and Tanzania. Many of the properties listed in Table 7.1 are not exclusive to lake sediments: it is the association of characteristics which is diagnostic. The greatest difficulties arise when trying to separate coarse, unfossiliferous beach gravels from ephemeral-stream deposits, and here sedimentary structures are of crucial importance (Picard and High, 1973; Reineck and Singh, 1973; Collinson et al., 1978) (Plates 26, 64-65).

The most useful grain-size parameters for differentiating the various types of lake sediments from other deposits proved to be mean grain size (M_ϕ), sorting (σ_ϕ) (Table 7.1), and % gravel, sand, silt and clay (Fig. 7.1). Skewness and kurtosis are highly variable and not very useful because many samples of lacustrine sediments are bimodal (Figs. 7.1, 7.2). It is not often that more than two major peaks are present. The most commonly occurring modes are as follows:

1) −3 to −2 Ø	Pumice, rhyolite and obsidian gravel
2) −1 to +1 Ø	Coarse to very coarse sand (quartz and feldspar phenocrysts and pumice fragments)
3) +3 to +4 Ø	Very fine sand (coarse ash and pumice fragments)
4) +5 to +7 Ø	Medium to fine silt (weathered pumice and fine ash, diatom fragments).

The existence of these distinct sediment populations reflects not only the source materials (chapter 7D) but also the processes of transport and deposition. Ashley (1978), in her study of sediment input into a Canadian lake, also found distinct modes at +3.8 Ø and +5.8 Ø, which she attributed to transport within the lake by traction and suspension respectively. Her suspended-clay mode (10.3 Ø) is not well-developed in the Ziway-Shala lake sediments, except in the Shala caldera.

The following sections discuss the origin and characteristics of the most important lacustrine sediment types.

Table 7.1: Characteristics of lake sediments in the Ziway-Shala Basin

1) Colour

Light grey or brown in surface exposures
(hue 5Y - 10YR, value 6-8)
- due to the removal of pedogenic clay coatings by abrasion and then oxidation on exposure

2) Texture of non-carbonate fraction

Mean grain size (M_{ϕ})

Ranges from -2.5 Ø (pebbles, S. Langano foreshore)
to +12.5 Ø (deep-water clay, Shala, -160 m)

Modality

Commonly monomodal or bimodal

Sorting (σ_{ϕ})

Ranges from 0.58 Ø (moderately well sorted) (coarse sand, O-itu foreshore)
to 4.65 Ø (extremely poorly sorted) (deepwater clay, Shala, -242 m)

Sorting by density is important as well as sorting by grain size.

Skewness and kurtosis

Highly variable.

Rounding of pebbles

Pumice clasts - subrounded to well-rounded
Harder clasts (lava and ignimbrite) - subangular to rounded (Plate 30)

3) Commonly-occurring bedding and sedimentary structures

Syndepositional structures:

Fine lamination, rhythmites
Even, horizontal bedding (Plates 64-65)
Even, draped bedding
Massive bedding (Plate 58)
} Low-energy environments

Graded bedding
- normal - fine-grained air-fall ashes
- reversed - buoyant pumice lapilli

Cross-stratification (usually in higher-energy foreshore, shoreface and delta-slope environments)
- trough
- planar
- microdelta (steeply inclined foresets) (Plates 25, 35)
- beach-accretion (Plate 26)
- climbing-ripple (Plate 75)

Table 7.1 continued:

Symmetrical, small-wave ripple marks
Reed-stem impressions
Gas-bubble (?) cavities
Shell concentrations, often with a load-cast or brecciated base (below)

Post-depositional structures:

Convolute bedding (Plate 76)
Load casts
Pseudobedding (brecciated mud clasts in a coarser matrix)

4) Composition

High proportion of volcanic glass
Very variable $CaCO_3$ content (0 - 80%) in the form of
- micrite
- stromatolites
- biogenic $CaCO_3$
- small nodules

Low organic content (0 - 8% organic C)
Diatom frustules may form more than 95% of sediment
Low in layer-silicate clays

5) Fossils

Diatoms (Plates 45, 49-51)
Green algae (visible in pollen preparations)
Plant macrofossils
- grass cuticles
- reed-stem impressions
- tiny charcoal fragments (from brush fires ?)

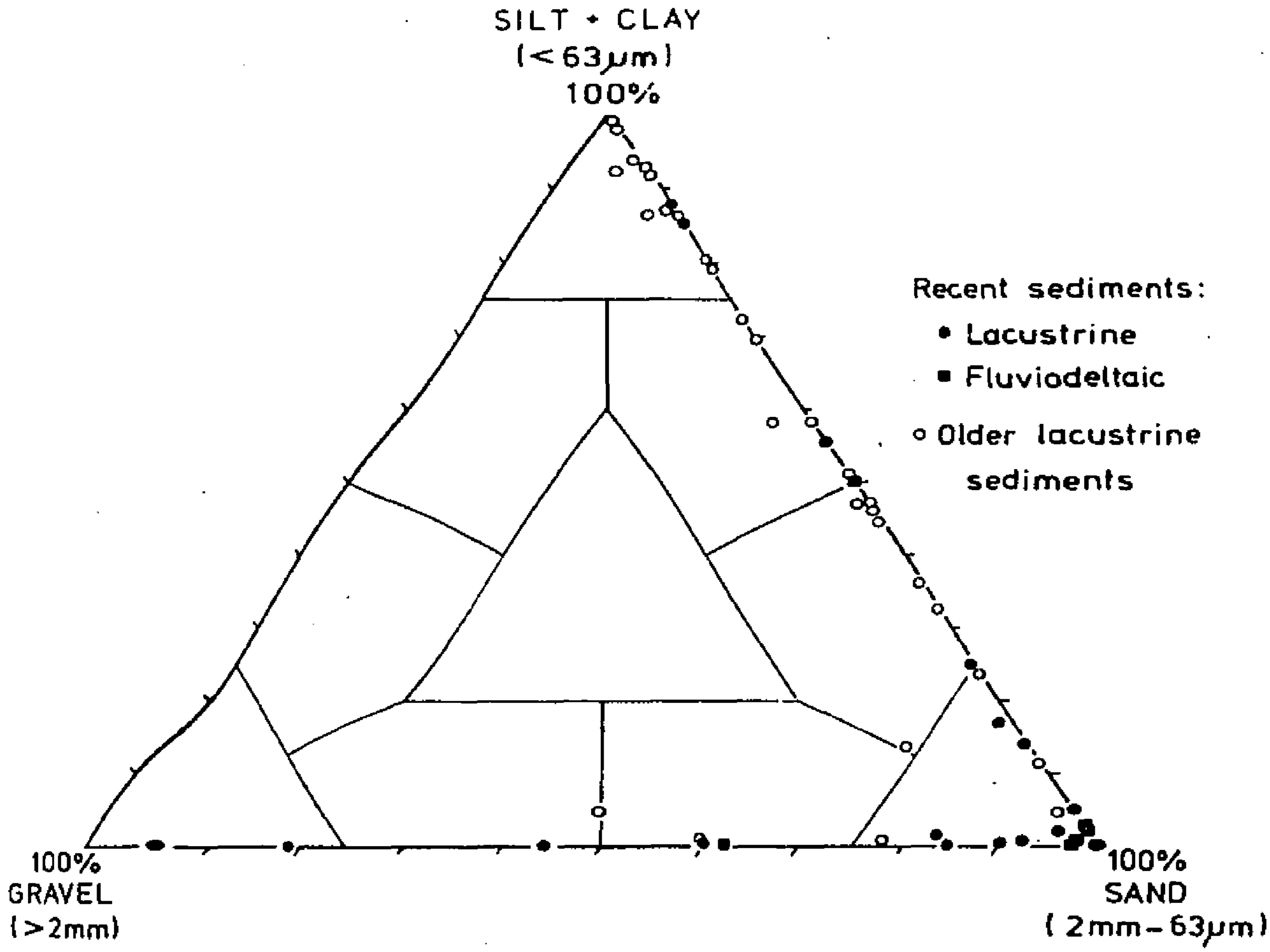
Abundant aquatic mollusca, Bivalves may be articulated (Plates 77, 78)
Terrestrial gastropods rare
Ostracods common in very fine to medium sands (Plate 44)
Fish bones (may be articulated - Plate 66) and otoliths
Freshwater crab remains

6) Stratigraphic sequence

Conformable relationships with swamp or fluviodeltaic deposits but not with fluvial or colluvial deposits (Walther's Law)
(Selley, 1976, pp. 309-310)

Terminology follows Born (1972); Reineck and Singh (1973).

Fig. 7.1: Grain-size distributions of lacustrine and fluvio-lacustrine sediments from the Ziway-Shala Basin (all samples analyzed to date) (for sources see Figs. 2.12, 6.4 and Appendix 7a).


SILT + CLAY
(<63μm)
100%
Recent sediments:
• Lacustrine
▪ Fluviodeltaic
○ Older lacustrine sediments
100%
GRAVEL
(>2mm)
100%
SAND
(2mm–63μm)

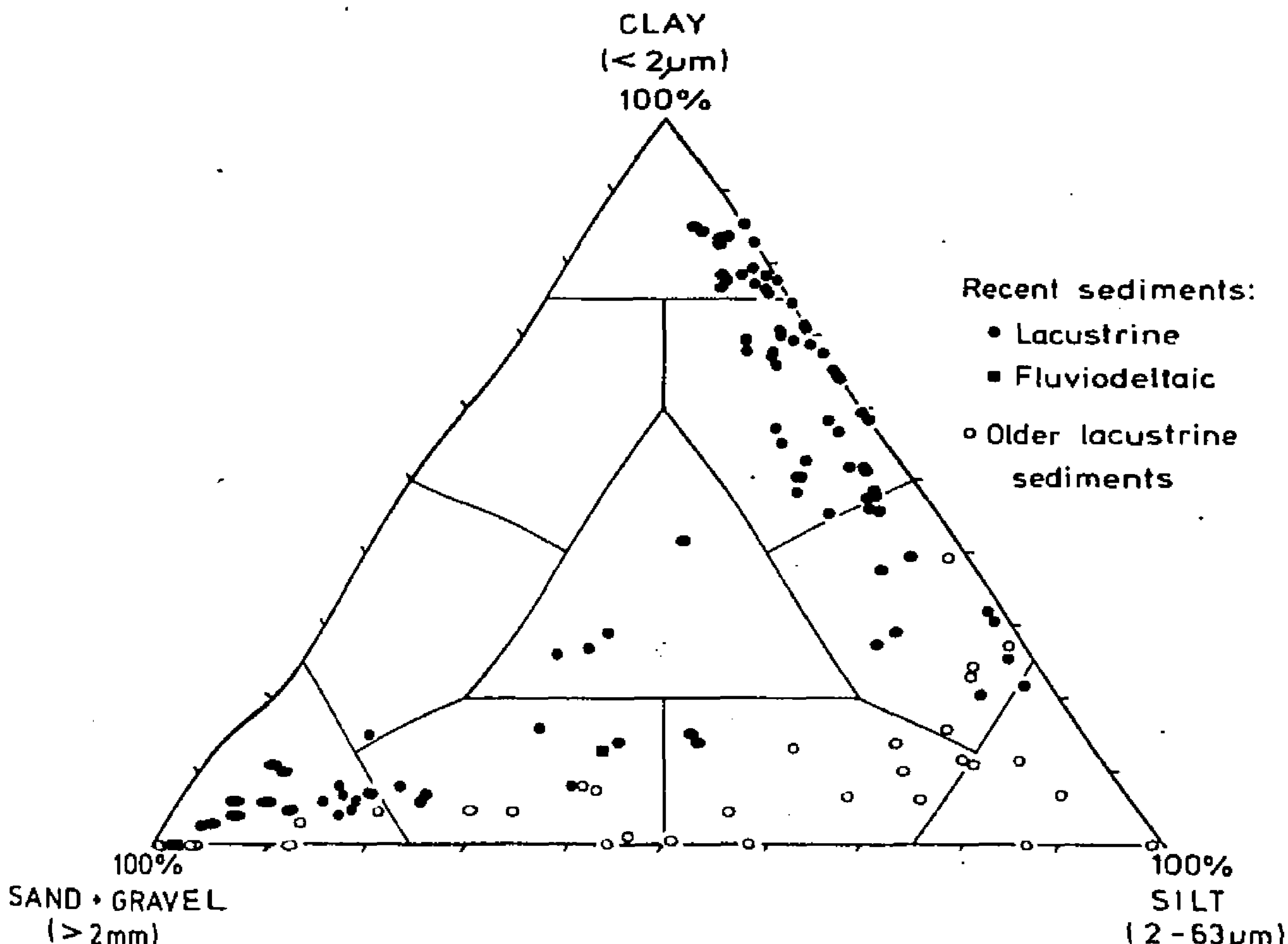
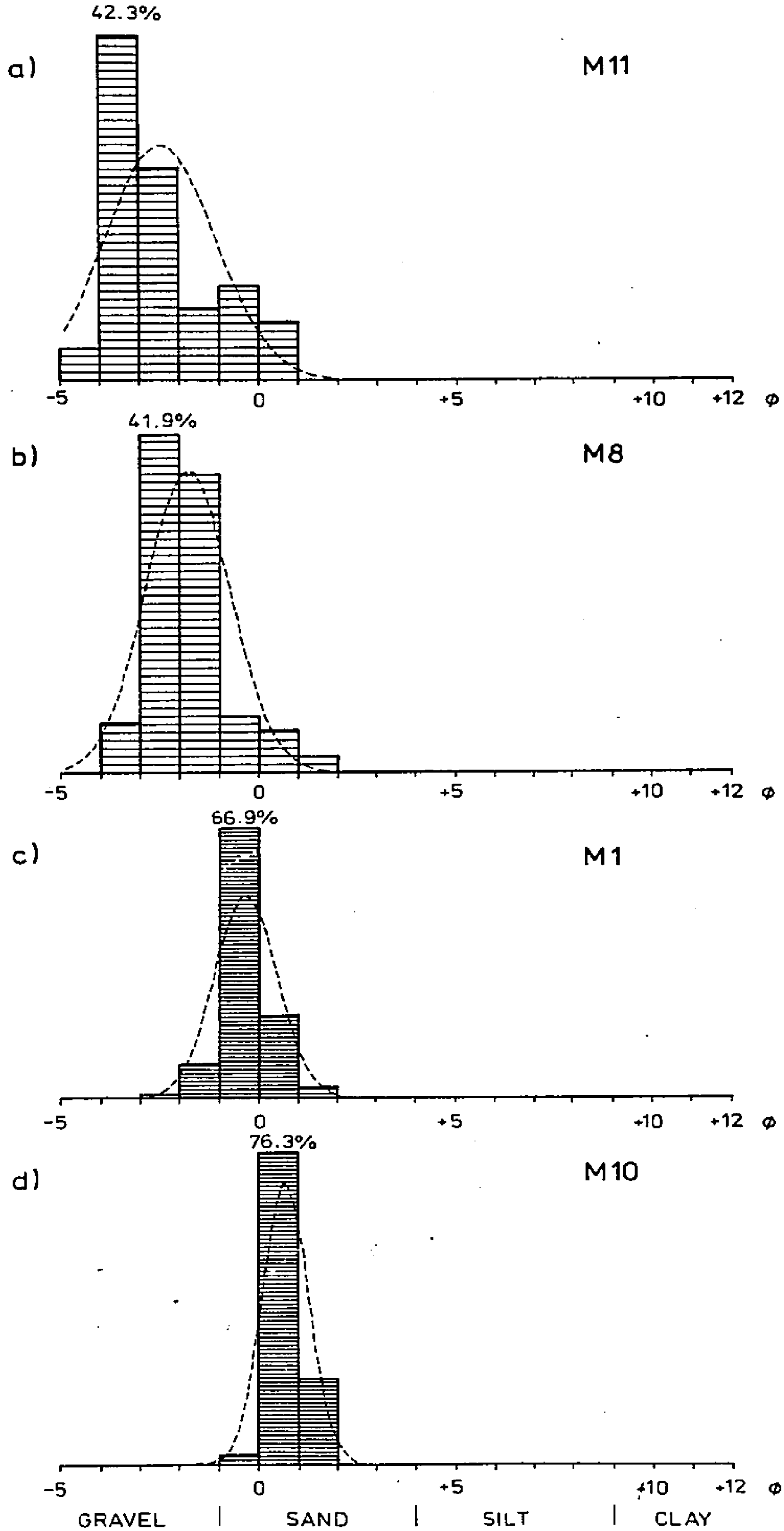
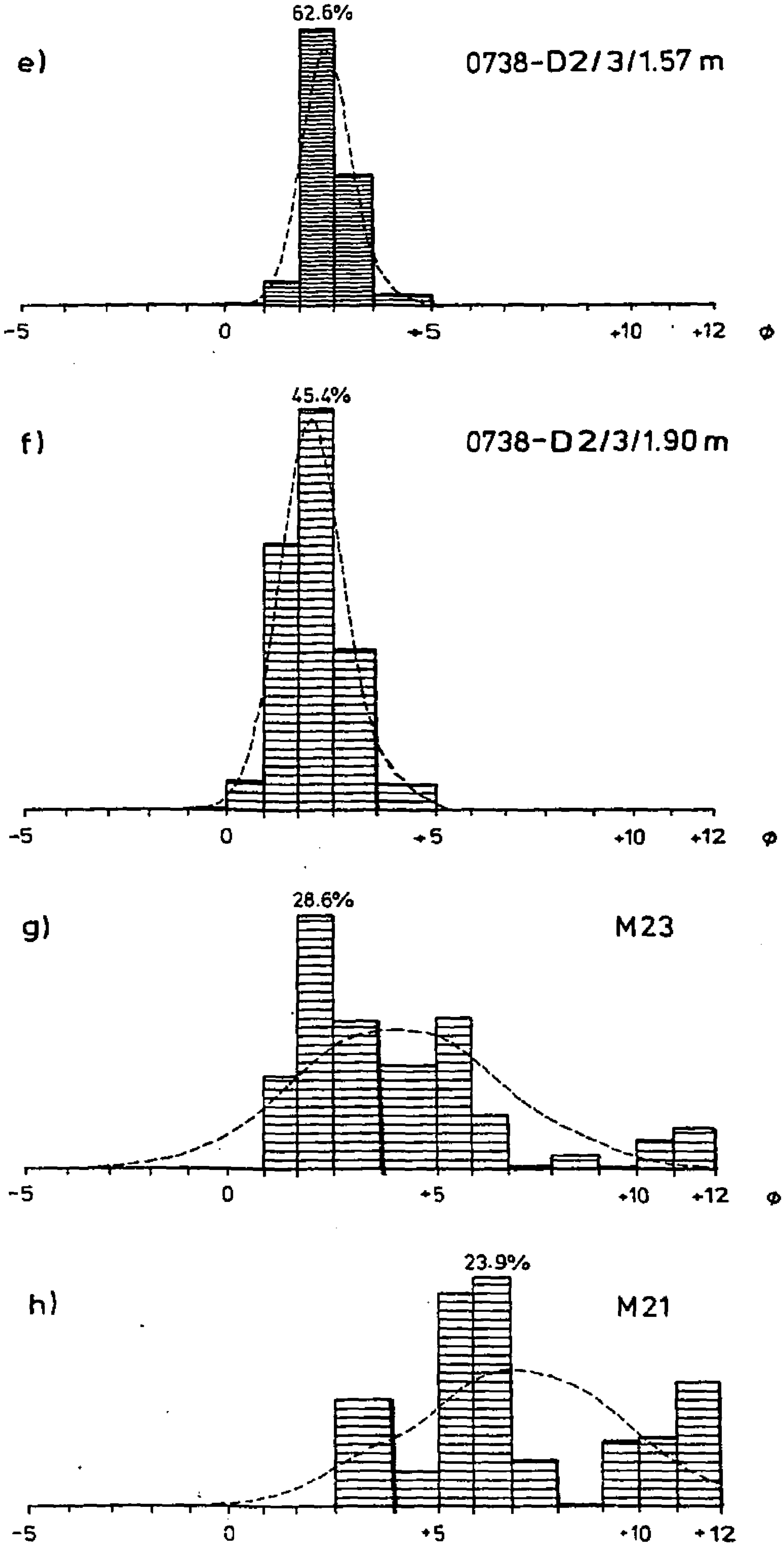

CLAY
(<2μm)
100%
Recent sediments:
• Lacustrine
▪ Fluviodeltaic
○ Older lacustrine sediments
100%
SAND + GRAVEL
(>2mm)
100%
SILT
(2–63μm)

Fig. 7.2: Selected grain-size histograms for lacustrine and fluviolacustrine sediments from the Ziway-Shala Basin (Appendix 7a).

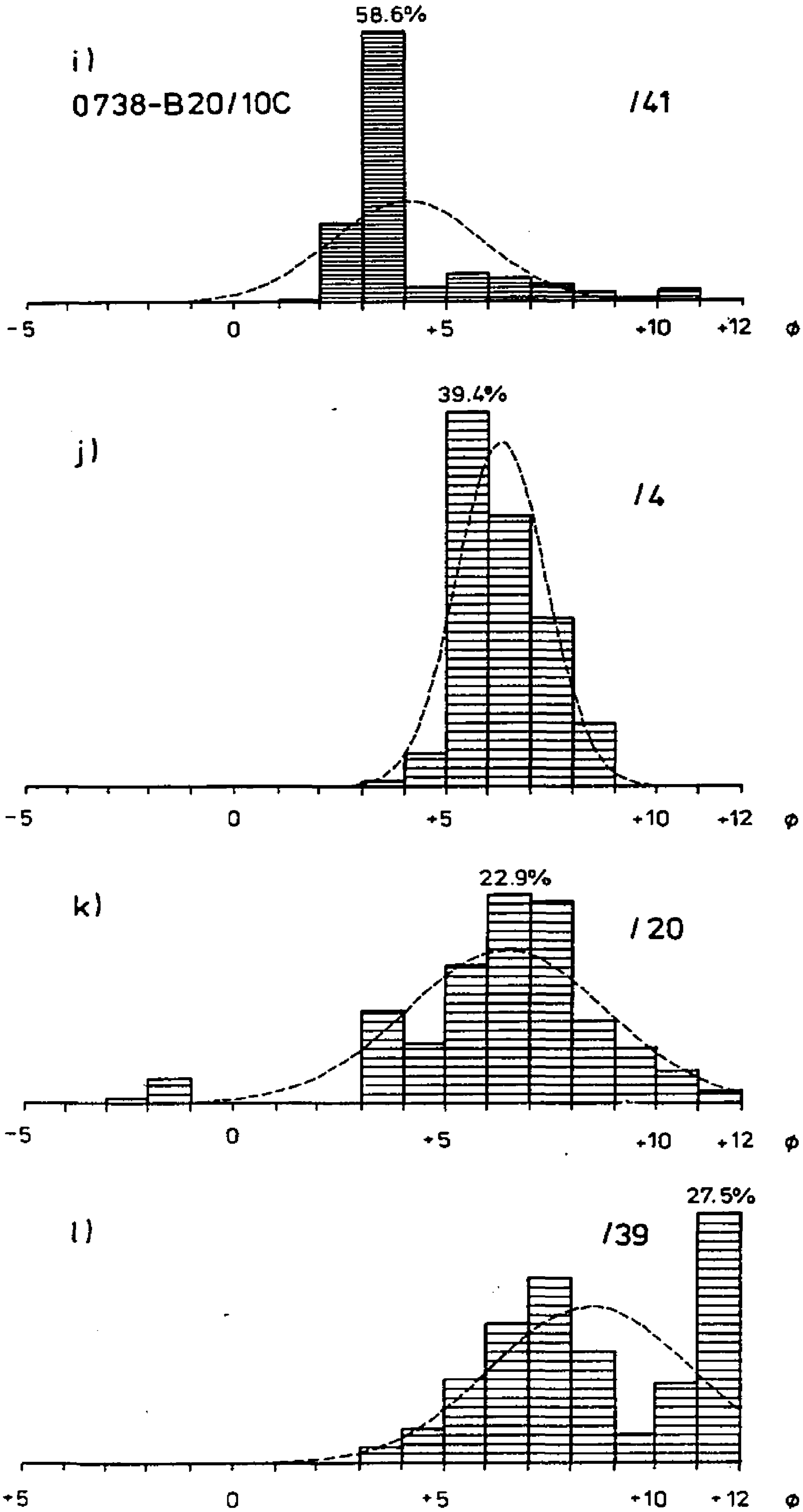
a) Foreshore gravels, SW. Langano
b) Foreshore gravels, O-itu, N. Langano
c) Foreshore sands, NE. Shala (Plate 18)
d) Foreshore sands, O-itu, N. Langano
e) Microdelta foresets, Ajewa fm., unit D6, NE. Shala
f) Microdelta foresets, Ajewa fm., unit D6, NE. Shala
g) Mudflats, SW. Abiyata
h) Mudflats, N. Abiyata
i) Ripplemarked sands, Bulbula fm., member 2a
j) Thin-bedded silts, Bulbula fm., member 4
k) Diatomaceous marl with floated pumice pebbles, Bulbula fm., mbr. 3b (non-carbonate fraction) (Plates 50, 51)
l) High-carbonate marl, Bulbula fm., mbr. 2a (non-carbonate fraction) (Plate 47)

(Dotted lines show normal curves with same $M\phi$ and σ_ϕ as sample distributions.)


42.3%
a)
M11
-5
0
+5
+10
+12
φ
41.9%
b)
M8
-5
0
+5
+10
+12
φ
66.9%
c)
M1
-5
0
+5
+10
+12
φ
76.3%
d)
M10
-5
0
+5
+10
+12
φ
GRAVEL
SAND
SILT
CLAY



e)
62.6%
0738-D2/3/1.57 m
-5
0
+5
+10
+12
φ
f)
45.4%
0738-D2/3/1.90 m
-5
0
+5
+10
+12
φ
g)
28.6%
M23
-5
0
+5
+10
+12
φ
h)
23.9%
M21
-5
0
+5
+10
+12



i)
0738-B20/10C
/41
58.6%
j)
/4
39.4%
k)
/20
22.9%
l)
/39
27.5%
-5
0
+5
+10
+12
ϕ

7D PRINCIPAL LACUSTRINE SEDIMENT TYPES IN THE ZIWAY-SHALA BASIN

I. Clastic sediments

a. Volcaniclastic[1] origin

The dominant feature of all the lake deposits in the basin is the high proportion of volcanic glass (obsidian, weathered ignimbrite, pumice and ash) in the allogenic fraction. This leads to the unusual abundance of silt, very fine sand and granule sizes compared with sediments derived from the weathering of plutonic, metamorphic or sedimentary terrains in which these fractions are generally scarce (Yatsu, 1955; Smalley and Vita-Finzi, 1968). Most of the pyroclastic[1] material seems to have been derived from Mt. Alutu, but a large proportion has passed into temporary storage in slope deposits and fluvial gravels before being reworked by the lake. The stratigraphy of the Bulbula Plain suggests that the volcano entered a particularly explosive post-caldera phase shortly before 21,000 BP (the approximate age of the base of the Abernosa pumice member), continuing to erupt intermittently through the Holocene right up to the last few centuries (chapter 2AIII). During lowstands, most of the pumice must have been deposited where it fell from the eruption plume. In contrast, during high lake phases the fine ash falling onto the water surface settled out to form thin ash bands, but the buoyant lapilli probably floated for long distances, eventually sinking in deep water or drifting onto the foreshore (Plate 17).

b. Variation of grain size with water depth

It has often been assumed that the carbonate-free grain size of lacustrine deposits can be used as a general guide to water depth (Twenhofel, 1932, p. 824), although it may not always be very useful for differentiating the various nearshore environments (Solohub and Klovan, 1970). Because the distribution of sediments in the shoaling zone is largely a response to the balance between the net hydrodynamic force generated by waves, directed onshore,

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1. These terms are used as defined by Fisher (1966).

and the force of gravity, directed offshore (Graf, 1976), most geologists believe that the transition from littoral to fine-grained offshore facies is more abrupt in lakes, which have a shallower wave base and experience negligible tidal effects, than in the marine environment (Picard and High, 1972). Fine-grained facies are however to be expected in marginal environments of low wave energy as well as in deep water (Richardson, 1969b; Butzer, 1971; McLachlan and McLachlan, 1971; Picard and High, 1972). In the Ziway-Shala Basin, floated pumice pebbles are frequently encountered in deep-water muds.

Twenhofel's assumption, which underlies many of the interpretations made in chapter 6, has seldom been tested in low-latitude lakes. Some empirical support is given by Bowler (1970), Born (1972) and Yuretich (1976). The grain-size data from the modern lakes (chapter 2EIV) do not provide an adequate test of the model, although certain indirect evidence suggests that it is also valid for the Ziway-Shala Basin. An orderly passage from nonlacustrine sediments (fluvial gravels or air-fall pumice) through gravels or sands to deepwater muds is very common in transgressive sequences in the Bulbula formation. Shell concentrations (see below), and structures such as microdelta or beach-accretion cross-stratification, climbing-ripple lamination, load casts and liquefaction features are common in the sandier facies, and fit in with deposition close to, or at, the shore. Further support for the basic model is provided by the relationship between the grain size of the sediments and their diatom assemblages (Figs. 6.6, 6.8).

c. Distribution

At most sites, clastic sediments laid down during shallow-water conditions gave way to diatomites or marls as the water-level rose. The Deka Wede and Kurkura sections are good examples (Figs. 6.3, 6.6, 6.7). In the basin as a whole, gravels and sands are therefore found above and below the thick accumulations of fine-grained sediments, and as marginal belts related to the highest shorelines (chapters 5, 6; Plates 25, 26, 35). The exceptions are those sheltered bays where the supply of sediment was low. Here one finds shallow-water diatomites or stromatolites (see below). There are also differences between the proportion of clastic material in the deposits of the various highstands, especially in the Ajewa area. We will return to this problem in chapter 8B.

II. Diatomites

Diatomites or diatom-rich muds form an important constituent of the lacustrine sediments. Two principal types can be distinguished: shallow-water diatomites, which are often very pure (the "Kertefa" facies); and profundal[1] diatomites or diatomaceous muds (the "Kurkura" facies). Both were deposited during deep-lake phases. Calcareous diatomite is also found as a transgressive and regressive facies in the Abiyata Basin (section IIIb).

a. Shallow-water diatomites

Most of the shallow-water diatomites, including those in the Kertefa area, are of early or mid-Holocene age, although Upper Pleistocene examples are also known (Fig. 7.3). With one exception these diatomites occur only above 1645 m. The sediment is typically pure white (2.5Y > 8/1), apart from localized iron staining along root holes or reed-stem impressions; very light; and fluffy in consistency. Measured bulk densities range from 0.17 to 0.45 g/cm^3 (Appendix 7, samples 1-5). Samples examined under the microscope are seen to consist almost entirely of diatom frustules, which are often extremely well preserved. Detrital particles are very infrequent. Calcium carbonate may be present (0 - 12%). In the Mudi section (0738-B14) the diatomites contain Melanoides shells and articulated Pisidium sp., but otherwise mollusca are scarce. Fine, horizontal or inclined laminations which are conspicuous at some sites (0738-B16, for example) probably reflect seasonal variations in stream-flow, but in other localities the diatomite forms a single, massive bed.

Several of these sites, such as 0738-B28, B16 and D2 (unit U2b) are found in protected, rocky bays, but this is not invariably true. In the Meki

1. Pertaining to or existing in the deeper part of a lake, below the limit of well-developed zones of vegetation (Gary et al., 1974), p.400) (see Appendix 8).

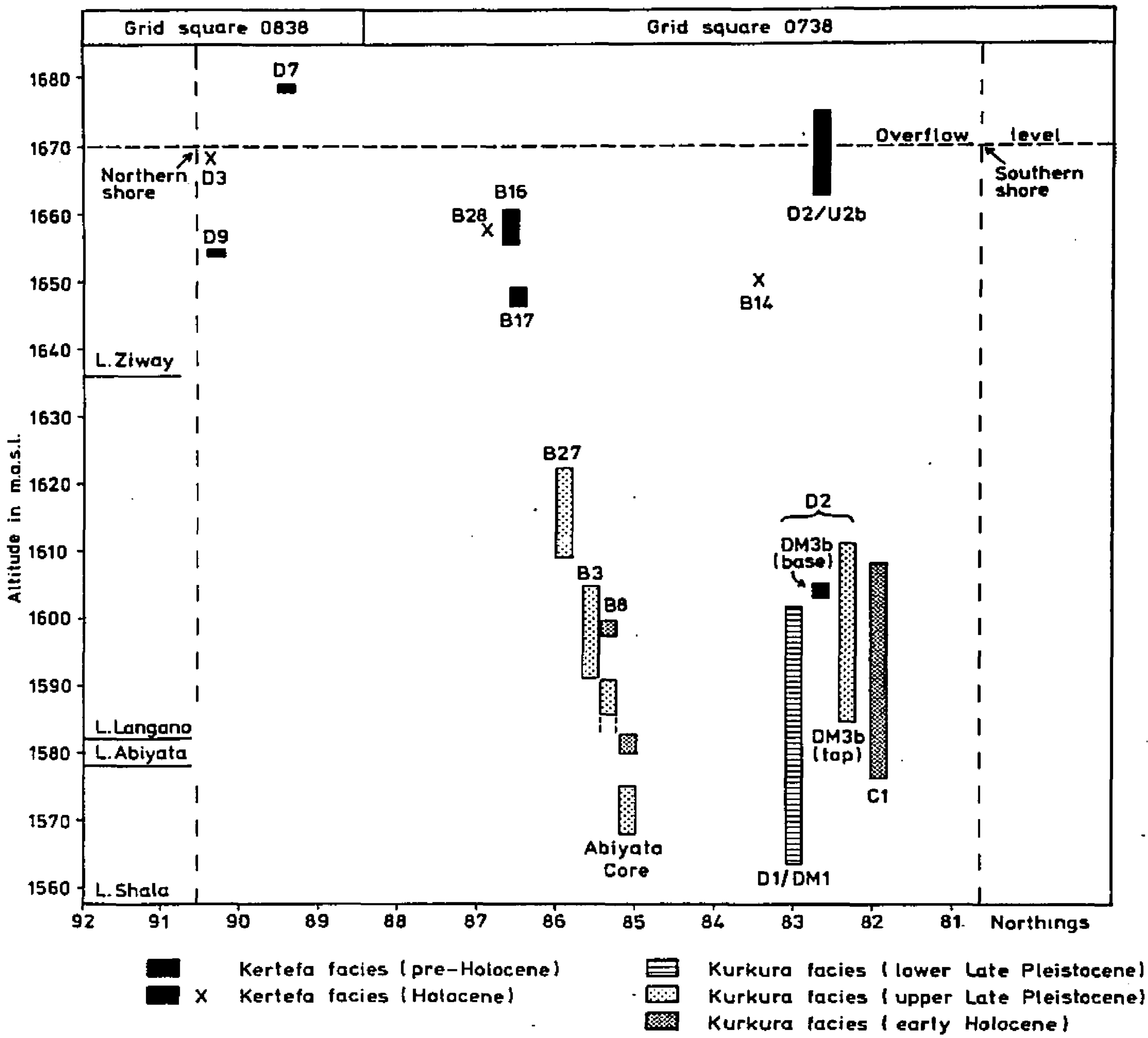

Fig. 7.3: Altitudinal distribution of diatomites in the Ziway-Shala Basin.

valley and at sites 0738-B14 and 0838-D7 the lake was probably bounded by pumiceous sediments. Paradoxically, the diatomites at these last two localities and at 0738-D2 (unit DM3b) occur in sequences of shallow-water sands with microdelta cross-bedding and/or climbing ripples, which suggest rapid sedimentation (Reineck and Singh, 1973, pp. 95-97). The change to diatomite deposition must indicate near-cessation of the clastic input. How this came about is not clear - perhaps through lateral shifting of stream channels or an abrupt rise in lake level.

Towards deeper water, for example going down the Meki valley from 0838-D3 to D4, or from 0738-B16 to B19/1, the diatomites grade first into diatomaceous silts and very fine sands with bulk densities of about 0.57 - 0.63 g/cm^3 (Appendix 7, samples 6-8), and then into predominantly clastic sediments (Fig. 6.10).

Dr. Françoise Gasse has kindly examined samples of the marginal diatomites from seven sites (Appendix 4). With the exception of the Middle or Upper Pleistocene diatomite at Rama Gebriel (0838-D7) (Fig. 5.15), the assemblages are very rich in freshwater epiphytic and benthic species. Towards the open lake first the facultative planktonic species *Fragilaria brevistriata* and *Cyclotella ocellata*, and then fully planktonic species such as *Stephanodiscus astraea* and *S.* aff. *minutus* become dominant. The flora from Rama Gebriel is much less diverse and is unusual in that it indicates relatively high or variable salinity, although littoral (largely epiphytic) species are again very abundant (Appendix 4).

The picture which emerges is one of shallow, confined bays and inlets in which pockets of aquatic macrophytes sheltered large populations of benthic and epiphytic diatoms. Probably the vegetation was fairly open, since the dense shade typical of papyrus swamps tends to inhibit algal growth (Howard-William, 1972). It is difficult to account for the lack of mollusca, especially pulmonates, in this setting, though low dissolved oxygen levels may have been responsible. The supply of incoming sediment was low; in some cases because the shoreline was very rocky, and in others probably because the adjacent slopes were densely vegetated and the runoff was filtered by passage through reed beds (Thompson, 1976). In the case of the Meki Valley, the bulk of the sediment was being deposited further upstream (chapter 6E).

b. Deep-water diatomites

The second group includes the Upper Pleistocene Kurkura diatomites, the Holocene diatomaceous clays in the Abiyata Core, and the laminated diatomite (units D1, DM1) in the Ajewa formation. These are found in the centre of the lake basin, below about 1625 m (Fig. 7.3), and are rarely as pure as the marginal diatomites, which may suggest that the waters of the open lake were relatively turbid. In general, measured bulk densities range from 0.50 to 0.72 g/cm^3 (Appendix 7, samples 9-15), with the exception of the early Upper Pleistocene (?) diatomite D1/DM1, which is very pure and fluffy in consistency (0.38 g/cm^3). It was inferred in chapter 6D that the latter was deposited in deep water under permanently or near-permanently stratified conditions, in the newly-flooded Shala caldera. Its highly variable calcite content (0 - 71%) suggests marked fluctuations in water balance and evaporative concentration.

The Kurkura diatomites are massive or only poorly laminated, which is probably due to bioturbation. This suggests that oxygenated water was in contact with the bottom sediments in the Bulbula Plain. They are low in $CaCO_3$ (0-9%) and organic carbon ($\leq$ 2.9%), containing $10^8 - 10^9$ diatom valves/g of dry sediment (Fig. 6.8); a moderate figure compared with the most prolific diatomites in the Afar, such as Abhé II, which reach $10^{15} - 10^{17}$ valves/g (Gasse, 1977b; Gasse and Descourtieux, in press). The Kurkura diatom assemblages are largely planktonic. Cyclotella ocellata and tropical Melosira spp. are the dominant taxa.

Diatomaceous lower Holocene sediments are found in the Bulbula Village section (0738-B8) and in the Abiyata Core (2.33 - 5.27 m). Although the Abiyata muds contain 10^7-10^9 valves/g (Fig. 6.8), their diatom content is masked by carbonate (7-19%), organic matter (1.6 - 6.6% organic C) and clay. Multiple ash bands are interbedded with the fine sediments. The most characteristic species are Cyclotella ocellata, Stephanodiscus astraea and its variety minutula, and Fragilaria brevistriata.

As a group, the profundal diatomites are much thicker and more extensive than the Kertefa type, and show much less lateral variation. They were deposited where potential water depths exceed 45 m (Fig. 7.3), and their planktonic floras record deep-lake conditions. Unlike the marginal diatomites, they are sometimes draped over highly irregular bottom topography. The laminated Ajewa diatomite (D1/DM1) mantles slopes of up to $19\frac{1}{2}^{o}$ with little sign of flowage or deformation. Diatomites still plaster the steep inner walls of the Chitu Haro crater (Plate 33) although there some slumping has occurred. The amazing ability of benthic diatoms to stabilize sediment has been attributed to the mucilage secreted by certain species (Holland et al., 1974).

c. Rate of accumulation

Washbourn-Kamau (1971) and Isaac (1977) have attempted to estimate the time-span represented by ancient diatomites in the Kenyan Rift. They used published rates of deposition for Holocene and pre-Holocene diatomites (Table 7.2, refs. a-f). When lake or ocean cores are used it is necessary to correct for the moisture content of the sediment. The resulting deposition rate in $g/cm^3/y$ can then be converted into linear measure using the actual density of the fossil diatomite under consideration. A value of 0.60 g/cm^3 will be used here to represent the profundal diatomites.

Measured accumulation rates for the Ziway-Shala Basin range from 1 m in 100 y for the basal Holocene beds in the Abiyata Core to 1 m in 1310 y for the Ziway-Shala V deep-water phase. However the accuracy of these extremes is questionable (chapter 6CIVa). Geze obtained a rate of 1 m in 489 y for the silty Kertefa diatomites in the vicinity of section 0738-B17, whereas the much purer laminated diatomite (unit D1/DM1) in the Ajewa formation contains an average of 1100-1150 varve couplets/m.

A deposition rate of 1 m in 600 y will be taken to represent the Kurkura diatomites in the Bulbula formation, which contain a moderate amount of clastic material. This figure is based on the rates obtained for impure deep-water diatomites in Lakes Abhé and Naivasha (Table 7.2). It yields estimates of 8400 y and 6300 y for the deposition of members Kurkura 3 and 4 combined, in sections 0738-B3 and B27 respectively, and 1650 y for member 2 in section

Table 7.2: Rates of accumulation of diatomaceous sediments and marls

Locality	Source	Rate of accumulation ($g/cm^2/y$)	Rate of accumulation (years per metre)
1) **Diatomites:**			
a) Valle dell'Inferno, Rome (Middle Pleistocene varved diatomites)	Bonadonna 1965		952
b) Luneburg Heath, Germany	Dewall 1928		588
c) Munster, Luneburg Heath, Germany	Giesenhagen 1925		909
d) Gulf of California (marine diatomaceous opal)	Calvert 1966	0.0006-0.174 av. 0.05	345-100,000 * 1200
e) L. Naivasha core (diatomaceous mud with 11-21% organic matter)	Richardson & Richardson 1972		
9200-5650 BP		0.109	550*
5650 BP - present		0.049	1220*
f) L. Tanganyika core (diatomite with 10-16% organic matter)	Livingstone 1965	0.0161-0.0732	820-3730*
g) L. Abhé core (Abhe III clayey calcareous diatomite, 25,600-23,900 BP)	Gasse and Delibrias 1977		625
h) Kertefa area, Ziway-Shala Basin (diatomaceous silts, 9360-4960 BP)	Geze 1975		489
i) L. Abiyata core (diatomaceous clays with ash bands, 9950-7970 BP)	This study		100-1310
j) L. Shala, section 0738-D2, unit DM1 (laminated diatomite)	This study		1100-1150

Table 7.2 continued:

Locality		Source	Rate of accumulation ($g/cm^2/y$)	Rate of accumulation (years per metre)
2)	Marls:			
k)	L. Abhé core (Abhé V, 7250-5830 BP)	Gasse 1975		660
l)	As Ela, Abhé Basin (7610-4120 BP)	Gasse 1975		2220
m)	L. Afrera (diatomaceous marl, 8750-7300 BP)	Gasse 1974b		592
n)	Bulbula Plain, Ziway-Shala Basin, section 0738-B20/10C, member Deka Wede 2 (marls and calcareous sands, 24,000-22,050 BP)	This study		328
		Average for Ethiopia and Djibouti		950
3)	Average Rift-floor sedimentation rates:			
o)	L. Abhé core (25,600 BP to present)	Gasse and Delibrias 1977		1290
p)	L. Abiyata core (10,370 BP to present)	This study		1240
q)	Bulbula Plain, Ziway-Shala Basin, section 0738-B20 (27,050 BP to present)	This study		1110
		Average for Ethiopia and Djibouti		1210

* Taking an average density of 0.60 g/cm^3.

All rates are given to three significant figures.

B27. The duration of lacustral phases Ziway–Shala II and III was therefore of the order of 1650 y and 6300 - 8400 y at these sites.

III. Carbonate sediments

Calcareous deposits in lakes can be chemical, biogenic or biochemical in origin (Reeves, 1968; Müller et al., 1972; Brammer, 1978). In the Ziway-Shala Basin there is abundant evidence to suggest that we are dealing mainly with the second and third categories. These include the skeletal remains of mollusca and ostracods, and fine-grained calcite (marl and stromatolites) which appears to have been precipitated partly as a result of photosynthesis by lacustrine organisms.

Precipitation of nonskeletal calcium carbonate in the lacustrine environment may result from a number of factors which tend to shift the equilibrium below to the right:

$$Ca^{++} + 2HCO_3^- \rightleftharpoons CaCO_3 + H_2O + CO_2 \qquad \text{(Bathurst, 1975, p.232)}$$

These include:

1) Mixing of water masses of different composition
   (Müller et al., 1972; Hecky and Degens, 1973; Hewer, 1975; Baumann et al., 1975).
2) Evaporative concentration
   (Neev and Emery, 1967; Müller et al., 1972).
3) Seasonal or long-term changes in p-t conditions
   (Deevey, 1953; Turner, 1970; Müller et al., 1972; Terlecky, 1974; Yakushko et al., 1978).
4) Photosynthetic extraction of CO_2 by
   a) blue-green algae (cyanophytes) (Eardley, 1966; Barron, 1975)
   b) diatoms (Stoffers and Fischbeck, 1974; Gasse, 1977b)
   c) green algae (chlorophytes) (Dodson, 1974; Beadle, 1974, p. 210)
   d) calcareous algae (charophytes) (Terlecky, 1974)
   e) submerged leaves of aquatic macrophytes like
      Elodea and *Potamogeton* (Ruttner, 1963, pp. 61-73); Müller, 1971; Förstner, 1973; Brammer, 1978).

5) Aerobic or anaerobic bacterial activity, usually within the sediment (Neev and Emery, 1967; Krumbein, 1972; Adolphe and Billy, 1974).
6) Long-term changes in the supply of calcium in solution (Deevey, 1953; Yakushko et al., 1978; Hecky, 1978).

Some of these processes, particularly 1), 4a) and 4c-e), usually operate most efficiently near the lake margin. Others such as 2) and 4b) frequently cause significant precipitation of $CaCO_3$ in the pelagic zone. This gives rise to the phenomena known as "whitings" (Bloch et al., 1944; Müller et al., 1972; Strong, 1976). In the Afar and in NW. Europe, accumulation of deep-water marls is commonly linked with the development of dense, monospecific blooms of the planktonic diatom Stephanodiscus astraea and its varieties (Turner, 1970; Gasse, 1977b).

a. Marls

i. Occurrence

Although calcium carbonate is present in small amounts in diatomites and muds, fine-grained sediments containing $\geqslant$ 25% $CaCO_3$ are almost entirely restricted to the Bulbula formation in the Deka Wede area. These deposits have been described in chapter 6CII (Plates 39-57). Potential water depths at the Deka Wede site were 45-70 m or more. The diatomaceous marl deposits range in age from Late Pleistocene to mid-Holocene (Ziway-Shala III to VI), and $CaCO_3$ maxima are believed to correspond to lake-level highstands.

ii. Origin

Any model of the origin of the Deka Wede marls needs to take into account their localized occurrence in close proximity to the former shore; and their content of benthic and epiphytic diatoms. Marl formation is very characteristic of hard-water lakes in temperate latitudes (Terlecky, 1974; Brammer, 1978), but less is known about the processes occurring in low latitudes, particularly in $NaHCO_3$-dominated lakes. No modern analogues in eastern Africa have yet been discovered, although restricted littoral carbonate deposits are found around Lakes Kivu, Idi Amin Dada, Tanganyika, Hayq and

Awasa, which have carbonate alkalinities in the range 6-17 meq/l (Talling and Talling, 1965; Baxter and Golobitsh, 1970; Beadle, 1974; Grove et al., 1975).

For deposition of relatively pure carbonate sediments in dry climates we need to look further afield; to eastern Europe and western Asia. Modern and fossil marl deposits are found in Hungary (Muller, 1970), Turkey (de Ridder, 1965; Irion, 1973), Syria (Kaiser et al., 1973), Israel (Neev and Emery, 1967; Serruya, 1971), Iran (Hutchinson and Cowgill, 1963), Afghanistan (Förstner, 1973), Tibet (Hutchinson et al., 1943) and the U.S.S.R. (Turovskiy and Sheko, 1973; Kuznetsov, 1975). The mineral species formed include calcite, high-Mg calcite, aragonite, proto-dolomite, huntite and magnesite. All but the first three are diagenetic. Low-Mg calcite is typical of waters with an Mg/Ca ratio $\leq$ 2 (Muller et al., 1972). All the present-day Rift-floor lakes in the Ziway-Shala Basin fall in this category (Mg/Ca 0.66 - 1.02). Significant shifts in the past seem unlikely. In the Afar, where marl is the characteristic early- and mid-Holocene deep-water facies (Gasse and Street, 1978a,b), calcite, high-Mg calcite and aragonite deposits are all found, depending on the Mg/Ca ratio of the inflows into the lakes.

In the $NaHCO_3$-dominated Rift lakes, the supply of Ca^{++} is likely to be the limiting factor. Examining the list of potential mechanisms for the formation of the Deka Wede marls, two possibilities suggest themselves. These are 1) the localized entry of hot springs into the lake; and 4) the photosynthetic activity of organisms living close to the shore. Fumaroles and steam vents are found today at several sites near the foot of Mr. Alutu (Fig. 2.6). Past hot-spring activity is suggested by the strong silicification of the 1670 m lake terrace in this area (Plate 25) (Lloyd, 1977). Although increased runoff from the Deka Wede headwaters might serve as an alternative source of Ca^{++}, similar marl deposits are not found off the mouths of other streams in this area, whereas calcite precipitation is known to occur offshore from the Shala springs (Baumann et al., 1975).

It is also possible that marl formation in the Deka Wede area was mediated by photosynthetic control of pH. Of the most likely littoral organisms, green and blue-green algae tend to prefer exposed rocky shores with clear water. Significantly, no traces of algal filaments or spherulites are visible in thin sections of the marls (Plates 46-47). A dense fringe of submerged weed is suggested by the abundance of epiphytic diatoms, and by the proliferation of large pulmonates such as Lymnaea sp. in Holocene deposits (B20/14) (chapter 6CIIc). Precipitation of carbonate on submerged leaves of Potamogeton has been reported from Lake Naivasha (2-5 meq/l) by Beadle (1932). However the widespread association of littoral vegetation with relatively pure "Kertefa" diatomites (chapter 7DIIa) suggests that some more localized factor such as hot spring activity was required to initiate marl formation in this case.

iii) Rate of accumulation

A deposition rate of 1 m in 328 y was calculated for member Deka Wede 2 in the Bulbula formation (Table 7.2). Comparisons with published rates for hard-water lakes do not seem very profitable, but this figure is higher than those obtained for Holocene marls in the Afar (1 m in 592 to 2220 y). The difference is probably accounted for in part by the high silt and sand content in the middle of member 2 (Fig. 6.3), but could also reflect the difficulties of applying radiocarbon dating to lacustrine carbonates (Street, Gillespie and Delibrias, in prep.). The rates given in Table 7.2 allow us to estimate the minimum duration of episodes Ziway-Shala II, IV and V at the Deka Wede type site (Table 7.3).

b. Calcareous muds and diatomites

Fine-grained lake deposits with up to about 20% disseminated $CaCO_3$ in the form of low-Mg calcite are common in the Bulbula and Ajewa formations. They occur in both Upper Pleistocene and Holocene sequences. The associated diatom floras indicate that two distinct depositional environments are represented. These are described below:

i) Brackish-water deposits

In the Upper Pleistocene Kurkura diatomites, carbonate maxima are typical of transgressive and regressive facies (Figs. 6.6-6.8). They are associated with a littoral, brackish-water diatom flora, in contrast to the more

Table 7.3: Estimated duration of marl deposition at the Deka Wede type site (0738-B20/10C-11)

Member	Lacustral Episode	Estimated minimum duration (years)	
		Range	At average rate of 1 m in 950 y [1]
1a	Ziway-Shala IIA	410-2778	1190
1b	Ziway-Shala IIB	820-5555	2380
3a	Ziway-Shala IV	98-667	290
3b	Ziway-Shala VA	361-2444	1050
3c	Ziway-Shala VB	164-1111	480

1. Mean rate for sites in Ethiopia and Djibouti (Table 7.2)

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massive profundal diatomites in members Kurkura 3 and 4 (Table 6.3), which contain freshwater, planktonic assemblages, and are characteristically carbonate-free. This is completely opposite to the pattern of sedimentation found in the Deka Wede area.

Carbonate contents of up to 19% are also found in the lower Holocene section of the Abiyata Core (Fig. 6.8), in association with the passage from littoral, brackish-water diatoms (zone C_2) to a deep-water, oligohalobous assemblage dominated by Cyclotella ocellata, Stephanodiscus astraea and its variety minutula, and Fragilaria brevistriata (zone C_3).

The best explanation for this pattern seems to be that a belt of carbonate deposition was situated in the inner part of the offshore zone, migrating with it as the water level rose or fell. Reviewing the list of possible mechanisms (p.318), mixing, evaporative concentration or photosynthetic extraction of CO_2 seem most likely to be implicated. Hecky and Degens (1973) suggest that inorganic precipitation of $CaCO_3$ from inflowing river waters is particularly important under brackish conditions. This would explain the peripheral (but not littoral) distribution of the carbonate, which was probably formed close to the shore but transported out into deeper water by wave and current action.

ii. Pelagic deposits

During the early Holocene highstand (Ziway-Shala V), carbonate precipitation continued at a reducedrate in the centre of the lake, in association with a Stephanodiscus-dominated diatom flora (Fig. 6.8). This contrasts strikingly with the carbonate-free deep-water sedimentation during Late Pleistocene phases Ziway-Shala IIIA and IIIB; implying greater evaporative concentration, warmer water temperatures, faster photosynthetic removal of CO_2 by the phytoplankton (mechanisms 4a-c), p. 318) and/or increased inputs of calcium from the catchment.

At the II International Symposium of Paleolimnology in 1976, Francoise Gasse and I put forward a speculative interpretation for the observed pattern of carbonate deposition (Gasse and Street, 1978a,b). We emphasized two main factors: the conditions prevailing in the lake and the calcium supply. We concluded that pelagic carbonate deposition during phase Ziway-Shala V, which also occurred in Lake Abhé and the other Afar lakes, was mainly due to more rapid hydrological cycling and increased phytoplanktonic productivity compared with the cooler Late Pleistocene. As noted above, Stephanodiscus astraea and its varieties are characteristically associated with profundal marl deposits in Ethiopia and Europe (Turner, 1970; Gasse, 1977b). It is not however certain whether it is the photosynthetic extraction of CO_2 by the diatoms which results in carbonate precipitation, or whether this species is merely the one best adapted to living in lakes subject to frequent "whitings". The onset of wetter conditions and the spread of montane forest may also have increased the supply of Ca^{++} in river waters through the rapid leaching of calcareous surface soils (cf. Deevey, 1953; Yakushko et al., 1978).

c. Stromatolites

Stromatolites (also known as algal biolithites) have been defined as 'laminated sedimentary structures, built by dense mats primarily of blue-green algae which selectively trap and bind sediment particles along their mucilaginous filaments' (Garrett, 1970). Their origin has been reviewed by

Hofmann (1975) and Grove et al. (1975). Johnson (1974) has published a very useful description of the lacustrine forms occurring in the Upper Member of the Koobi Fora Formation at Lake Turkana.

In lakes, stromatolites are an essentially littoral or supralittoral facies. They require warm, well-aerated waters with low turbidity. Growth of blue-green algae is apparently encouraged by the presence of hot springs (Grove et al., 1975). Modern examples are known from rocky shores of Lakes Hayq, Awasa and Kivu (Grove et al., 1975; Beadle, 1974, p. 210). The surface waters of these lakes range in pH from 8.2 to 9.5 and in carbonate alkalinity from 6 to 16 meq/l. Although algal limestones are found in other parts of the world, notably in Great Salt Lake (Eardley, 1966) and the Lisan Formation of Israel (Neev and Emery, 1967; Buchbinder et al., 1974), they often formed in sodium chloride-dominated waters in which much higher Ca^{++} levels are possible.

Fossil examples have been found in two localities in the Ziway-Shala Basin, at altitudes of 1648 - 1670 m. They were believed by Grove et al. (1975) to be Holocene in age, but this view can no longer be accepted. The sites involved are the eastern wall of the Shala caldera and the northern shore of the Dakadima "island" (Fig. 5.9). Two types of limestone occur at Lake Shala: a honeycomb variety with vertical tubes running through it, and a second type which is browner in colour, with digitate structures. They form eroded coatings ($\leq$ 3 cm) on ignimbrite outcrops (Grove et al., 1975). At Dakadima, the honeycomb type forms crusts up to 8.5 cm thick on large rhyolite boulders, which are preserved under a layer of bedded basaltic tuff up to 1 m thick overlain by lake sediments. The nearest known source for the hyaloclastites appears to be the bevelled cone in Adamitulu Village, some 4 km to the NNE.

The evidence from Shala and Dakadima fits in well with the picture of environmental conditions suggested by studies of modern stromatolites. Samples of the honeycomb type from the Ajewa area were examined in thin section by Dr. J.W. Schopf of the Department of Geology at U.C.L.A. They show remains of algal filaments and exhibit signs of recrystallization. A few gastropod shells are present. The shoreline in both areas was exposed and rocky, with signs of past or present hydrothermal activity (Fig. 2.6 and Lloyd, 1977).

A Late Pleistocene age for the stromatolites in the Ajewa area now seems likely. A waterworn stromatolite clast was found in fluvial gravels (member UM1) at profile 74, stratigraphically below the shell bed dated 8320 ± 175 BP (Fig. 6.13). This suggests that the exposed limestones are contemporary with the buried shelly pavement in unit UM1 at P67, which has a probable minimum age of 14,400 years (chapter 6DIId). Thin tufa coatings are common on large boulders in unit UM1 between profiles 67 and 80B (1630-1651 m). At Dakadima the stromatolites also predate the 1670 m beach gravels. Moreover all the likely sources for the overlying hyaloclastites are of Late Pleistocene age (chapter 2AIII). There is therefore no good evidence for stromatolite formation during the Holocene.

d. Shell beds

i) Introduction

Beds of freshwater mollusca form a conspicuous component of the Bulbula and Ajewa formations. They have so far yielded 62% of all the ^{14}C dates from the lake basin (Appendix 2b), which makes an understanding of their mode of origin critical to the interpretation of the radiocarbon chronology.

I originally hoped to carry out a detailed analysis of the fauna of individual shell beds as a guide to the palaeooecology of the lakes. This objective was subsequently modified in the light of the ^{14}C dates from the Ajewa area (Table 6.7), which prove conclusively that the faunal assemblages reflect reworking and transport processes as well as ecological factors. As a result, a much larger number of samples is required in order to characterize each shell bed, necessitating an investigation well beyond the scope of the present study. The ecological inferences made in chapter 8 are therefore based on presence/absence data for each species in the basin as a whole (Table 8.1).

Several important features of the shell concentrations require explanation:

1) The high density of shells and shell fragments compared with adjacent beds.
2) Their great horizontal ($\lesssim$ 2 km) and vertical extent ($\lesssim$ 34 m).
3) Their restricted occurrence in space and time. (Shell beds are very much more common in Holocene than ~~in~~ Upper Pleistocene sediments.)

4) Their frequent association with reverse-graded pumice granules and pebbles.

In the following sections, several lines of evidence are used to reconstruct the depositional environments of the shell beds. These include any ecological inferences which can be drawn from the mollusca and associated remains, such as fish bones; the state of preservation of the shell material; and the general context of each bed, involving its relationships to underlying and overlying strata and to the former shore. Although the literature on modern bioclastic lake deposits is surprisingly sparse, some comparisons can be made with the fossiliferous bands found in older lacustrine formations in eastern Africa (Bishop, 1969; Van Damme and Gautier, 1972; Johnson and Raynolds, 1976; Johanson et al., 1978; Vondra and Bowen, 1978), and with surveys conducted in Lake Chad, in which dense shell banks similar to the early- and mid-Holocene examples appear to be forming at the present day (L.C. Beadle, in litt. 3/12/75).

ii) Faunal elements

The impoverished nilotic freshwater fauna of the Ziway-Shala Basin (Tables 2.7, 7.4, 8.1) has clear affinities with the mollusc faunas of the Awash and Turkana Basins. It is especially close to the Holocene member IV of the Kibish Formation (Brown, 1965; Brown and Lemma, 1970; Van Damme and Gautier, 1972). However the subfossil occurrence of a number of palaearctic species in the Ziway-Shala Basin has been noted by Brown (1973b) (chapter 6CV).

Mandahl-Barth (1954) divides the East African aquatic molluscs into several broad categories with preferences for the following habitats:

- smaller, permanent or temporary waterbodies (dams, ponds and small streams)
- papyrus swamps
- small lakes
- and the great lakes.

In a large waterbody such as the united Ziway-Shala lake, mollusca from all four categories may inhabit different subenvironments, although the second group is represented by only one species, Biomphalaria sudanica. It is therefore more appropriate to distinguish three major faunal elements found in the lacustrine sediments (Table 7.4):

Table 7.4: Ecological groupings of subfossil mollusca found in the Ziway-Shala lake sediments

1) Circumlittoral species

Pulmonata:

Lymnaea natalensis (Krauss)
Bulinus "truncatus" (Audouin)
Planorbis planorbis parenzani Bacci + p
Ceratophallus natalensis (Krauss)
C. bicarinatus (Mandahl-Barth)
C. blanfordi Brown +
Biomphalaria sudanica (Martens)
B. pfeifferi (Dunker) *
B. barthi Brown * +
Burnupia sp. +

Prosobranchiata:

Valvata nilotica Jickeli + p (?)

2) Benthic species

a) "Great lakes" element

Prosobranchiata:

Bellamya unicolor (Olivier) +

Bivalvia:

Caelatura aegyptiaca (Cailliaud) +
C. (Nitia) monceti (Bouguignat) +
Corbicula pusilla Philippi +

b) Less restricted element

Prosobranchiata:

Melanoides tuberculata (Muller)

Bivalvia:

Sphaerium hartmanni (Jickeli) +
Pisidium kenianum Preston + (? - normally a riverine species)
P. moitesserianum Paladilhe +
P. milium Held + p
P. nitidum Jenyns + p
P. subtruncatum Malm + p
Pisidium spp.n. +

Table 7.4 continued:

3) Land snails

Pulmonata:

Cerastua sp.
Cecilioides sp.

* These "species" appear to intergrade

+	Not found living in the area today	ecology inferred from other areas
p	Palaearctic	

Sources:

Pilsbry and Bequaert (1927)
Gardner (1932)
Bacci (1941, 1951)
Mandahl-Barth (1954)
Brown (1965, 1973a)
Kuiper (1966)
Brown and Lemma (1970)
Beedham (1972)
Brown and Mandahl-Barth (1973)
Beadle (1974)
Goll and Aweitu (1974a,b)
Ellis (1978)

1) circumlittoral species
2) benthic (mud-dwelling) taxa
3) land snails.

Group 2) can be subdivided into species which are restricted to large permanent lakes and rivers - the "great lakes" element — and those which may also be found in smaller lakes and/or streams.

1) The circumlittoral fauna

This group includes the pulmonate snails and probably the palaearctic prosobranch Valvata nilotica, which is common in still and slow-flowing waters in the Ethiopian Highlands (Brown, 1965, 1973b) and is frequently associated with planorbid and lymnaeid snails in subfossil assemblages.

Freshwater pulmonates are basically browsers. They spend their lives mostly on aquatic vegetation, feeding on algae, especially green algae and diatoms, or on decomposing plant tissues (Berrie, 1970). They are very seldom collective alive from depths greater than 25 m (Mandahl-Barth, 1954; Beadle, 1974, p. 237). High flow velocities (>0.75 m/s) and sediment-laden water appear to be inimical to most species (Harris, 1965; Jordan and Webbe, 1969; Berrie, 1970; Brown and Lemma, 1970).

Many pulmonates have special adaptations to a swamp-dwelling existence, such as tolerance of low oxygen concentrations and an ability to survive seasonal fluctuations in water level by aestivating (burrowing in the mud) (Jones, 1964; Jordan and Webbe, 1969, pp. 30-32). Research has concentrated on the schistosomiasis vectors Bulinus and Biomphalaria, because of their importance in disease transmission, but has at the same time revealed the habits of several other species. Biomphalaria sudanica thrives even in the anoxic waters of dense papyrus swamps (Beadle, 1974, pp. 251-253). Bulinus spp. are often found in stagnant or semi-stagnant water, or in streams with a high content of organic matter, where they feed on deposits of rapidly decaying organic detritus (Berrie, 1970, pp. 58, 62). Biomphalaria pfeifferi apparently prefers conditions intermediate between Bulinus and Lymnaea natalensis, which is restricted to relatively well-oxygenated waters and requires a diet of fresh plant material in the form of green algae, diatoms and submerged leaves. The freshwater limpets (Ancylus and Burnupia) live on stones, plant stems or leaves in well-oxygenated water, flowing or agitated by the wind as at the edge of a lake (D.S. Brown, in litt. 15/10/75).

In Lake Ziway, snails inhabit the marginal swamps (Table 2.7). Biomphalaria sudanica occupies a rather specialized and interesting niche which differs significantly from Lake Awasa. It is restricted to swampy lagoons which are sheltered from open water by thickets of Aeschynomene (Plate 16) or occasionally Typha. During the dry season, it aestivates when the level of Lake Ziway falls below about 1635.9 m, which is close to the estimated 20-year mean water level (1636.0 m) (Fig. 5.12) (Goll and Aweitu, 1974a,b; Makin et al., 1976; Goll, in litt. 29/3/75). The Aeschynomene-lagoon habitat is characteristically found where the 1636 m and 1637 m contours are closely spaced, with a gradient of 1:25 to 1:1.25. Since the life cycle of the snails is so sensitively adapted to local conditions, it is likely that this pattern evolved over a long period of time, and also existed in the early and mid-Holocene lakes.

2) The benthic fauna

The benthic (gill-breathing) molluscs include the prosobranch snails Bellamya and Melanoides and a number of bivalves. They are able to live at much greater depths than pulmonates. Mandahl-Barth (1954, pp. 187-188) records that Melanoides tuberculata, Bellamya unicolor, Sphaerium stuhlmanni, Corbicula africana and Pisidium fistulosum have all been collected live at -64 m in Lake Victoria, although he found Caelatura only down to -12 m.

The "great lakes" element comprises species which are restricted to large, permanent lakes and rivers. They are common in lakes such as Victoria, Kioga, Mobutu Sese Seko, Idi Amin Dada, Tanganyika, Tana and Chad; in the lacustrine deposits of the Turkana Basin; and in the Nile and Awash rivers. (Pilsbry and Bequaert, 1927; Mandahl-Barth, 1954; Brown, 1965; Van Damme and Gautier, 1972; Lévêque, 1972; Brown, pers. comm.). Melanoides tuberculata is also abundant in, but not exclusive to, these habitats.

Many of the "great lakes" group can tolerate turbulent open water and high flow rates. Harris (1965) found that the large freshwater mussels (unionids) could withstand current velocities of at least 2.4 m/s. The majority of benthic species feed on algae and organic detritus. Surveys in Lake Chad (maximum depth 10 m) suggests that they exhibit some preference for particular substrates. Corbicula, which likes sandy bottoms, is highly correlated with the unionid Caelatura, although the latter has a wider distribution (Lévêque, 1972). In Lake Tanganyika, Caelatura lives on sandy muds near the shore, feeding on particles of algae and organic matter stirred up by the waves (Beadle, 1974, p. 226). It is interesting to note that its larvae (called glochidia) are parasitic on fish (Mandahl-Barth, 1954, p.124). Bellamya occurs in large numbers on organic mud in the centre of Lake Chad, whereas Melanoides is not characteristic of any particular habitat (Lévêque, 1972).

The tiny orb and pea mussels (Sphaerium and Pisidium spp.) are found in a wide spectrum of low-energy habitats ranging from streams and small lakes to the bottom muds of large lakes. They may also climb onto submerged plants (Mandahl-Barth, 1954, p. 162; Ellis, 1978). Both genera are widespread in the cooler Ethiopian Highlands (Bacci, 1951; Kuiper, 1966; Brown, 1973b).

3) Land snails

Occasionally land snails are found in lacustrine shell beds. They are most commonly encountered in Holocene sediments in the Ajewa and Deka Halela embayments, northeast Shala, and in the central Bulbula Plain. But in section D738-B19/1, tiny land snails (Cecilioides sp.) occur in littoral gravels of presumed Late Pleistocene age in association with palaearctic pisidia. Given the abundance of land snails in recent colluvium (unit U3 of the Ajewa formation, for example) it seems most likely that occasional individuals were reworked by the rising lake from pre-existing slope deposits; but they could also have been transported into the lake by ephemeral streams.

iii. General ecological inferences

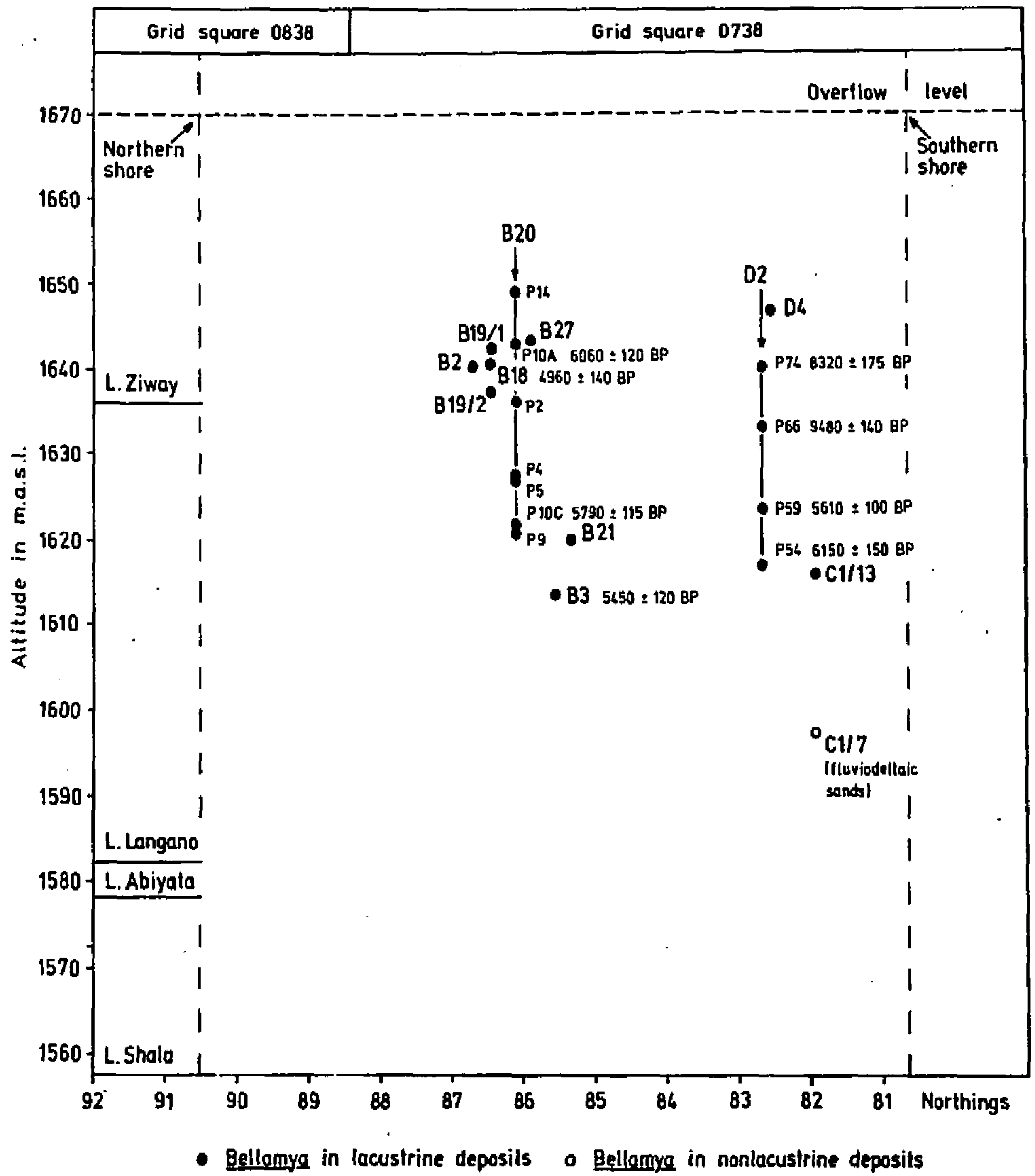
- Salinity and alkalinity ranges

Opinion is divided over the extent to which the distribution of African freshwater mollusca is affected by salinity and alkalinity. Living specimens have not been discovered in waters with conductivities (K_{20}) less than about 25 μmhos/cm, probably because of limiting levels of calcium or other electrolytes (Jordan and Webbe, 1969, chapter II; Beadle, 1974, chapter 5). Beadle and Taylor found that the minimum Ca^{++} concentration required for normal development of Lymnaea natalensis and Biomphalaria sudanica embryos was 2 mg/l (Beadle, 1974, p. 52).

Little is known as yet about the upper limit for mollusca in saline and alkaline waters, which is more relevant to this study. The most tolerant species is often said to be Melanoides tuberculata, although there is little modern data to justify this claim. Snails are apparently absent from Lake Langano (alkalinity 14.6 meq/l), while the fauna of Lake Turkana (20-25 meq/l) is stunted and depauperate. In these alkaline lake waters calcium is also reduced to very low levels, although it is not certain that this is the limiting factor. Lévêque (1972) has discovered that the density of benthic gastropods in Lake Chad drops off rapidly above a conductivity (K_{25}) of 500-600 μmhos/cm (0.4 - 0.5 ‰), well before the pulmonates begin to succumb. The most sensitive species, Bellamya unicolor, has an optimum range from 200-400 μmhos. Unionids such as Caelatura are also most abundant in fresh waters.

The limits suggested by Lévêque appear rather low in comparison with eastern Africa (Beadle, 1974, p. 171). Bellamya unicolor has been collected live from Lakes Malawi, Tanganyika, Victoria, Kioga, Mobutu Sese Seko, Tana and Abaya, which have conductivities in the range 100 - 1000 μmhos and alkalinities of 0.25 - 10 meq/l (Pilsbry and Bequaert, 1927; Brunelli et al., 1941; Mandahl-Barth, 1954; Brown, 1965; Gautier, 1970; Talling and Talling, 1965; United Nations, 1973). This species has not been found living in Awasa (5 - 11 meq/l), Idi Amin Dada (6 - 11 meq/l) or Kivu (15 - 19 meq/l). In the Ziway-Shala Basin, subfossil shells of Bellamya were only found at altitudes about 1610 m (Fig. 7.4), which suggests that a critical dilution threshold was crossed as lake level rose. On the basis of its present distribution this limit seems to lie at an alkalinity of about 10 meq/l.

Fig. 7.4: Altitudinal distribution of Bellamya unicolor in the Ziway-Shala Basin (with ^{14}C dates in years BP)



Grid square 0838
Grid square 0738
Overflow level
Northern shore
Southern shore
Altitude in m.a.s.l.
1670
1660
1650
1640
1630
1620
1610
1600
1590
1580
1570
1560
L. Ziway
L. Langano
L. Abiyata
L. Shala
B20
P14
B19/1
B27
P10A 6060 ± 120 BP
B2
B18 4960 ± 140 BP
B19/2
P2
P4
P5
P10C 5790 ± 115 BP
P9
B21
B3 5450 ± 120 BP
D2
D4
P74 8320 ± 175 BP
P66 9480 ± 140 BP
P59 5610 ± 100 BP
P54 6150 ± 150 BP
C1/13
C1/7 (fluviodeltaic sands)
92
91
90
89
88
87
86
85
84
83
82
81
Northings
● Bellamya in lacustrine deposits
o Bellamya in nonlacustrine deposits

- Temperature ranges

The influence of water temperatures on mollusc distributions is also poorly understood. Only a few species, notably Melanoides tuberculata, can thrive above 30°C (Pilsbry and Bequaert, 1927, pp. 522-23). Experiments on Biomphalaria pfeifferi have established that it reaches its highest growth and reproductive rates at about 25°C, and that continuous periods with temperatures of 25-27°C or above result in high mortality (Berrie, 1970). This species therefore sets an upper limit for lake-water temperatures during the Ziway-Shala V and VI maxima.

Brown (1973) has suggested that the subfossil distribution of so-called "palaearctic" species may reflect lower temperatures in the past. Valvata nilotica and Pisidium spp. are now found in Ethiopian highland streams but not on the Rift floor. Valvata is restricted to altitudes about 2000 m although it is very common in Holocene shell beds (Table 8.1). The variety of tiny Pisidium spp. discovered in the Upper Pleistocene and lower Holocene lake beds also indicate zoogeographical connections with North Africa and the Near East. Their subfossil distribution may reflect: 1) migration of palaearctic species down from the highlands as a result of lower temperatures; 2) transport by migrating birds from higher latitudes, which might be a side-effect of climatic change; or 3) the influence of other limiting factors such as salinity (Gardner, 1932; Kuiper, 1966; Brown, 1973). But there is no easy way to eliminate any of these possibilities.

iv. Origin

It is apparent from the literature that there are two kinds of shell beds: autochthonous and allochthonous. The former contain faunas characteristic of a single environment and have accumulated in situ. The latter show signs of reworking by streams, waves or currents; in the form of abrasion, breakage, size sorting and mixing of species from different habitats. In both lakes and seas, shells are commonly concentrated on the foreshore by wave action (Worthington, 1929; Born, 1972; Reineck and Singh, 1973; Collinson et al., 1978). In the East Turkana area this has been referred to rather ponderously as the 'arenaceous bioclastic carbonate facies' (Vondra and

Bowen, 1978). Transported shells are also characteristic of the lacustrine shoreface and delta-slope environments (Born, 1972), and may accumulate just above wave base during heavy storms (Hertweck, 1971). The latter may be the origin of the characteristic "shell zone" found at depths of 7-12 m in the Baltic lakes (Ruttner, 1963). Research by Hertweck (1971) in the Tyrrhenian Sea has shown that autochthonous (unmixed) shell assemblages are restricted to parts of the offshore zone where current action is negligible. In the context of the present study, this leads us to expect undisturbed offshore shell beds consisting predominantly of benthic species: however in the lower-energy lacustrine environment the accumulation of pulmonates in situ in marginal swamps also seems a strong possibility.

Various lines of evidence have been used to assess the degree of transport and reworking which has occurred:

- The presence of articulated bivalves is a sign that disturbance has been minimal (Van Damme and Gautier, 1972; Reineck and Singh, 1973, p. 136). Figure 7.5 shows that Corbicula and Caelatura with occluded valves are only common in early and mid-Holocene lake sediments between about 1615 and 1655 m, corresponding to potential water depths of 15-55 m. We are led to infer that most of the shell concentrations at lower elevations have been transported.

If the prominent shell bands with articulated bivalves are followed down-slope, for example from profile 10A to 10C in the Deka Wede gully (Fig. 6.2), they become thinner. Beyond a certain point the occluded valuves disappear and broken fragments become more common. This may reflect limited transport of shell material from the growth zone into deeper water by predators or current action.

- Abrasion and size sorting are characteristic of transported assemblages. In the marine environment, Driscoll (1970) has shown experimentally that destruction of bivalves occurs 150 to 1000 times faster on a sandy or gravelly foreshore than on a low-energy, muddy bottom. Size reduction is more efficient in very coarse or very fine sand than in medium sand, and increases as grain sorting (σ_{ϕ}) decreases. The most susceptible species are those with high surface area/unit weight (Driscoll and Weltin, 1973). In the Ziway-Shala

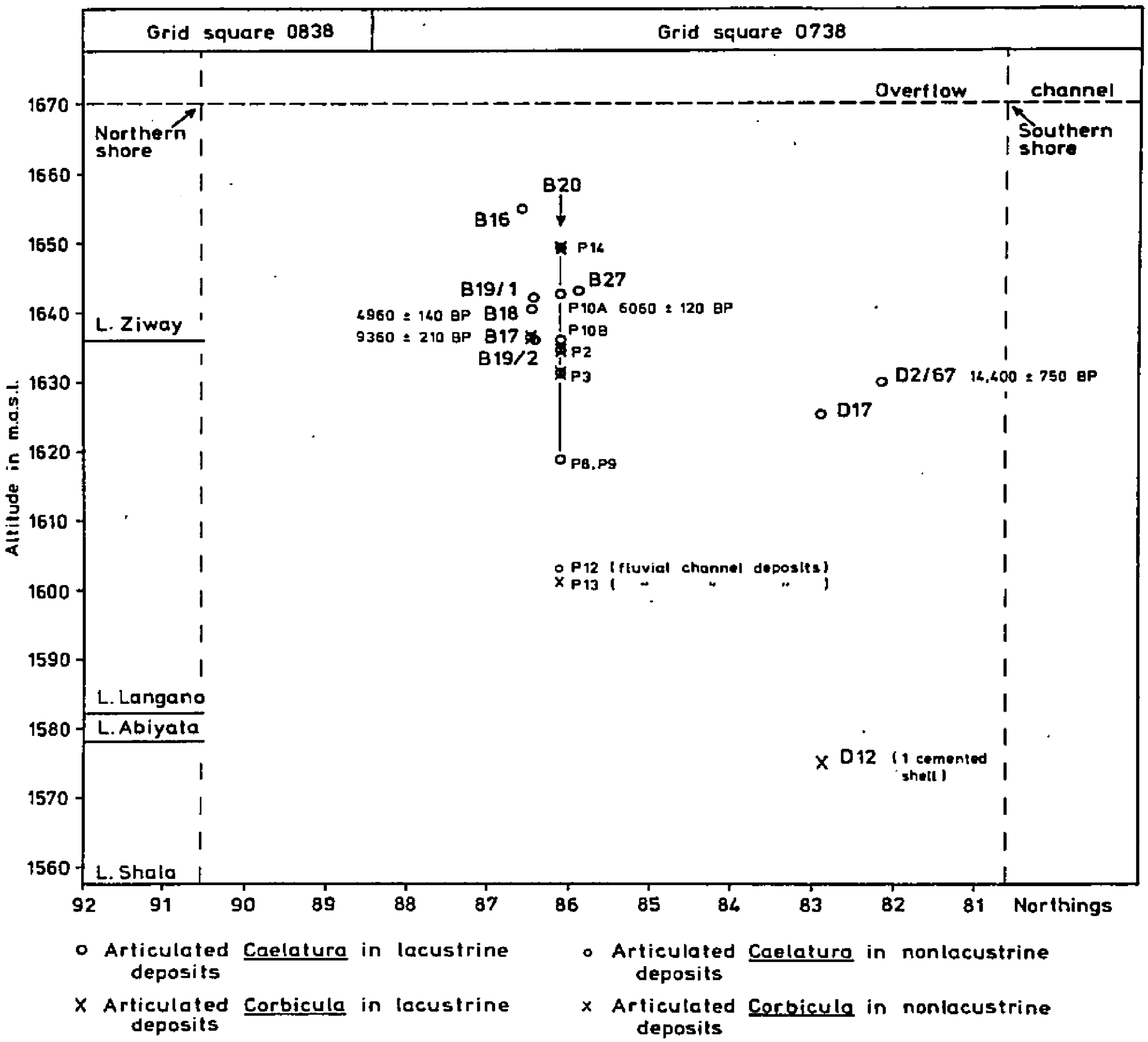


Fig. 7.5: Altitudinal distribution of shell beds containing <u>Caelatura</u> and <u>Corbicula</u> with occluded valves (^{14}C dates in years BP).

Basin, evidence for abrasion is most common in fossil assemblages from beach or micro delta deposits.

- Species composition. The relative resistance of the freshwater mollusca to reworking and wave action can be assessed from the depauperate assemblages collected on the foreshores of Langano and Abiyata. These are probably reworked from submerged sediments or by the Bulbula River. They are dominated by large, robust, relatively compact forms. Corbicula makes up 85-100% and Melanoides 0-14%. Biomphalaria pfeifferi and Bellamya are occasionally represented. The shells are opaque and rolled in appearance, with holes and broken apices (gastropods) or valves (Corbicula).
Small planorbids (Ceratophallus, Planorbis), sphaeriids (Pisidium, Sphaerium), and genera with fragile shells such as Lymnaea and Caelatura (unionid shells consist of very thin aragonite plates which disintegrate readily after death) are not represented.

The following tentative classification of the shell beds in the Ziway-Shala Basin is based primarily on the degree of reworking and secondarily on the concentration of shells and shell fragments. It does not pretend to be more than a preliminary approach to a very complex problem.

1) Assemblages with only minor signs of reworking

a) Rich concentrations

These are of two types: those in which benthic species account for a significant proportion of the biomass, and those in which pulmonates are dominant. They appear to have formed in open water and in marginal swamps, respectively. Although articulated bivalves are a characteristic of the first type, none of these beds are truly autochthonous, since there is always some mixing of faunal elements.

- Shell banks formed in open water. These are the very rich and laterally extensive deposits which are plotted on Figures 7.4 and 7.5. They almost all occur in one of two stratigraphically distinct situations; either near the base of the Holocene lake beds or in coarse gravels of mid-Holocene age (Plates 77 and 78). The basal beds formed during the Ziway-Shala V transgression as the lake became rapidly fresher. They characteristically occur in thin-bedded silts, sands and gravels which were probably deposited in the inshore zone, and are overlain by fine-grained silts or diatomites.

The association of shell beds with lake transgressions is not coincidental. Studies of artificial lakes have shown that temporary, explosive increases in plant and animal populations commonly occur during the filling phase. This is due to the influx of nutrients resulting from the decay of flooded vegetation, especially trees; the leaching of soils; and shoreline erosion (McLachlan, 1974, 1977). When filling is completed, productivity declines towards an equilibrium level determined by the supply of nutrients and their recycling within the lake (Mitchell, 1973).

The presence of articulated bivalves in thick accumulations of coarse gravels is more surprising. Figure 6.1 reveals that the Holocene lake deposits in the Bulbula Plain south of the Hondola sill are extremely thin. Fine silts and clays of early Holocene age are preserved in the Bulbula Village (0738-B8) and Abiyata Core sections (Fig. 7.3), but no deep-water deposits of mid-Holocene age have been found. We can infer that fine sediment was prevented from accumulating in this area, perhaps by bottom currents. Beadle (1974, p. 178-179) has described the formation of shell pavements by current winnowing at depths of 20-30 m in Lake Tanganyika. The mid-Holocene shell beds in sections 0738-B18, B27(?) and B3 may be "condensation deposits" (Reineck and Singh, 1973, p. 136) of this type. However the consistent radiocarbon dates of 5300-5550 BP from B3 (Table 6.3) indicate that the accumulation of shells occurred within a fairly short time span.

- Shell beds formed in marginal swamps. Dense concentrations of pulmonate snails, including the very large "swamp-dwelling" form of *Lymnaea natalensis* (p.216), are found in the Deka Halela embayment, northeast Shala. Fish bones, crab remains and nodules of bog iron (Pilsbry and Bequaert, 1927, p. 529) occur with the shells. The fauna suggests sheltered but not completely stagnant conditions. These beds have not yet been dated, but are believed to be mid-Holocene in age.

b) Sparse concentrations

Scattered shells, particularly articulated Caelatura and Corbicula, are common in permanent stream-channel deposits within the Bulbula formation. Shell horizons with pisidia and small pulmonates are also present in lacustrine beds of Upper Pleistocene and lower Holocene age in sections 0738-B2 (Fig. 6.11) and D13. The reasons for the predominance of tiny, fragile species in these lacustrine assemblages are uncertain, but we can infer low-energy conditions with sparse aquatic vegetation (Van Damme and Gautier, 1972). The presence of several palaearctic species make cooler water temperatures conceivable.

2) Reworked or water-sorted assemblages

a) Nonlacustrine deposits

Freshwater mollusca are occasionally found in colluvial or fluvio-deltaic deposits. A good example of the former is unit U3 in the Ajewa formation. Reworked shells are also present in the delta foresets at site 0738-C1/7 (Plate 35). In both situations the shells are infrequent and usually broken or rolled. Melanoides, Bellamya, Bulinus and Corbicula are most abundant.

b) Beach deposits

Concentrations of abraded shells are common in sandy or pebbly deposits close to the highest shoreline. The best example is site 0738-D1 in the Mirrga graben, which lies within a metre or two of the overflow level. As predicted earlier, Melanoides is one of the most resistant taxa, although Bulinus is also very common here. The shells show typical abrasion features similar to those found in modern beach deposits.

In this kind of exposed situation the danger of reworking of older shells is particularly high. This is probably the explanation for the discordant dates which were obtained by Haynes and Haas (1974) on individual species fractions from sandy, shallow-water deposits at a single site on terrace III, West Ziway, at an altitude of 1653 m.

c) The prominent shell beds in the Ajewa formation

The discontinuous basal shell concentrations in units D3 and UM4 of the Ajewa formation have been described in detail in chapter 6DIIb. The principal argument for reworking is the distribution of radiocarbon dates (Table 6.7, Fig. 6.13), but there is considerable internal evidence to confirm this view. The shell lenses contain a wide mixture of freshwater benthic and pulmonate species with occasional land snails. Lymnaea is very rare and the unionids (Caelatura spp.) are often fragmented, especially at the upstream end where the early Holocene ^{14}C dates were obtained. The lateral extent of the shells and their presence in a discrete, gravelly layer within otherwise fine-grained deposits suggests that they were emplaced by a widespread and catastrophic event, which also resulted in massive fish mortality. In chapter 6DIIb it was suggested that a dramatic fall in lake level fits in best with the field evidence; but the problem cannot be regarded as conclusively resolved.

The likelihood of reworking makes it difficult to make ecological inferences from the fauna in the Ajewa concentrations. However, if we accept the hypothesis that the mollusca which yielded mid-Holocene dates are contemporary, or nearly so, with the formation of the shell beds, then we can deduce that conditions along the southern shores of the Ajewa embayment were suitable for Biomphalaria sudanica, which implies the existence of shallow lagoons protected by Aeschynomene thickets. Subfossil shells of this species have been found in the area by a number of collectors (Brown, 1965), and a single large specimen was found in situ at P54. It had undergone a second growth spurt following aestivation (P. Goll, pers. comm. 1974). This suggests that the delicate ecological adjustment between the life cycle of this snail and seasonal fluctuations in lake level was already in existence at this time.

iv. Conclusions

1) It is now evident that the shall concentrations in the Ziway-Shala Basin have originated in a number of different ways. Reworking is probably a greater threat to the validity of ^{14}C dates on shell than contamination by younger carbon during recrystallization (Appendix 2c). The most reliable beds

for dating purposes seem to be the prolific concentrations in the Bulbula Plain which contain articulated bivalves, although these samples will only yield minimum estimates for the contemporary lake level. Their reliability is supported by the consistent dates from 0738-B3 (Table 6.3). In chapter 8B, the ages derived from reworked shell concentrations are only used where they do not conflict with other evidence.

2) Because of the problems posed by reworking, palaeoecological inferences based on the mollusc faunas are necessarily limited. The most useful indicators appear to be the following:

a) Pulmonate snails, which are browsers, suggest the presence of a fringe of aquatic vegetation. B. sudanica is restricted in Lake Ziway to the lagoons which form in the lee of Aeschynomene thickets.

b) The large benthic taxa Bellamya unicolor and Caelatura spp. are found only in major freshwater lakes or rivers. The upper alkalinity limit for B. unicolor in East Africa is about 10 meq/l.

c) Biomphalaria pfeifferi cannot tolerate water temperatures much higher than 25°C.

3) The very prolific concentrations of subfossil benthic mollusca suggest an aquatic ecosystem with high net primary productivity. The enormous molluscan biomass in Lake Chad, which locally exceeds 1000 - 2500 kg/ha (Bellamya alone accounts for 34 - 45% of this total), is maintained by an abundance of organic detritus and warm, shallow, wind-stirred waters (Lévêque, 1972; Beadle, 1974, p. 171). Such conditions may well have prevailed during the early and mid-Holocene on the gently shelving lake bottom in the Bulbula Plain and West Ziway areas, to lakeward of the fringing swamps. It is difficult to account for the apparent lack of such rich shell concentrations in the Upper Pleistocene lake beds: since it seems unlikely that calcium levels were low enough to be limiting, we can only guess that lower water temperatures and/or lower primary productivity were responsible.

CHAPTER EIGHT

Late Quaternary Environments in the Ziway-Shala Basin

8A SUMMARY

Chapter 8B brings together the information derived from the study of the lake shorelines and sediments, in order to derive an integrated picture of environmental change through time. The sequence of lake-level fluctuations is reconstructed from three lines of evidence:

1) The elevations and relative ages of the major shorelines (chapter 5).
2) The combined stratigraphic sequence derived from key sections (chapters 6 and 7) (Table 6.9).
3) An analysis of all the ^{14}C dates from the lake basin with respect to elevation (Fig. 8.1). These only cover the last 28,000 years.

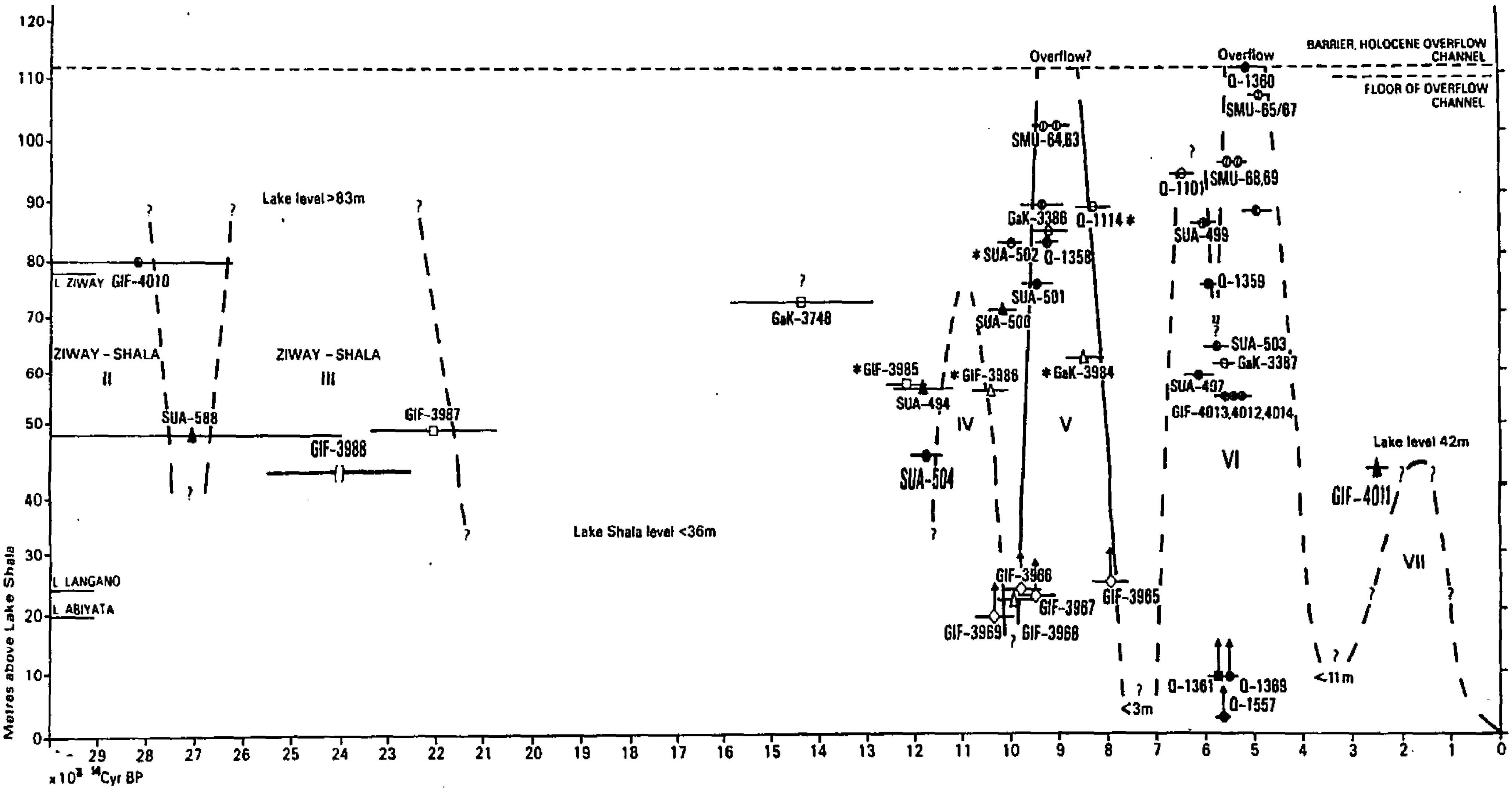
Chapter 8C discusses current evidence on past snowlines, vegetation changes and soil formation in the Ziway-Shala Basin, which is relevant to the selection of palaeo-temperatures and runoff coefficients for use in chapter 9. The implications for the basin as a whole are summarized in Table 8.1, side by side with the lacustrine evidence.

8B THE LAKES

I. Introduction

This survey of the changes in lacustrine environments through time will concentrate on two major aspects: water depth and salinity. Lake area, which can readily be calculated for a given shoreline elevation once the bathymetry is known, is a direct measure of water balance. Salinity, which is an indicator of dilution by rainfall and runoff, can be reconstructed from independent diatom and molluscan evidence, and therefore serves as a check on the results. In the following chapter a simple water-budget approach will be used to convert the variations in lake area into estimates of past rainfall.

Fig. 8.1: Water-level curve for the Ziway-Shala Basin (fluctuations of Lake Shala) since 30,000 BP.



¹⁴C Control: Key

Lacustrine sediments

- O Mollusc shells
  - Haynes and Haas (1974)
  - Gèze (1975)
  - Grove et al (1975)
  - This study
- □ Other carbonates
- ◇ Organic matter
- ■ Fish bones

} This study

Non-lacustrine sediments

- ● Fluviatile mollusc shells
- ▲ Charcoal
- △ Soil organic matter

} This study

Error bars are $\pm 2\sigma$

↑ Minimum elevation for contemporary lake-level

* Date inconsistent with curve

Chapters 5 and 6 discussed the geomorphic and sedimentary evidence for fluctuations in lake level. The stratigraphic model derived from the study of individual sections is summarized in Table 6.9, and forms the framework of the present chapter. Its reliability is broadly confirmed by the water-level curve shown in Figure 8.1, which is based on a total of thirty-nine ^{14}C dates obtained by all three groups who have worked in the area, and takes into account all the dated sites in the basin (Appendix 2b). Published altitudes for the S.M.U. samples have been corrected using the more precise levelling survey by L.R.D. (Makin et al., 1976). The lake-level curve plotted is a best-fit envelope to the lacustrine samples. In places there are large discrepancies between different dating materials ($\leqslant$ 1750 y). It is now quite obvious that the response time of the lake was shorter than the combined errors inherent in ^{14}C dating. The solution adopted here is to assume that charcoal is the most reliable, followed in descending order by the organic fraction of lake sediments, shells, marl, tufa and soil humus (Olsson, 1968; Thurber, 1972; Stuiver, 1964; Morrison and Frye, 1965; Benson, 1978).

Figure 8.1 contributes little to our understanding of Late Pleistocene lake-level fluctuations, because the dating framework for this period is inadequate. It shows only the broad outlines of phases Ziway-Shala II and III. The major lowstand which followed has been called the "terminal Pleistocene arid phase" (Gasse and Street, 1978a,b; Gasse et al., in press; Street, in press). This is not strictly correct, since it is now known that the "Holocene" lacustral phase Ziway-Shala IV had begun and ended before the beginning of the Holocene, which is now usually taken to be 10,000 BP (Appendix 2a). However the existing terminology will be retained here in order to avoid confusion.

The lake-level curve confirms the complexity of the Holocene fluctuations reconstructed from the Deka Wede section, though even now the Bulbula IIIB and IIIC oscillations are not resolved within the early Holocene peak (Ziway-Shala V). The distinctness of the Ziway-Shala IV/V regression is only apparent from detailed comparisons of the Deka Wede and Ajewa sections with the Abiyata Core. It was not shown on earlier verions of this curve

(Gasse and Street, 1978a,b; Street, in press,a). The form of the Ziway-Shala VI peak is based on the new evidence from the Roricha fish bed (chapter 6DII).

The S.M.U. dates from the Ziway terraces show that the lake probably overflowed during at least part of phases Ziway-Shala V and VI. It therefore occupied the 1670 m level at least twice. The earlier rise (Ziway-Shala IV) does not appear to have reached the outlet.

The small late Holocene Ziway-Shala VII transgression is recorded by the 1595-1600 m shorelines and by beach and delta deposits at 1600-1601 m in the Ajewa, Deka Halela and Rukesa sections. Ziway-Shala VIII, which is not shown on Figure 8.1, corresponds to Nilsson's, and possibly Omer-Cooper's, drowned trees, and to the +6 m shoreline of Lake Shala.

The complete sequence of lake-level fluctuations is summarized in Table 8.1, together with the characteristics of each palaeolake. The dates given are based entirely on evidence from within the basin. In the following sections this sequence will be reviewed in more detail, beginning with the first lacustral phase which can be related stratigraphically to the fluctuations of upper Late Pleistocene time (Ziway-Shala I). The fragmentary evidence for high lake-levels of Middle Pleistocene or lower Late Pleistocene age in the Gademotta-Ziway area has already been discussed in chapters 3 and 5C and will not be repeated here.

II. Characteristics of the Late Quaternary lakes

a. Ziway-Shala I

This early lake is represented by the finely laminated, faulted diatomite (units D1 and DM1) on the northeast side of Lake Shala. This apparently records the first flooding of the newly-formed caldera. We do not know if the lake communicated with the Abiyata Basin or if the water-level rise had any climatic significance. The diatomite outcrop extends to at least 1602 m. The preservation of fine laminae implies that the deep water was seasonally or permanently anoxic (chapter 6DII). Allowing for an oxygenated layer of 50 - 100 m, this suggests a surface elevation of 1650-1700 m or more.

Sediment inputs were very low, perhaps because the Langano-Abiyata-Shala isthmus was flooded, perhaps because of different conditions in the catchment, which may at that stage have been very small, steep and rocky; or, alternatively, because the lake waters were sufficiently alkaline to cause immediate flocculation of clays. However, the pH must have been lower than today in order to allow the preservation of diatom frustules (chapter 2EIV).

No faunal remains have so far been found in this diatomite. Mollusca and ostracods could easily have been destroyed by solution. The diatom floras have not yet been studied, but the rhythmic laminations suggest seasonal variations in water temperatures and diatom productivity. Longer-term changes in the alkalinity of the lake are indicated by their highly variable calcite content (chapter 7DIIb).

The minimum duration of this episode, as estimated from the thickness of the couplets, is 4800 - 5000 years. It took place after deposition of the caldera-wall ignimbrites (dated 0.20 - 0.24 m.y.), perhaps as recently as 100,000 years ago.

b. The Ziway-Shala I/II regression

In the Shala Basin, the top of the laminated diatomite has invariably been removed by erosion. It is cut by deep channels resembling the present Ajewa gully in cross section. Some of these are infilled with diatomite gravel or calcareous colluvium.

In the Abiyata Basin this period is apparently represented by all or part of the 36.6 m of fluvial sands and diatomaceous shallow-water deposits which underlie the "Kurkura" diatomites in the Abiyata Core. These must represent a long interlacustral interval. Using the average rates of deposition given in Table 7.2, we can estimate its duration as 41,000-47,000 years or more.

During this period coarse gravels appear to have been deposited on alluvial fans in the Alutu piedmont (sections 0738-B20 and B27). Like the other evidence cited above, they conjure up a picture of rather harsh, semi-arid conditions.

c. Ziway-Shala II

This lacustral phase was first defined in the Deka Wede type section (= Bulbula I of Gasse and Street, 1978a). It represents the initial, hesitating transgression of the upper Late Pleistocene lake. The deposits of member Deka Wede 1 consist of fine, ostracod-bearing sands, which are weakly laminated in places (Fig. 6.3). The slight regression evident between member 1a and 1b is more clearly seen in the Kurkura section 0738-B27, where there is an erosional unconformity separating the two lowest lacustrine units (Fig. 6.7). If we assume that the *Corbicula* shells from the Gerbi section date from the second, more important, fluctuation (Fig. 8.1), then it appears that lake level reached at least 1609 and 1638 m during phases Ziway-Shala IIA and IIB. These estimates correspond to lake areas of roughly 1400 and 2000 km^2 respectively (Fig. 8.2). Ziway may have been united to the other lakes during phase IIB but probably not during phase IIA.

In the Kurkura section 0738-B27 the two maxima are represented by very fine pumice sand and impure diatomite. In the Ajewa area, the lateral equivalent of member Deka Wede 1 may be the fluviodeltaic sands and overlying shallow-water diatomite in units DM3a and 3b. However only one lake-level fluctuation is evident here.

The total duration of episodes Ziway-Shala IIA and IIB seems to have been short: probably of the order of 1200 and 2400 years respectively (Table 7.3). If the ^{14}C dates from the Deka Wede section can be relied on, then the onset of the long upper Late Pleistocene lacustral period (phases II and III) can be tentatively placed around 30,000 BP or shortly before. If the new date of 28,200 BP on sample GIF-4010 is also a true reflection of the age of the shells and not an artifact of contamination (cf. Butzer *et al.*, 1969), then the Ziway-Shala IIB maximum occurred around 29,000 to 28,000 BP.[1]

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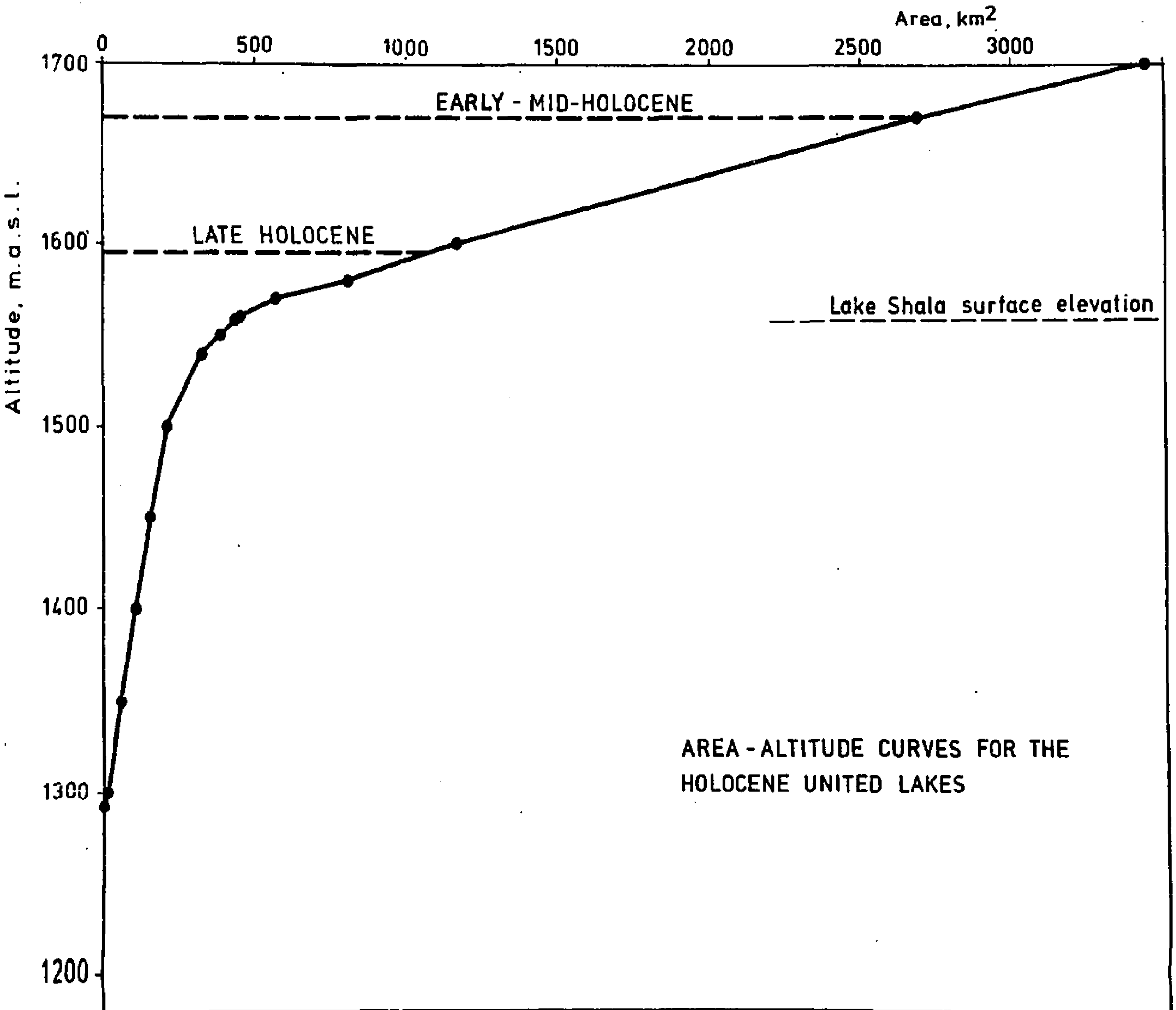
1. Note added in proof:
A date of 24,590 ± 550 BP just received for the top of the Kurkura diatomite sequence in section 0738-B8 (corresponding to the end of phase Ziway-Shala IIIC) indicates that the Deka Wede dates may be too young, and that the chronology of these Late Pleistocene highstands is still open to question.

Fig. 8.2: Area-altitude curve for Lake Ziway-Shala (data from U.S. Defense Mapping Agency 1:250,000 maps and sources listed in Fig. 2.8).


Area, km2
0
500
1000
1500
2000
2500
3000
1700
1600
1500
1400
1300
1200
Altitude, m.a.s.l.
EARLY - MID-HOLOCENE
LATE HOLOCENE
Lake Shala surface elevation
AREA - ALTITUDE CURVES FOR THE
HOLOCENE UNITED LAKES

Little is known about ecological conditions during phase Ziway-Shala II. Member Deka Wede Ib contains few diatoms. Its flora suggests a littoral, brackish environment, whereas the assemblage from the base of unit DM3b is similar to the Holocene freshwater Kertefa diatomites (chapter 7DIIa; Appendix 4b). Since diatom studies of the 19.4 - 14.2 m levels in the Abiyata Core, which are believed to correlate with phase Ziway-Shala II, have not yet been completed, it is not yet possible to resolve this controversy by reference to the core sequence. However, the observed scarcity of mollusc and fish remains in sediments of this age suggests that the Ziway-Shala II lake was rather brackish, or at best, of fluctuating salinity and alkalinity.

d. The Ziway-Shala II/III regression

At the end of phase Ziway-Shala IIB, lake level dropped sharply to 1591 m or below (from Kurkura section 0738-B3). This event can be provisionally dated between 28,200 $\pm$ 1000 and 27,050 $\pm$ 1540 BP. The preliminary diatom analyses published by Gèze (1975) confirm that Abiyata ceased to overflow into Shala and became brackish. In the Ajewa section the shallow-water diatomite which forms the base of unit DM3b is overlain by a thin layer of pumice gravel with diatomite clasts, which seems to record a brief spell of fluvial reworking of the underlying sediments.

In the Alutu piedmont, the exposed Ziway-Shala II lake deposits were also partially dissected. In the Deka Wede valley, a thin soil developed on ephemeral stream gravels around 27,000 BP. The area was occupied by prehistoric hunters who probably obtained obsidian for tool-making from Mt. Alutu.

e. Ziway-Shala III

Shortly after 27,000 BP a renewed transgression flooded the Bulbula Plain. The lacustral phase Ziway-Shala III (= Bulbula II) was originally defined in section 0738-B20, where member Deka Wede 2 consists of up to 1.85 m of shelly, fluviodeltaic sands overlain by 6.45 m of marls and calcareous sands. The marls have been dated 24,000 $\pm$ 750 BP close to the base and 22,050 $\pm$ 750 BP at the top. The apparent hiatus in sedimentation between 27,000 and 24,000 BP may be the result of slight contamination of the basal marl with younger carbon, although thin-section study does not reveal any visible recrystallization of the carbonate matrix. The estimate of 6000 – 8500 y for the total duration of phase III, based on the thickness of the Kurkura diatomites (Table 7.3), is probably more realistic.

The Deka Wede type section indicates that there were two major high-stands, Ziway-Shala IIIA and IIIB. Phase IIIA lasted considerably longer than IIIB. Comparison with the Kurkura sections (Table 6.4) suggests that lake level reached at least 1617 and 1622 m respectively. The sediments deposited in the southern Bulbula Plain were noncalcareous diatomites in both cases. A third oscillation, Ziway-Shala IIIC, is represented by shallow-water silts and beach deposits in the Kurkura area (0738-B27), which overlie fluvial sands and gravels resting on an eroded diatomite surface.

In the Shala Basin, the Ziway-Shala III lacustral phase appears to be recorded by pale yellow, diatomaceous mudstones (unit DM3b), again with a low carbonate content. A water-laid pumice layer up to 65 cm thick near the base may be derived from Chabbi rather than Alutu, since it is only found in the Shala caldera.

There is some circumstantial evidence to suggest that lake level reached about 1680 m during phases IIIA and IIIB. The diatom flora of both these episodes is rather similar and resembles Ziway-Shala VI, when the united lake overflowed through the Meki-Dubeta col at 1670 m. The sedimentary facies also indicate stable, deep-water conditions. There are eroded terrace remnants at about 1680 m west of Ziway Town which tie in with the evidence from the overflow channel for a pre-Holocene outlet at 1675-1680 m. And lastly, there are indications that the algal stromatolites found at elevations of 1648-1670 m are of pre-Holocene age (chapter 7DIIIc).

The Ziway-Shala IIIC lake appears to have reached a final maximum of about 1641 m. This estimate is based on the altitudes of the shelly beach gravels which immediately underlie the Abernosa pumice member in the northern Bulbula Plain, at altitudes of 1638-1641 m (sections 0738-B17 and B19/1). These have not yet been dated.

During phases IIA and IIB, the lake therefore covered an area of the order of 3000 km^2, uniting all four major waterbodies (Fig. 8.2). This figure is necessarily approximate, because the bathymetry has been changed by sedimentation and faulting. The palaeolake had a maximum depth of at least 388 m in the South Basin, which is about 122 m deeper than the present Lake Shala. The North Basin had a maximum depth of at least 54 m. In the central Bulbula Plain, which was then less of a bottleneck than it had become by the early Holocene, water depths were of the order of 65-90 m during phase IIIA.

The Ziway-Shala IIIC palaeolake was not so extensive - about 2000 km^2 in area - and probably did not overflow. Maximum water depths were at least 349 m and 15 m respectively in the two basins.

The diatom floras in Kurkura section 0738-B3 and in the Abiyata Core have been studied in some detail (Figs. 6.6, 6.8), whereas only pilot samples from the Deka Wede (member 2a) and Ajewa sections (unit DM3b) have been examined by Dr. Gasse. At all four sites the characteristic association during phase IIA consisted of Cyclotella ocellata with numerous tropical or endemic Melosira spp. and Fragilaria brevistriata (Gèze, 1975; Gasse, 1975, p. 369; Descourtieux, 1977; Gasse and Descourtieux, in press; F. Gasse, pers. comm. 3/5/76). During the IIIA/IIB recession conditions in the Abiyata Basin became brackish. The IIB highstand was very similar to IIA, but a conspicuous "nordic" and "alpine" element was present in the flora ($\lesssim 5\%$).

The Ziway-Shala III palaeolake had quite a diverse mollusc fauna. At present large benthic species such as Bellamya appear conspicuously lacking, although none of the samples collected so far correspond to the IIA and IIB maxima. The IIIC assemblage from 0738-B19 consists largely of the tiny palaearctic bivalve Pisidium subtruncatum Malm. Pisidium spp. are also found at the top of the diatomaceous mudstone in section 0738-D13, together with a number of rather frail-looking pulmonates.

Viewed together, the diatom and mollusc evidence suggests a large, stable, oligohaline lake with medium-low alkalinity (2-10 meq/l), high dissolved silica in its surface waters (>10 mg/l), and only moderate organic productivity

(Richardson, 1968, 1969; Kilham, 1971b; Gasse, 1977b; Gasse and Descourtieux, in press; Richardson et al., 1978). Melosira spp. are abundant today in a range of oligohaline lakes including Naivasha (1884 m.a.s.l.), Tana (1830 m), Victoria (1134 m), George (916 m) and Gamari (340 m), all of which have surface temperatures of 19-20^{o}C or above (Gasse, 1975). Although Melosira spp. can tolerate quite shallow habitats with high turbidity (Talling, 1976b), the thickness of the diatomites and the low percentages of littoral species suggest prolonged deep-water conditions in this case.

Fine-grained carbonates were deposited in the inner offshore zone during the IIIA transgression, but calcite precipitation in the centre of the lake basin ceased during the IIIA and IIIB maxima, presumably because the lake was overflowing and therefore too dilute (chapter 7DIIIb). However localized marl deposition in the Deka Wede embayment culminated during these high-stands, probably due to an enhanced input of calcium from the springs at the foot of Alutu. Algal stromatolites formed on exposed rocky shores where the waters were clear and well aerated. There is some evidence that hot springs encouraged algal growth (chapter 7DIIIc).

It is quite possible that surface water temperatures were cooler than present during phases IIIB and IIIC, perhaps by as much as 4-5^{o}C. This would explain the small but significant number of nordic and alpine diatoms and the appearance of palaearctic pisidia. However diatoms should be used only with caution as temperature indicators, because of the much greater influence of factors like alkalinity, dissolved silica, and trophic status (Richardson et al., 1978). It appears from the rarity of laminations in the Kurkura diatomites that the sediments in the Bulbula Plain were in contact with oxygenated bottom water, although their present low organic content may reflect oxidation during the terminal Pleistocene lowstand.

The sparse mollusc faunas suggest that littoral productivity was low compared with the early and mid-Holocene. The infrequent, frail pulmonates indicate a rather patchy fringe of weed around the lake. Possibly the absence of Bellamya and the dominance of small benthic species reflect a general scarcity of organic detritus, although biogeographic factors such as the existence of a water connection with the Awash system may also have influenced species diversity.

f. The Ziway-Shala III/IV regression

The terminal Pleistocene arid period which followed was the most pronounced in the last 30,000 years. Lake level fell dramatically some time after 22,000 $\pm$ 750 BP to levels at or below present. Lloyd (1977, p. 56) has even suggested that Lake Ziway dried out completely.

The first indication of environmental deterioration may be the influx of reddish sediment during phase IIIB, which coloured the diatomite of member Kurkura 5 a distinctive pinkish-orange. The most likely source are the latosols in the Gidu and Langano drainages. Fine red silt from the Gidu can be seen entering Lake Shala in Plate 7. The high turbidity of Lake Langano at the present day has been attributed to deforestation and soil erosion on the Southeastern Escarpment (Wenner, 1973b; Grove et al., 1975; Lloyd, 1977).

During the IIIC highstand conditions were fluctuating and unstable. The final retreat of the lake in the Abiyata Basin is recorded by thin ostracod marls and layers of pumice gravel in the Kurkura (0738-B3) and Bulbula Village (0738-B8) sections. These seem to reflect an oscillating, and possibly brackish, environment like the present north shore of Abiyata, since mollusca are very scarce.

During the succeeding nonlacustral interval up to 13 m of air-fall pumice and ash accumulated on the exposed lake floor and on the slopes of Mt. Alutu (the Abernosa pumice member). The volcano apparently entered a more explosive phase at about the time of the IIIA/IIIB regression, when thick water-laid ashes appear in the Kurkura sections (Gèze, 1975). After the lake retreated the pumice was able to accumulate where it fell. A minimum of thirteen eruptive events is recorded in 0738-B27. Brownish, calcareous, weathering horizons are present, but soil formation was minimal compared with Holocene pumice deposits (Haynes and Haas, 1974; Laury and Albritton, 1975). For example there was little translocation of clay. This may indicate that local conditions were rather dry (Gasse and Street, 1978a). However it must not be forgotten that the frequent eruptions may have had injurious effects on the vegetation, due to acid rains, physical smothering, or to the high permeability and low nutrient status of rhyolitic pumice (Wilcox, 1959; Haynes and Haas, 1974).

Contemporary fluvial activity is indicated by shallow, gravel-filled channels within the Abernosa member (section 0738-B17, Fig. 6.10). Towards the close of this period, the drainage system in the Bulbula Plain became incised almost to the level of the modern gully floors, suggesting that Lake Abiyata had shrunk to approximately its present extent. Primary air-fall pumice is not found in the Abiyata Core, which seems to have a depositional hiatus at this level.

In the Shala Basin, the minimum level of the lake cannot at present be established. The III/IV interval is represented by up to 4 m of gritty loams with calcrete horizons. These colluvial/alluvial deposits are hard to interpret. They show increasing signs of water sorting downslope, and contain a mixture of unrolled and waterworn artifacts, but apparently no land snails. Possibly the latter have been destroyed by solution. These deposits probably accumulated through unconcentrated surface wash acting under rather open vegetation conditions, since recent overgrazing is resulting in the localized formation of a pebbly surface mulch (Fig. 6.13). Towards the close of this interlacustral period, increased geomorphic stability allowed the formation of a soil on top of the slope deposits.

g. Ziway-Shala IV

It is not certain when lake level first began to rise again. By 11,800 $\pm$ 150 BP the water-table in the Bulbula Plain had recovered sufficiently to allow large unionids to grow in permanent streams at the foot of Alutu (profile 0738-B20/12). The lake subsequently flooded the Deka Wede embayment, depositing a thin layer of marl over the valley sides. Sedimentation conditions in the Deka Wede "ria" were very calm and sheltered. Disseminated charcoal and soil humus from the palaeosol underlying the marl have been dated 11,870 $\pm$ 300 BP and 10,450 $\pm$ 180 BP respectively. The basal marl itself yielded an age of 12,200 $\pm$ 200 BP. In a case of conflict like this, it is usual to accept the charcoal date as the most reliable (Thurber, 1972), but it must be borne in mind that the charcoal was scattered throughout the top 10-20 cm of the soil, and may therefore be a little older than the time of inundation (Blong and Gillespie, 1978). The marl date is probably a minimum of 300 y too old, due to a contribution of dead carbon from the spring waters which fed the lake in this area (Broecker and Walton, 1959; Thurber and Broecker, 1970; Thurber, 1972). Slight contamination of the soil by younger humus or roots would easily explain its measured age.

A compromise date of 11,500 BP will be provisionally accepted for the initial transgression of the lake at 1615 m. Lacustrine conditions apparently lasted only a few hundred years at this elevation (Table 7.3).

In the Shala Basin the IVA episode is represented by the swampy lake-margin deposits of unit UM3 in the Ajewa formation. It is not certain how high the lake rose. The hydromorphic soil in unit UM3 extends to at least 1632 m, although recognizably lacustrine sediments are not found about 1623 m. This implies a lake of the order of 1900 km^2 (Fig. 8.2). Shala, Abiyata and Langano must have been united, but probably not Ziway.

The diatom assemblage in the Abiyata Core indicates that the lake freshened rapidly to a salinity on the oligohaline-mesohaline borderline, with alkalinities ranging from medium-low to medium-high ($<$ 10 to 50 meq/l) (Gasse and Descourtieux, in press). This is borne out by the mollusc fauna of unit UM3 in the Ajewa formation. Only a few, tolerant taxa such as Melanoides and Bulinus are present, and neither Bellamya nor unionids are included.

Several independent indicators confirm that a fringe of aquatic vegetation existed at least locally in the Bulbula Plain and the Ajewa embayment. A variety of epiphytic diatoms are found in the Deka Wede section (member 3a). Many of them are common on submerged weed in Lakes Ziway, Tana and Hayq (1-10 meq/l) (Gasse, 1975; Baxter and Golobitsh, 1970). The small but significant pulmonate element in the mollusc fauna, and the characteristic swamp soil developed in unit UM3 of the Ajewa formation, tell the same story.

h. The Ziway-Shala IV/V regression

The Ziway-Shala IV maximum ended abruptly between 10,370 ± 180 BP, the age of the top of the clayey silt bed in the Abiyata Core, and 10,220 ± 140 BP. The latter date was obtained from a large piece of carbonized *Acacia* wood in unit UM3 of the Ajewa formation. Lake level fell below 1579 m at the site of the Abiyata Core, and the lake became brackish. The palaeosol dated 9950 ± 170 BP records the establishment of a periodically-flooded lake-side grassland (probably *Sporobolus spicatus*) with alkaline pools. Meanwhile, the exposed lake deposits in the Deka Wede embayment were slightly gullied by surface runoff.

i. Ziway-Shala V

A renewed transgression started in earnest after 9950 ± 170 BP. By 9810 ± 170 BP or slightly later, depending on the interpretation of the relevant ^{14}C dates, Abiyata had freshened considerably and was supporting a planktonic diatom flora (Descourtieux, 1977). This suggests that overflow into Shala had already begun. By the time the shoreline reached 1633 m, a rich freshwater mollusc and fish fauna had become established in the Shala Basin. The oldest basal shell date from unit UM4 in the Ajewa formation is 10,010 ± 150 BP (SUA-502). This is compatible with the measured age of 10,220 ± 140 BP from the underlying unit UM3, but appears slightly too old by comparison with the group of organic matter dates from the Abiyata Core (Fig. 6.8), which are probably more reliable in view of the clear evidence for reworking of the shells in unit UM4 (chapters 6 and 7).

If we then disregard SUA-502, the remainder of the radiocarbon dates from basal shell beds yield consistent estimates of 0.11 to 0.13 m/y for the average rate of lake-level rise between 1580 and 1659 m.a.s.l. This is, as might be expected, much slower than the rates experienced over short periods today (up to 1.9 m/y), but still indicates an annual water surplus of the order of 100-300 mcm/y (from Fig. 8.2).

The lake probably first overflowed around 9400-9300 BP. The lip of the outlet may initially have stood at 1673 m or even higher (chapter 5BIa), but was at one stage downcut to a minimum of 1660 m. It is by no means clear when back-filling of the channel floor to 1667 m occurred, although we can infer that it took place during a recession prior to the last high level, which is dated 5180 $\pm$ 55 BP (Q-1360). Beadle (1974, p. 231) describes how blockage of the outlet from Lake Malawi occurred during a lake-level minimum between 1915 and 1935 AD, due to formation of sand bars across the channel mouth, followed by the accumulation of alluvial silts which were consolidated by vegetation. Outflow finally resumed in 1935, at a level 6 m higher than in 1915. Since then Lake Malawi has remained comparatively high.

The SMU dates prove conclusively that the Ziway-Shala V lake stood for a significant period at the level of terrace IV (Haynes and Haas, 1974; Laury and Albritton, 1975). For lack of more precise information, it will be assumed from now on that outflow took place at the very prominent 1670 m shoreline level. But minor oscillations did occur, notably the Bulbula IIIB/IIIC regression which brought lake level temporarily back to 1617 m or below (Fig. 6.3) (Descourtieux, 1977). The duration of the entire Ziway-Shala V episode was quite short, probably only about 1500 years at the Deka Wede type locality (Table 7.3).

Common characteristics of the 1670 m palaeolakes

The Ziway-Shala V and VI palaeolakes were very similar in their geometry and general characteristics. The water-surface area at overflow (1670 m) was about 2690 km^2 (Table 8.2). It is unlikely to have been less than 2470 km^2 (1660 m) or more than about 2770 km^2 (1673 m) (Fig. 8.2). The shoreline configuration has been described in detail in chapter 5BI. The united lake consisted of a northern (970 km^2, maximum depth ca. 44 m) and a southern basin (1720 km^2, maximum depth ca. 378 m), whose water masses exchanged over the shallow ($<$ 20 m) Hondola sill (Fig. 5.1). It was therefore intermediate in depth and area between the present Lakes Idi Amin Dada (Edward) and Kivu (Balek, 1977), and nearly equalled Lake Turkana in volume (Yuretich, 1976). Estimates of the fetch and the depth of the wave-mixed layer in each basin are given in Table 8.2.

Table 8.2: Physical characteristics of the 1670 and 1600 m lakes

	1670 m lakes	1595-1600 m lake
Maximum depth	ca. 378 m	ca. 303 m
Increase in depth (Shala Basin)	112 m	37 m
Mean depth	76 m	41 m
Area	2690 km^2	1590 km^2
Ratio total lake area: present lake area	2.26 : 1	1.34 : 1
*Volume	204 km^3	68 km^3
Increase in volume relative to present	160 km^3	23 km^3
Ratio total lake volume: present lake volume	4.57 : 1	1.52 : 1
**Estimated turnover time (volume/annual water loss)	$<$ 38 y	21 y
Effective maximum fetch:		
North Basin	45 km	?
South Basin	30 km	27 km
***Estimated depth of wave-mixed layer:		
North Basin - force 3	9 m	?
- force 8	$>$ 30 m	?
South Basin - force 3	7.5 m	7 m
- force 8	26.5 m	25 m

* From the formula in Hutchinson (1975, p.166).

** Assuming an annual evaporation loss of 2 m (chapter 9).

*** From Smith and Sinclair (1972).

The increase in lake volume at 1670 m, 160 km^3, is equivalent to a 4.57-fold dilution compared with today. If the weighted mean salinity of the present lakes is taken as 13.7‰ (from the data in Table 2.3), this allows us to calculate a maximum limit of 3‰ for the Ziway-Shala V and VI lakes, neglecting the effects of the outflow and the input of salts by the tributary rivers. If it is further assumed that mixing with the Shala deep water was negligible, then this upper limit is reducedto 1.6‰. The diatom and mollusc assemblages indicate that dilution was considerably greater than these figures imply.

The relation of the outlet to the Meki Valley suggests that the 1670 m lakes were fed chiefly by the Katar, Gidu and Langano drainages, while the bulk of the Meki flow was diverted straight through the spillway, at least during the rains (chapter 5BIa). If confirmed, this possibility would further reduce the salinity of the lake, since modern runoff from the eastern escarpment is considerably more dilute than the Meki River (Table 2.5).

Because the major rivers probably accounted for the bulk of the inflow, as they do today, it is likely that there was a wind-assisted net circulation of the lake waters southwards and westwards from the Ziway and Langano Basins into the Shala Basin, where the (evaporation-precipitation) deficit is greatest (McKerchar and Douglas, 1974; Kingham, 1975; Makin et al., 1976). We can therefore imagine a slight increase in salinity and alkalinity towards the south end of the lake, similar to the situation obtaining today in Lake Turkana (Yuretich, 1976).

The sediment carried by the Meki was deposited as a thick alluvial fill in the present valley above Meki Town (chapter 6E). It is probable that the Katar also dumped most of its load in the flooded Shetemata graben. Significant quantities of fine-grained pumiceous sediment were supplied to the lake by local runoff, and accumulated together with diatom frustules on low-angle underwater fans. A large proportion of the beach and shallow-water deposits in the basin were probably derived from shoreline erosion.

Characteristics of the Ziway-Shala V lake

During this episode the deep-water sediments laid down at the Abiyata Core site were diatomaceous clays with up to 19% $CaCO_3$. In the Shala Basin, only the shell concentrations at the upstream end of unit UM4 can be attributed with confidence to this phase. Marl deposition once again predominated locally in the Deka Wede embayment.

Between the levels dated 9810 $\pm$ 170 BP and 7970 $\pm$ 150 BP the diatom assemblages in the Abiyata Core are largely planktonic (Descourtieux, 1977). The flora indicates a large, oligohaline lake with medium to medium-low alkalinity (< 30 meq/l) (Fig. 6.8). But the fluctuating presence of Fragilaria brevistriata betrays a certain ecological instability compared with the much deeper and more extensive early Holocene Lake Abhé (Gasse and Descourtieux, in press). This species is sometimes associated with colonies of the blue-green alga Microcystis in the plankton (Haworth, 1977).

Stephanodiscus astraea and its varieties reach a maximum abundance of 54%. This species is common today in deep lakes such as Victoria (79 m), Mobutu Sese Seko (Albert) (58 m), Idi Amin Dada (112 m), Kivu (500 m) and Tanganyika (1500 m) (Richardson and Richardson, 1972; Degens et al., 1973; Gasse, 1975). It attains its greatest abundance in dilute lakes (0.050 to 0.5‰) containing less than 10 mg/l dissolved silica in their surface waters (Richardson, 1968, 1969a; Kilham, 1971b; Gasse, 1977b). During the early Holocene, S. astraea proliferated in an astonishing variety of expanded lakes, including Abhé, Dobi-Hanlé and Afrera in the Afar (Gasse, 1975; Gasse and Street, 1978a), Turkana (B. Owen, pers. comm., 1977), Naivasha (Richardson and Richardson, 1972), Mobutu Sese Seko (Harvey, 1976), Victoria (Kendall, 1969) and Manyara (Holdship, 1976). In many cases it was also associated with the precipitation of calcite from the surface waters (Gasse, 1977b).

The pattern of carbonate sedimentation during phase Ziway-Shala V clearly has implications for the hydrology and water chemistry of the united lake, although the problems of interpretation are by no means resolved (chapter 7DIII) (Gasse and Street, 1978a,b). During the Late Pleistocene maxima IIIA and IIIB, carbonate sedimentation was apparently restricted to the littoral and inner offshore zones, and to the Deka Wede embayment, where it can be attributed to localized hot-spring activity. In contrast, during phase V, low-Mg calcite accumulated in abundance during the initial brackish-water stage, but precipitation continued into the high-water, Stephanodiscus phase. The underlying causative factors identified in chapter 7 appear to be higher water temperatures; an increased annual water flux through the lake; higher phytoplankton productivity, associated with the dominance of S. astraea and its varieties; and an enhanced calcium supply. This hypothesis conflicts slightly with the model put forward by Hecky and Degens (1973, p. 33), who state that calcareous offshore deposits can only form in lakes which are sufficiently dilute for Ca^{++} to be mixed into open water rather than being immediately precipitated around the periphery, as occurs under brackish conditions. Their argument implies that Lake Ziway-Shala V was less alkaline than Lakes Ziway-Shala IIIA and IIIB, which runs counter to the diatom evidence (Fig. 6.8). However Hecky and Degens were dealing with Lake Kivu, which has always been more alkaline than Lake Ziway-Shala, to judge from its diatom record. It is possible to reconcile all the available evidence by assuming that the waters of Lakes Ziway-Shala IIIA and IIIB lay in the range presently occupied by a third class of lakes such as Naivasha which are too dilute for significant carbonate precipitation to occur except around their margins, where photosynthetic control of pH is more effective (Beadle, 1932; Ruttner, 1963; Richardson and Richardson, 1972; Mothersill, 1976).

The faunal assemblages confirm and supplement the environmental picture deduced from the diatoms. The Ziway-Shala V lake supported a large biomass of pulmonate, prosobranch and bivalve molluscs. These evidently lived in lush fringing vegetation, or as in Lake Chad, in dense offshore banks where they fed on algae and organic detritus. Although three species of palaearctic pisidia are found in section 0738-B2, the lower Holocene assemblages are generally much more tropical in character and vigour than the Late Pleistocene ones. Freshwater species such as Bellamya and Caelatura accounted for a large proportion of the total biomass. In addition we find freshwater ostracods and abundant rolled fish bones, although few species have been successfully identified.

The present distributions of the mollusca, particularly Bellamya, allow us to define the ecological characteristics of the Ziway-Shala V lake more precisely. Bellamya is today restricted to lakes of low to medium-low alkalinity (<1 to 10 meq/l) (chapter 7DIIId). In Lake Chad it reaches its greatest abundance at the lower end of this range, but this may not apply to eastern Africa (Beadle, 1974, p. 171). Two possible analogues for the early Holocene united lake are therefore Lakes Mobutu Sese Seko (= Albert) (7-9 meq/l) and Abaya (8.5 - 10 meq/l), which both have outlets, and which support similar diatom floras and closely related, though more diverse, mollusc and fish faunas (Worthington, 1929; Brunelli and Cannicci, 1941; Parenzan, 1941; Riedel, 1962; Brown, 1965; Talling and Talling, 1965; Beadle, 1974, ch. 10; Harvey, 1976). However, it is important to bear in mind that Ziway-Shala V was much deeper than either and may have been more dilute. We can also deduce by comparison with Lake Idi Amin Dada (112 m deep) (Beadle, 1974, p.81) which is probably the nearest analogue from the point of view of circulation, that the Shala Basin was permanently or near-permanently stratified during the Ziway-Shala V episode, although the sedimentary evidence from the Ajewa area is not sufficient to test this hypothesis.

j. The Ziway-Shala V/VI regression

Lake-level began to fall sharply again before 8520 ± 200 BP, when a thin palaeosol formed on air-fall pumice in the Deka Wede embayment. There is a slight inconsistency, though probably not statistically significant, between this date and the measured age of 8320 ± 175 BP derived from shells in deltaic sediments at 1646 m in the Shala Basin (Q-1114; Fig. 8.1). Deep-water conditions came to an end shortly after 7970 BP in the Abiyata Basin.

The V/VI regression is most clearly seen in section 0738-B2, where the early Holocene lake beds are overlain by 3.7 m of colluvium. In the Abiyata Core, it is represented by gritty pumice sand with abundant phytoliths and very few diatoms. There is no indication that the core site dried out. The lake became brackish and alkaline (>50 meq/l) confirming that overflow into Shala had ceased. The diatom floras suggest a shallow eutrophic lake with dense blooms of algae, rather like Abiyata today (Descourtieux, 1977). Meanwhile the level of Lake Shala dropped to 1561 m or below.

k. Ziway-Shala VI

By 6500 BP, lake-level had risen again to 1651 m. Using the mean rate obtained for the Ziway-Shala V transgression, we can estimate that the water level began to rise around 7100 BP. The altitudinal distribution of dated shell beds is rather chaotic (Fig. 8.1) but shows that overflow through the Meki-Dubeta col at 1668.5 m or above took place around 5200-4900 BP. It is possible that there was a minor recession around 5700 BP which gave rise to the Roricha fish bed; a hypothesis which fits in well with the evidence for fluctuating conditions in the Deka Wede type section, although the magnitude of the drop in level is uncertain.

Sediments which can be firmly attributed to phase VI include: the partly dissected Deka Wede marls; units D3 and UM4 in the Ajewa formation; and the thin and suspiciously sandy sediments at the very top of the Abiyata core. In all probability most of the fresh depositional features along the 1670 m shoreline also date from this episode. In the central Bulbula Plain phase VI is represented mainly by thin and very shelly lag gravels. The reasons for the lack of fine sediments in this area are discussed further in chapter 7DIIId.

The shoreline configuration and bathymetry of the lake have already been documented (chapter 5BI and section j above). Its diatom flora is only known from the upper 55 cm of the Abiyata Core, and is similar to Ziway-Shala IIIA and IIIB, consisting largely of Cyclotella ocellata and Melosira granulata var. valida. This association seems to indicate an oligohaline lake of medium-low alkalinity (<10 meq/l), with more than 10 mg/l dissolved silica in its surface waters (Kilham, 1971b; Richardson et al., 1978). The companion species Fragilaria brevistrata and Cyclotella kützingiana reflect a clear tendency towards oligotrophism. Nordic and alpine species are absent (Descourtieux, 1977; Gasse and Descourtieux, in press). However this data should be interpreted with caution, because it is not certain that the lake-level maximum is recorded and in any case the deposits may have been reworked.

Fortunately the fauna is better known. This is the richest of any of the palaeolakes. The list includes 18 species of mollusca, 5 species of freshwater ostracods, a freshwater crab, and at least 6 species of fish. The relatively diverse and abundant mollusc fauna implies the existence of fringing vegetation and a rich organic substrate on the gently shelving Bulbula and West Ziway terraces, even though the organic matter has long since been oxidized. The large freshwater benthic taxa Bellamya and Caelatura are particularly common. Pockets of swamp in the Shala caldera provided calm conditions in which the "giant" swamp-dwelling form of Lymnaea natalensis (chapter 6CIIc) and large numbers of the pulmonates Bulinus "truncatus" and Biomphalaria pfeifferi could flourish. The presence of B. sudanica strongly suggests that Aeschynomene thickets and lagoons had become established along the shores of the Ajewa-Aneno embayment. Water temperatures were probably close to present or only slightly higher, judging from the abundance of B. pfeifferi.

The Roricha fish bed provides strong circumstantial evidence for the establishment and sudden breakdown of density stratification in the Shala Basin at the beginning of phase VI. Alternative explanations for the origin of this bed are discussed in chapter 6DII.

l. The Ziway-Shala VI/VII regression

Shortly after 4800 BP lake-level fell dramatically. The Ziway-Shala VI/VII unconformity extends down to 1569 m or below in the Ajewa area. It is likely that this level was reached before 4000 BP. The very regular pattern of strandlines in the southern Bulbula Plain leaves little doubt that recession was rapid and monotonic, although the crest spacing is probably too great for the ridges to be annual in origin. During the subsequent inter-lacustral interval the exposed sediments in the Ajewa were gullied and reworked by subaerial processes. Acacia charcoal from the resulting colluvial-alluvial deposits is dated 2510 $\pm$ 100 BP (unit DM5).

m. Ziway-Shala VII

Shortly after 2500 BP Shala rose briefly to about 1595-1600 m, uniting with Abiyata and Langano to form a lake about 1590 km^2 in area. No dates are available from this episode. The deposits attributed to it include spits, barrier beaches and shore dunes, small deltas and ostracod-rich shallow-water silts and sands. Fishbone clusters are very common in the sediments, but only Tilapia has been identified with certainty. Mollusca are very scarce and show signs of reworking.

Although we have no diatom analyses to rely on, we can calculate a maximum salinity for the Ziway-Shala VII lake based on volumetric considerations. A water volume of 67.6 km^3 in the South Basin represents a 1.53-fold dilution compared with today. This suggests an upper salinity limit of the order of 9.3‰ compared with the present weighted mean of 14.2‰ for the three southern lakes. If we assume that little mixing with the Shala deep water took place then this estimate is reduced by a further 3‰. This is in good agreement with the faunal evidence, which suggests a mesohaline lake with an alkalinity of 25 meq/l or more. Tilapia nilotica is a salt-tolerant, herbivorous fish, able to feed directly on phytoplankton (Beadle, 1974; Greenwood, 1976), and could therefore live quite happily on the algal blooms which characterize Abiyata and other shallow mesohaline lakes (Beadle, 1974).

n. The Ziway-Shala VII/VIII regression

The three southern lakes shrank to about their present extent at the end of the Ziway-Shala VII episode, leaving a magnificent series of recessional ridges which are now covered by dense forest along the southeast shores of Shala and Langano.

o. The Ziway-Shala VIII transgression

This small oscillation in lake-level affected Shala and possibly Abiyata, but did not result in a connection between the two. It has left widespread traces around Lake Shala in the form of drowned trees, bouldery strandlines, small deltas and abandoned spring heads up to 5.8 m S.D. The date of this rise is not known but may have been late nineteenth century. It is unlikely that there was any significant change in the water chemistry or ecology of the lake.

8C THE CATCHMENT

I. Introduction

In order to reconstruct the water budget of any former lake in the Ziway-Shala Basin, it is first necessary to estimate several important parameters (chapter 9). These include the mean annual air temperature, which is required in order to calculate the open-water evaporation rate; and the runoff coefficient for the catchment. Palaeotemperature estimates for the last glacial maximum in East Africa have been made from the snowline lowering on several glaciated massifs, using the present-day lapse rate (Osmaston, 1965, 1975) (Table 8.3). In this section the former snowline on Mt. Badda is reconstructed from the elevations of the terminal moraines, and used to determine the glacial-maximum temperature lowering. The pollen, diatom and soil-stratigraphic evidence from the catchment is then reviewed as an indication of possible variations in runoff through time.

II. Late Quaternary glaciation

a. The extent of former ice caps and cirque glaciers in Ethiopia

Although none of the Ethiopian mountains are glacierized today, traces of Pleistocene glaciation exist on a number of high massifs. These provide a check on the reliability of the snowline reconstruction for Mt. Badda. Nilsson (1935, 1940) was the first to describe glaciated valleys and moraines in the High Simen, in the Arussi Mountains, and in the mountains of Wollo Province (Guna and Amba Farit) (Fig. 8.3). The last two occurrences have never been substantiated by later workers. In the Simen he found evidence for two glaciations, which he assigned to the "Great Pluvial" and "Last Pluvial" respectively. The deposits attributed to the "Great Pluvial" have now been reinterpreted as periglacial slope mantles (Williams, Street and Dakin, 1978).

Nilsson's "younger glaciation" is recorded by small NW-facing cirques and well-preserved moraine ridges on Ras Dejen (4543 m), Bwahit (4430 m) and Silki-Abba Yared (4420 m).[1] Estimates of the former firnline range from

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1. Altitudes from the 1:50,000 map by Werdecker (1967), corrected to conform with the new USCGS datum (Messerli *et al.*, 1975).

Table 8.3: Summary of snowline data from Ethiopia and East Africa

Massif	Latitude	Summit (m.a.s.l.)	Estimated glacial-maximum firnline (m.a.s.l.)	Greatest development of glaciers	Wettest side today	Present precipitation maximum (mm)	Average lowering of firnline at glacial maximum (m)	Glacial maximum temperature depression (°C)	Modern firnline (m.a.s.l.)	Glaciated area (km^2)
Ethiopia										
Simen	13°N	4543	4200-4300	NW, N	SW?	1600	(650)	(4-5)[3]	(4900)[1]	60
Badda	8°N	4210	3750-3900 (av. 3850)	E, NE	E?	1600	(900)	(6)[4]	(4750)[2]	ca.140
Balé	7°N	4450	3700	N, NE	SW?	1500	(1050)	(7)[4]	(4750)[2]	?
East Africa										
Kenya	0°	5199	3660-4270	N, W	SE	2290	760	5[3]	4630-4790	ca.200
Kilimanjaro	3°S	5996	3660-5160	SE-SW	SE-SW	2030	880	6[3]	5360-5730	150
Ruwenzori	0°	5109	3510-4270	E, SE	E, SE	2670	640	4[3]	4500-4720	260

Sources:

Simen: Nilsson (1940); Schaller and Kuls (1972); Hastenrath (1974).

Badda: This study (Table 8.4).

Balé: Schaller and Kuls (1972); Messerli et al. (1977).

Kenya: Osmaston (1965); Baker (1967).

Kilimanjaro: Osmaston (1965).

Ruwenzori: Osmaston (1965).

Notes:

1. From mean annual 0°C isotherm in the free atmosphere (Hastenrath, 1974).
2. From mean annual 0°C isotherm in central Ethiopia (estimated from the surface temperature/altitude regression on p. 25. This corresponds closely to the firnline on the wetter mountains (Osmaston, 1965; Hastenrath, 1973), and yields a conservative estimate of snowline lowering.
3. Using average lapse rate for eastern Africa (6.5°C/1000 m) (Osmaston, 1965).
4. Using lapse rate of 6.75°C/1000 m calculated for central Ethiopia (p. 25).

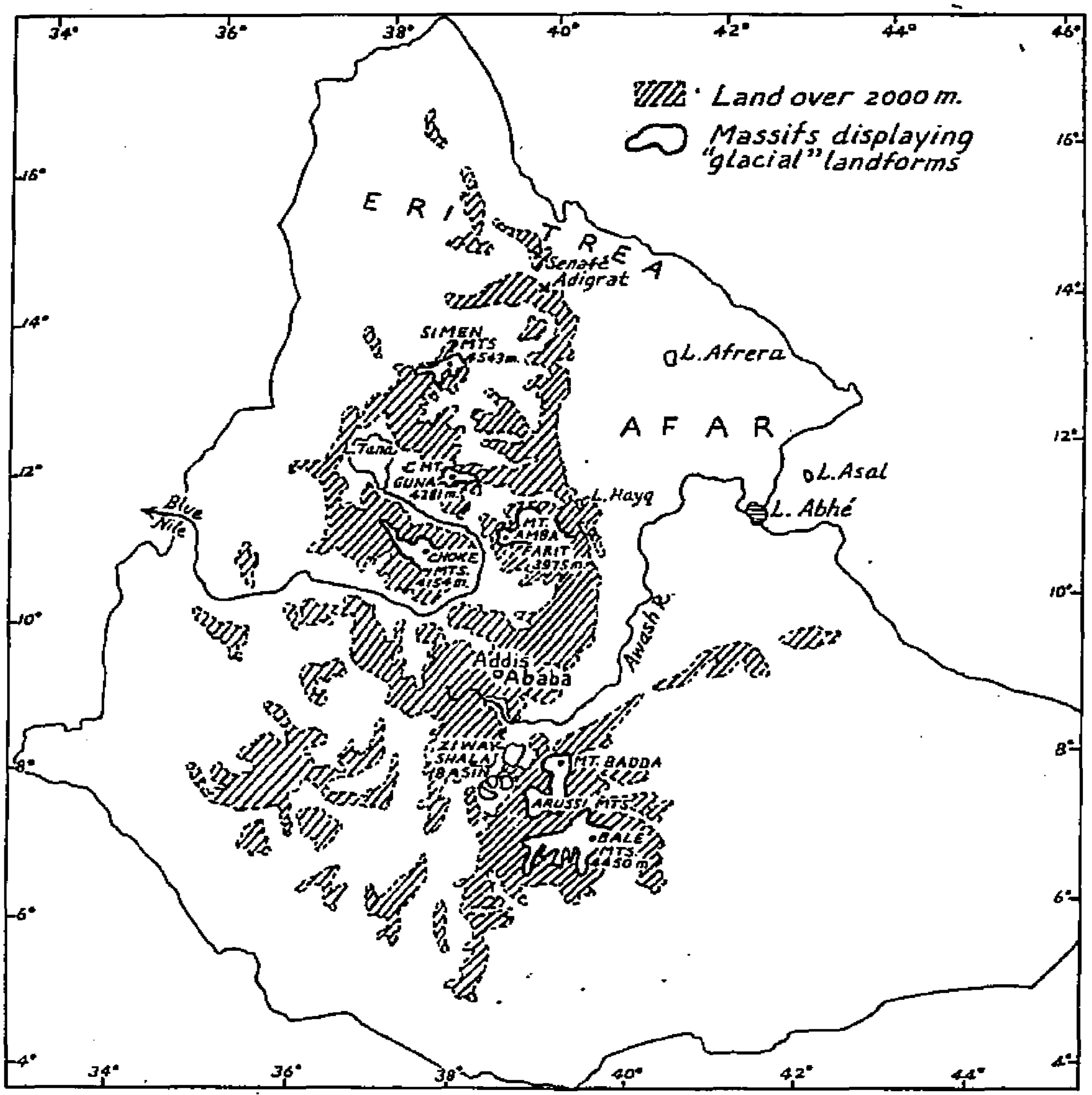


Fig. 8.3: Location map of Ethiopian massifs displaying "glacial" landforms.

4100 to 4400 m (Nilsson, 1940; Minucci, 1938; Büdel, 1954; Hövermann, 1954a; Werdecker, 1955, 1958; Hastenrath, 1974) (Table 8.3). Hastenrath has correlated the maximum advance with glacial stades IA – ID on Mt. Kenya (Baker, 1967) and with the Fourth (Main) Glaciation of Kilimanjaro (Humphries, 1972), thereby implying a Late Pleistocene age (Table 8.3). Although there are many records of snow falling on the massif this century, the present snowline does not intersect the highest peaks. Its elevation was estimated by Büdel (1954) and Werdecker (1955) to lie around 4700-4800 m. Hastenrath (1974) notes that the mean annual 0°C isotherm in the free atmosphere is reached at approximately 4900 m in the vicinity of the Simen.

The area covered by the "younger glaciation" of the Simen is very small compared with the extent of the ice caps which formed on the Balé and Arussi Mountains in southern Ethiopia. The glaciation of Balé, which is the fifth highest massif in Africa, was probably first discovered by Mooney (1963). Glacial modification is most pronounced on the north and northeast sides of the massif, especially in the Danka, Shiya and Tagona Valleys. Messerli et al. (1977) have calculated a former snowline elevation of 3700 m. They envisage an ice cap completely covering the Balé Plateau, feeding outlet glaciers 10-12 km in length which extended down to 3100-3200 m.

Other workers have described highly equivocal evidence for glaciation in the highlands of Eritrea and Gojjam (Mt. Choké) (Hövermann, 1954b; Kuls and Semmel, 1965). The features reported include small bouldery ridges and cirque-like valley heads. The outer limit of the "terminal moraines" described by Hovermann is so low (about 2000 m) in comparison with East Africa and the Mediterranean (Osmaston, 1965; Messerli, 1967) that his conclusions should be viewed with considerable scepticism (Kuls and Semmel, 1965). It is probable that many of the so-called "moraines" are landslide-debris tongues or other nonglacial features.

b. The glaciation of the Arussi Mts.

i) Previous investigations

Nilsson (1940) first suggested that at least two of the summit volcanoes of the Southeastern Plateau, Mts. Kaka (= Badda) and Chilalo, had been glaciated. The glacial limits have subsequently been remapped from aerial photographs by Potter (1976), Mohr and Potter (1976) and Hastenrath (1977). According to Potter (1976) the lowest terminal moraines extend to an altitude 350 m below the headwall in valley 2. He estimated that the total ice-covered area was 140 km^2. The latest map by Hastenrath (1977) adds very little, and it may be that he was denied full access to the aerial photography.

Small ice caps and cirque glaciers have also been mapped on Mts. Chilalo (4005 m), Kakka (4245 m) and Enkwolo (3850m). The glaciated areas shown by Mohr and Potter (1976) can be estimated to cover 19 km^2, 18 km^2 and 2 km^2 respectively (Gasse, Rognon and Street, in press).

ii) <u>This study</u>

The present study concentrated on Mt. Badda, which is not only the most accessible of the four massifs but also the only one which shows unequivocal moraines. In my view, most of the features mapped by Hastenrath on Chilalo and Kakka are lava flows and rock ridges. Field visits were made in January 1975 and again in March. The primary object was to obtain cores from moraine-dammed lakes or bogs: owing to vehicle problems and the early onset of the small rains, we were unable to investigate the lateral and terminal moraines in the field. However David Lauder and I succeeded in obtaining a 3 m core from a bog at 4000 m (Bog 3, Fig. 8.4). The moraines have been mapped using the 1967 USAF and 1972-73 D.O.S. aerial photos (Fig. 8.4). The elevations given are interpolated from the 20 m contours on the D.O.S. 1:50,000 sheet 0739A2, which is still under restricted access.

Mt. Badda, which is shown on the D.O.S. maps as the Galama Range, is a deeply dissected, basaltic-trachytic central volcano forming the northern end of the Sagatu Ridge; an Upper Pliocene-Lower Pleistocene fissure system which runs NNE across the Southeastern Plateau. Unfortunately the volcanism is too ancient (3.6 - 2.1 m.y.) to be used for dating the glaciations as has been done on Kilimanjaro (Humphries, 1972) and Mauna Kea, Hawaii (Porter <u>et al.</u>, 1977). The undulating ridge crest rises from 3850 m on the Digelu-Ticho road to 4190 m southeast of the main summit. Deep, glacial troughs

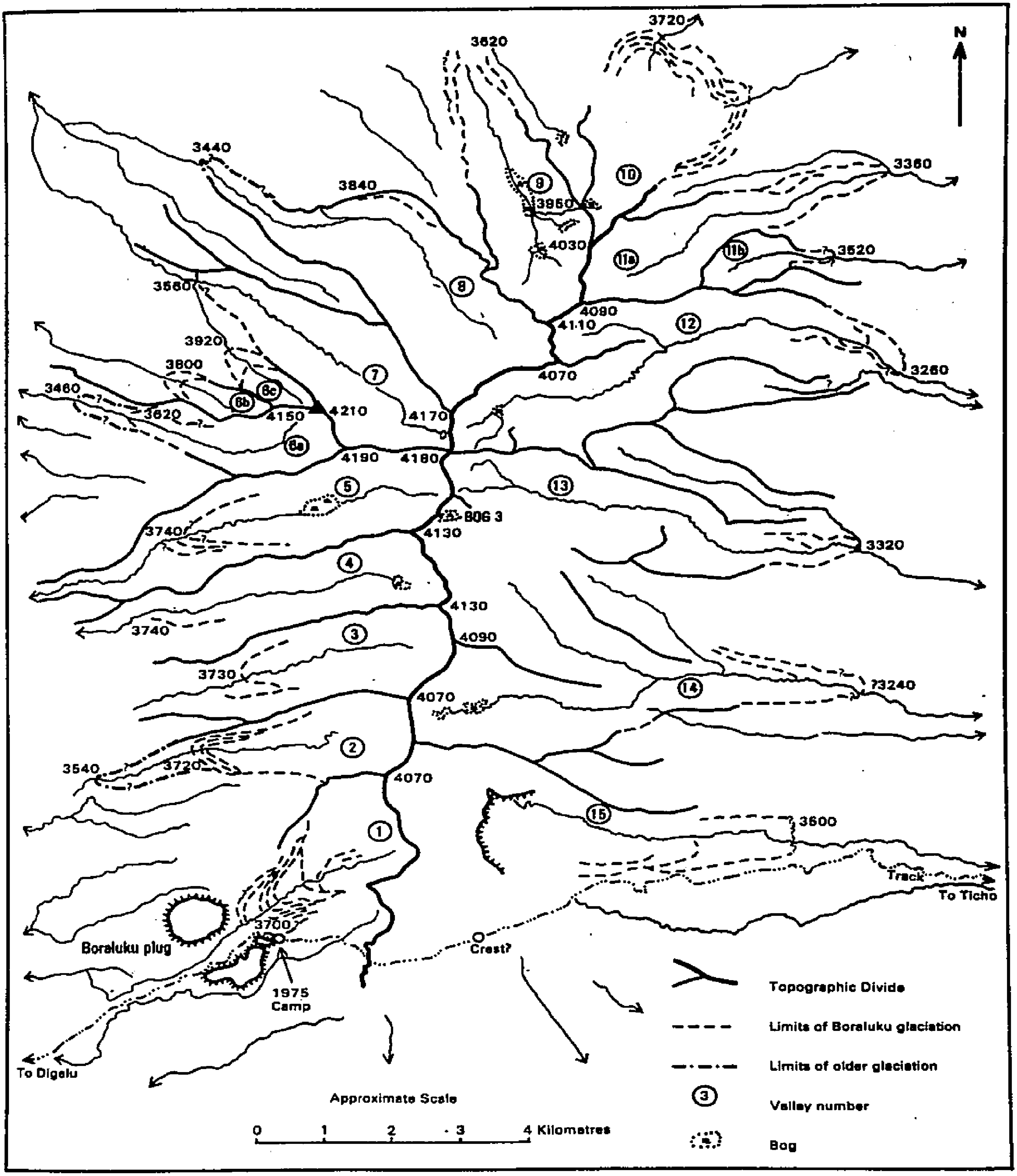


Fig. 8.4: Former extent of glaciers on Mt. Badda, Arussi Mts. (altitudes in m.a.s.l.).

radiate out from this divide. These show less glacial modification than the Tagona Valley or Mt. Kenya, but more than the Simen. Erosional features include cirque heads, rock steps, and a small tarn. The high peaks and cols have a rounded and ice-scoured appearance reminiscent of the Scottish Highlands (Plate 79), which suggests that the valleys were outlets for a small ice cap straddling the summit ridge. The pattern of striations supports this hypothesis, since they are aligned parallel to the axis of valley 2 but also trend across the divide between valleys 3 and 4 (Potter, 1976).

As noted by Nilsson (1940) and Potter (1976), there is evidence for two glacial advances on the western side of the range. The valleys change from a V-shaped to a U-shaped cross profile at about 3440-3540 m (Table 8.4), although any associated moraines have been largely removed by erosion. Further upstream there is usually a series of moderately sharp-crested, arcuate ridges which strongly suggest terminal moraines. Potter (1976) found that they consist of unconsolidated, subangular to subrounded cobbles and boulders up to 3 m in diameter. Lateral moraines are preserved in some valleys where the side walls are not too steep.

On the eastern flank the style of glaciation changes. The valleys are larger and have more tributary heads. Nested terminal moraines like those on the western flank do not occur; instead the deep troughs are bounded in their lower reaches by continuous, smoothly curving, sharp-crested, raised rims. These are particularly well developed in valley 11a. Although it is impossible to be sure without visiting them in the field, these ridges are interpreted as lateral moraines, for the following reasons:-

1) They are too continuous, smoothly-sloping and sharp-crested to be lava flows or non-glacial drainage divides.
2) They frequently display downslope vegetation stripes, which suggests that they consist of unconsolidated material.
3) They only border valleys which have large cirque-like valley heads and rock steps.
4) They resemble the Late Pleistocene (Stage I) moraines in the Gorges Valley, Mt. Kenya (Baker, 1967).

Table 8.4: Glacial limits on Mt. Badda, Arussi Mountains

Valley	Highest peak (m.a.s.l.)	Change from U-shaped to V-shaped cross profile (m.a.s.l.)	Lowest well-preserved terminal moraine of Boraluku stage (m.a.s.l.)	Mid-height Boraluku stage (m.a.s.l.)	Maximum glacier length, Boraluku glaciation (km)
Western side of divide					
1	4070		3700	3885	3.5
2	4070	3540	3720	3895	3.3
3	4130		3730	3930	3.6
4	4130		3740	3935	4.5
5	4190		3740	3965	4.5
6a	4210	3460	3620	3915	3.4
6b	4150		(3800)	3975	1.7
6c	4210		(3920)	4065	1.7
7	4210		3560	3885	5.0
8	4170	3440	< 3840	< 4005	>4.2
9	4110		3620	3865	5.1
10	4030		3720	3875	4.0
			Average	3906	
Eastern side of divide					
11a	4090		3360	3725	5.3
11b	?		3520	–	c. 2.1
12	4190		3260	3725	7.8
13	4190		3320	3755	6.6
14	4130		3240	3685	7.8
15	4070		3500	3785	4.7
			Average	3735	

The inferred lower limit of these glaciers on the eastern flank is 3240 to 3520 m (Table 8.4).

We are now faced with the problem of correlating the glacial evidence on the two sides of the massif. Either the eastern moraines are equivalent to the younger moraine complex on the western flank, or else they represent the older glaciation, in which case there is no sign of any features attributable to the younger glaciation on the eastern side. Hastenrath (1977) apparently overlooked the evidence for more intense glaciation of the eastern valleys, but it is probable that he only had access to the D.O.S. photos on which the ridge crest and parts of the eastern slopes are obscured by cloud.

A comparison of the morphology of the eastern and western-style moraines in adjacent valleys 10 and 11a strongly suggests that they are contemporary. This ice advance will be called the Boraluku glaciation, after the large volcanic plug which overlooks valley 1, where fresh, arcuate moraines are particularly well developed. The lower, glacially modified valleys and eroded moraines on the western flank are attributed to an older (unnamed) glaciation.

The Boraluku Glaciation

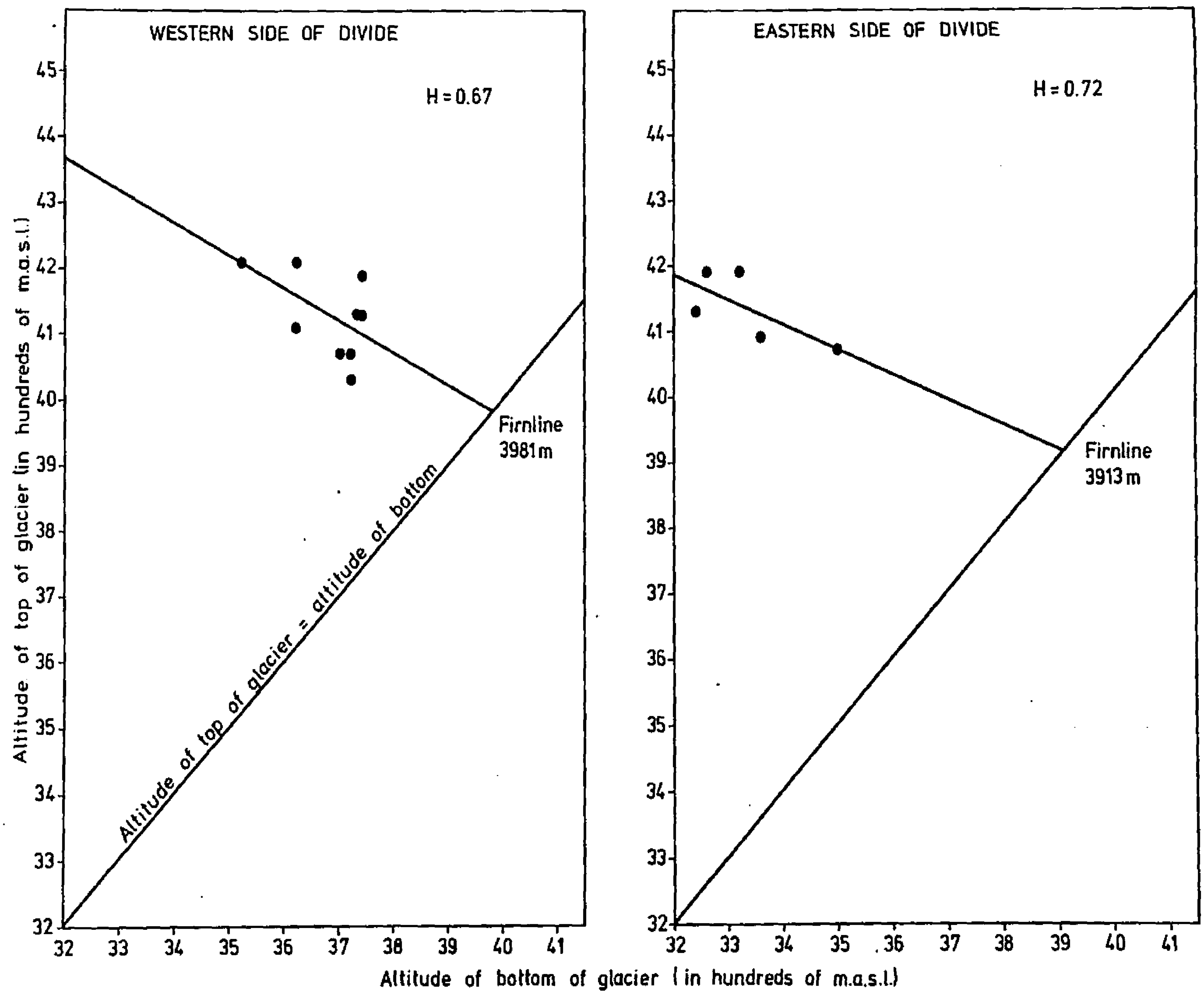
- Extent

The glaciers of the Boraluku glaciation ranged from about 1.7 to 7.8 km in length (from Fig. 8.4). Several of the western tongues, notably 1, 2 and 10, occupied wide, shallow troughs in which the most intense erosion was concentrated near the headwall. Osmaston (1965) believes that this is a characteristic of tropical glaciers. On the eastern flank the glaciers were longer and apparently more active.

- Firnline elevation

In the absence of area or accumulation data, the firnline of the Boraluku glaciation can be estimated using Hofer's mid-height method or the more precise general height method of Osmaston (1965, 1975) (Appendix 8). For this purpose the glaciers were divided into a western and an eastern group. The results of the general height method are shown in Figure 8.5. Both groups

Fig. 8.5: General height graphs for the glaciers of the Boraluku stage, Mt. Badda (for explanation see Appendix 8).



WESTERN SIDE OF DIVIDE
H = 0.67
Firnline 3981 m
Altitude of top of glacier = altitude of bottom
EASTERN SIDE OF DIVIDE
H = 0.72
Firnline 3913 m
Altitude of top of glacier (in hundreds of m.a.s.l.)
Altitude of bottom of glacier (in hundreds of m.a.s.l.)

yield a Höfer coefficient greater than 0.50, which is characteristic of plateau glaciers (Osmaston, 1975, p. 231); but due to the small number of points the relationships are not statistically significant ($0.10 < P < 0.20$). A mean firnline for each group has therefore been determined by the mid-height method, which gives a mean value 170 m lower on the eastern side. The average for the whole massif, 3845 m, falls mid-way between the estimates for the Simen and the Balé Mountains, and corresponds to a lowering in mean annual temperature of the order of $6^{\circ}C$ (Table 8.3), assuming no change in precipitation.

- Date of deglacierization

A 3 m-long core was collected from the centre of a small peat-filled basin (Bog 3) near the headwall of valley 14, at about 4000 m. The basal 10 cm of peat have yielded a ^{14}C date of 11,500 ± 200 BP (GIF-3891). This is a minimum limit for deglacierization, since the corer did not reach sand or gravel, but we can conclude that even the highest cirques were ice-free before 11,500 BP. The arcuate recessional moraines of the Boraluku stage may therefore be appreciably older. This view is consistent with the lack of any fresh unvegetated moraines attributable to neoglacial readvances (Osmaston, 1965; Baker, 1967; Humphries, 1972).

- Effect on the hydrology of the Ziway-Shala Basin

The total extent of ice within the Ziway-Shala catchment was quite small; about 60 km^2. On Mt. Badda only valleys 1-5 (21 km^2 of ice) drained in this direction, although most of the glaciated area mapped by Mohr and Potter (1976) on the other three massifs lies within the basin. Even during deglacierization, the input of meltwater into the Katar River was probably very much less than the latter's present interannual range of discharge (Table 2.1, Fig. 5.11), based on data from the rapidly shrinking Lewis Glacier on Mt. Kenya which show that net wastage is only about 0.37 m/y water equivalent (Hastenrath, 1975).

III. Vegetation history and bog development

a. The Mt. Badda Core

Bog 3 lies in a small cirque basin with a rock step (Plate 80). Although it is presently surrounded by Afroalpine Alchemilla moorland, the bog surface itself is covered by tussock-forming species (probably Cyperaceae) with patches of Sphagnum moss.

Madame Delibrias has kindly dated five samples of peat at Gif-sur-Yvette. Figure 8.6 reveals an apparent abrupt increase in sedimentation rate around 3400 BP, from 0.128 mm/y to 0.558 mm/y. Dr. Alan Hamilton and Dr. Françoise Gasse have undertaken the study of the pollen and diatoms in the core. Their findings are summarized in Figure 8.7.

b. Pollen analysis

The Badda diagram is divided into three pollen assemblage zones: B1 to B3 (Hamilton, 1977a,b). The ages of the B1/B2 and B2/B3 boundaries, as interpolated from Figure 8.6, are 6400 and 1850 BP. The vegetation shifts indicated by the pollen curves are summarized in Figure 8.7. The most important feature is the change from a very dry to a more mesic vegetation around the B1/B2 boundary. Dr. Hamilton feels strongly on the basis of the East African evidence that this change must have taken place at or before 9500 BP, and that the date of 7490 ± 140 BP from 232-250 cm is therefore suspect, perhaps as a result of contamination during a later stillstand of peat growth. His view derives some support from a new date of 7920 ± 80 BP (Q-1362) for the beginning of the forest period on the Balé Mountains (Hamilton 1977a) (Appendix 2b).

Zone B3 is generally similar to zone B2, but there are increased values of scrub pollen types, together with weeds such as Rumex and Plantago which are indicative of human interference. This zone therefore records the beginnings of agricultural clearance within the montane forest belt.

c. Diatom analysis

Whereas the pollen evidence reflects the broad pattern of vegetation change across the Southeastern Plateau, the diatoms in the Mt. Badda core

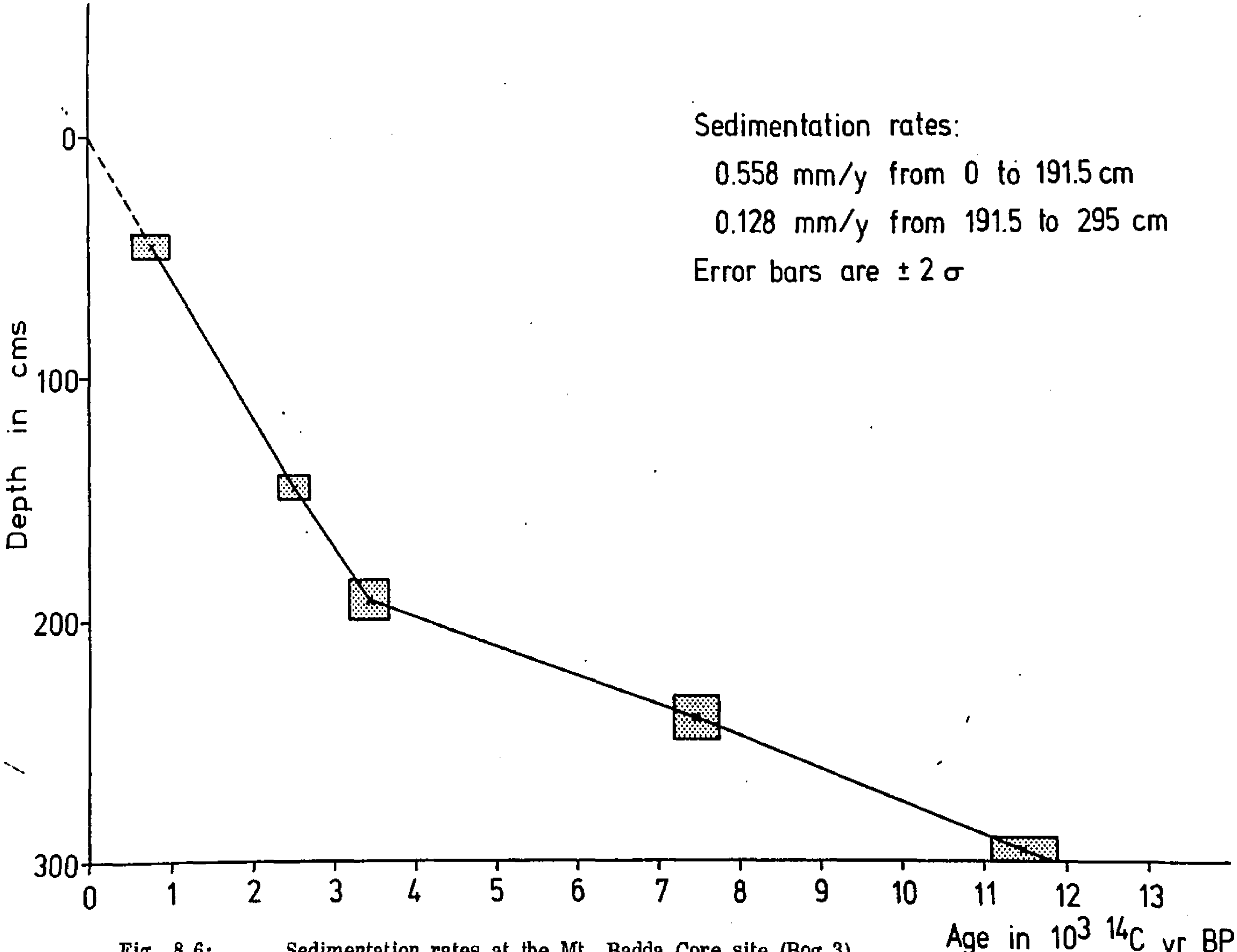
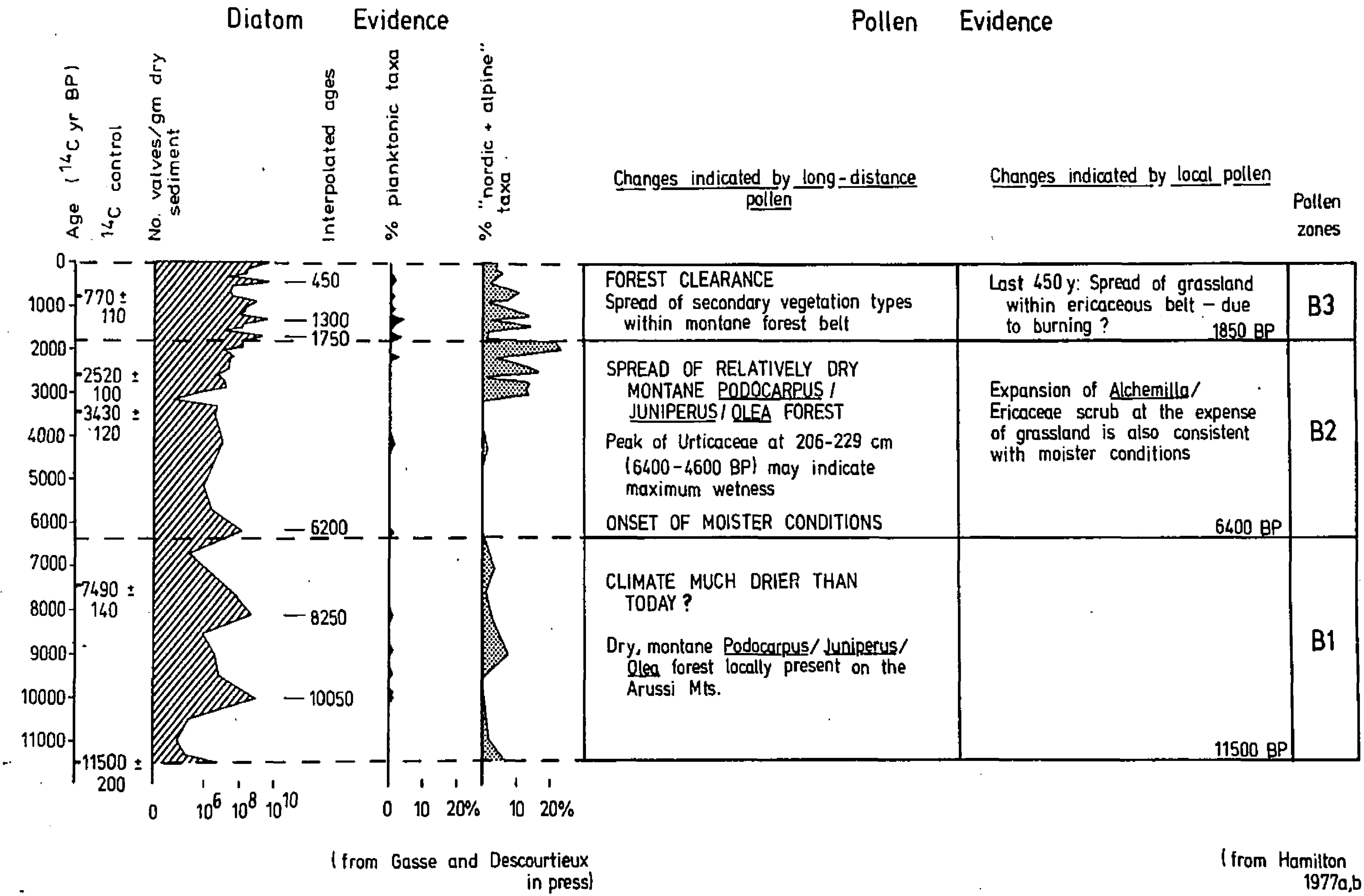


Fig. 8.6: Sedimentation rates at the Mt. Badda Core site (Bog 3).

Fig. 8.7: Summary of palaeoecological data from the Mt. Badda Core.

MT. BADDA CORE



record much more local, short-term events which resulted in an alternation of lacustrine and bog conditions at the core site itself (Gasse, 1978; Gasse and Descourtieux, in press). The findings are discussed briefly here because Dr. Gasse has suggested that the lacustrine episodes correspond to high lake levels in the Rift and are therefore indicative of increased runoff from the montane areas of the catchment.

The diatom content of the peat oscillates from 7.5×10^4 to 1.3×10^9 valves/g of dry sediment, with peaks at roughly 25 cm intervals (Fig. 8.6). The maxima are associated with an alkalophilous diatom assemblage, suggesting a truly aquatic habitat rich in diatoms and higher plants. The water was apparently oligohaline, alkaline (pH $>$ 7) and tending towards mesotrophism. We can picture a small tarn fed by runoff from the surrounding basaltic and trachytic volcanics. The minima correspond to an acidophilous assemblage with numerous subaerial species. This implies a damp, acidic (pH $<$ 6) environment with low trophic status. Oxidizing conditions led to partial dissolution of the diatom frustules. In this case, it is likely that we are seeing a trend towards an ombrogenous bog with Sphagnum growing between the tussocks, such as we find today at the core site (Plate 80). The second association also contains a significant "nordic" and "alpine" element which is particularly noticeable after 3150 BP (Fig. 8.7). However it would be a mistake to interpret the abrupt oscillations of these "cold" species as a function of temperature changes alone; according to Gasse and Descourtieux, they are more likely to reflect factors such as water chemistry and nutrient availability.

The cyclical variations in diatom content can be interpreted as the result of changes in local runoff (Model 1). According to this hypothesis, the "alkaline" peaks record the passage to wetter climatic conditions. The principal events can be dated about 10,050, 8250, 6200, 1750 and 1300 BP by interpolation between the ^{14}C dates (Fig. 8.7). The first three fall within lacustral phases Ziway-Shala V and VI in the Rift: the last two may conceivably match up with Ziway-Shala VII. The base of the core corresponds to a smaller maximum which could be correlated with Ziway-Shala IV. The intervening acidic bog episodes would then correspond to intervals of drier, or more stable, climate. This is similar to the model recently proposed by

Aaby (1976) for the Danish raised bogs. The main problem with this hypothesis is that it does not explain the change from a 1800–2000 y periodicity to a 400-450 y periodicity around 3400 BP, which is associated with an apparent quadrupling of the sedimentation rate.

Alternatively, Dr. Gasse has suggested that the diatom peaks and troughs reflect an autogenic cycle of bog growth and degeneration, without climatic implications (Model 2). This model was first proposed by Sernander (von Post and Sernander, 1910) for bogs in northwest Europe. The strongest argument in favour is the cyclical occurrence of the acidophilous diatom assemblage as a function of depth. The change in apparent sedimentation rate at 3400 BP could reflect either a speeding-up of the cycle of bog growth and decay, or contamination of the basal sediments by root growth during the long dry interval which appears to have followed phase Ziway-Shala VI in the Rift (Fig. 8.1).

This controversy cannot be resolved at present. The ideal way to test the two hypotheses would be to obtain further cores from other bogs in the area to see if the cycles could be correlated from one to the other. Model 2 predicts a random distribution, whereas model 1 leads us to expect a consistent pattern related to the hydrological fluctuations which affected the Rift lakes. It is also important to discover whether the four-fold increase in mean sedimentation after 3500 BP is a widespread phenomenon.

IV. Soil formation

Evidence from the LSA archaeological sites studied by S.M.U. (Haynes and Haas, 1974; Laury and Albritton, 1975) strongly suggests that surface soils developed on pumice deposits in the Rift which lie outside the 1670 m shoreline began to form during lacustral phases Ziway-Shala IV and V. The S.M.U. dates tie in with widespread evidence for soil formation and slope stabilization in the Afar following the terminal Pleistocene arid phase (Williams and Dakin, 1976; Gasse, Rognon and Street, in press). Although we have no pollen evidence extending back before 11,500 BP, it seems likely that the two most important factors were increased soil moisture and vegetation cover, resulting from a dramatic improvement in the precipitation/evaporation ratio.

V. Conclusions and problems to be resolved

The first part of this chapter outlined the sequence of fluctuations recorded by the Rift Valley lakes during the Late Quaternary. These are summarized in Table 8.1. The information derived from the lakes is complemented by three lines of evidence from the catchment: geomorphic studies of the glacial moraines on the Arussi Mountains; pollen and diatom analyses of the Mount Badda Core; and soil stratigraphy on the Rift floor.

The new radiocarbon dates from the Mt. Badda Core confirm that the last glaciation of the Arussi Mountains took place prior to 11,500 BP. The glacial snowline lay at about 38[illegible]0 m, corresponding to a temperature lowering of 6 ^{o}C. It appears increasingly unlikely that this ice advance corresponded with the very dry Ziway-Shala III/IV interval. Gasse, Street and Rognon (in press) have suggested that it occurred during phase III, which seems to have provided an optimum combination of cooler and moister conditions (chapter 10). Glacial maxima were also experienced between 30,000 and 20,000 BP in Australia and tropical Mexico, where the subsequent period was also very dry (Bowler, 1975; Heine, 1977). Dating the outermost glacial moraines on Mt. Badda is therefore an important priority.

Pollen analyses of the Mt. Badda core suggest that vegetation changes may have played an important role in influencing runoff from the highlands. The main points to note are the persistence of patches of dry Podocarpus/ Juniperus/Olea forest throughout the last 11,500 years, despite the existence of a very dry climate in pollen zone B1; the expansion of this forest, including certain mesic elements, in zone B2; and the onset of human disturbance at the B2/B3 boundary, considerably earlier than in East Africa. The pollen assemblage from zone B1 suggests that conditions in Ethiopia during the early Holocene were much drier than in the Kenyan and Ugandan mountains (Hamilton, 1977a,b). If this difference also existed during the Ziway-Shala III/IV interval, then it is easy to explain the conspicuous poverty of the Ethiopian montane flora and faunas compared with East Africa (Hamilton, 1976).

An important controversy surrounds the timing of the B1/B2 boundary. According to the ^{14}C dates from Mt. Badda, this is situated at 6400 BP.

Dr. Hamilton proposes an age of 9500 BP on the basis of the East African evidence. In the wetter Balé Mountains, the main post-glacial forest period appears to have begun before 7900 BP, but it is quite conceivable that recovery of the forest proceeded at different rates on different massifs. If the ^{14}C dates from the Badda Core are accepted at face value, then vegetation conditions were more open during phases Ziway-Shala IV and V than during phase VI, which may have resulted in slightly greater runoff and erosion rates. The early introduction of agriculture, ca. 1850 BP, also raises the intriguing possibility that increased runoff and primitive cultivation techniques contributed, at least in part, to the Ziway-Shala VII lacustral phase (Pereira, 1962; Dagg and Blackie, 1965; Likens et al., 1970; Rapp et al., 1972; Pennington, 1975). So far the impact of pastoralism, which spread onto the Southeastern Plateau around 3500 BP (J.D. Clark, pers. comm.) has not been detected in the pollen record.

One important difference between the pollen and lake-level records is immediately apparent. The gradual changes in montane vegetation over periods of 1000-3000 years do not reflect the dramatic fluctuations in lake-level seen in Figure 8.1. This is not surprising in view of the slowness of vegetation migration and succession (Coope, 1977). The best indicator of hydrological conditions in the highlands may prove to be the highly sensitive diatom assemblages in lakes and bogs. The pioneer work by Dr. Gasse on the Mt. Badda Core has revealed cyclical oscillations from lacustrine to bog conditions. If these prove to have a climatic significance, they will provide the key to reconstructing changes in streamflow from the well-watered montane areas of the catchment.

CHAPTER NINE

Palaeohydrology of the Ziway-Shala Basin

9A SUMMARY

Past rainfall totals have been calculated for three different time periods by the water-balance method of C. Washbourn-Kamau (Washbourn, 1967), using as input estimates of past runoff, lake area and open-water evaporation. The results are shown in Table 9.2. They indicate that the "terminal Pleistocene" lowstand was linked with a significant reduction in rainfall and runoff, whereas Holocene lake-level maxima seem to have been caused by greatly increased rainfall. The model does not take into account the possible complicating effects of changes in cloud cover, seasonality, windiness or groundwater levels.

9B INTRODUCTION

Lake-level fluctuations are now recognized to be one of the most direct indicators of Late Quaternary climatic fluctuations in areas of internal drainage (GARP 1975; Street and Grove, 1976). The hydrological factors controlling the water budget of a closed lake are essentially simple, and have been recognized for more than 250 years. In the words of Edmund Halley (1715),

> Now I conceive that as all these lakes receive rivers, and have no exit or discharge, so it will be necessary that their waters rise and cover the land, until such time as their surfaces are sufficiently extended, so as to exhale in vapour that water which is poured in by the rivers; and consequently that lakes must be larger or smaller, according to the quantity of the fresh water they receive.

This relationship can be expressed by the following equation:

$$A_B P_B k + A_L P = A_L E \qquad (1)$$

where A_B is catchment area (km^2)

A_L is lake area (km^2)

P_B is mean annual precipitation over the catchment (mm)

P_L is mean annual precipitation on the lake surface (mm)

k is the runoff coefficient (the proportion of rain falling on the catchment which ultimately reaches the lake)

and E is mean annual evaporation from the lake surface (mm).

Past precipitation totals (P_B) can be estimated from equation (1) for any time intervals provided that the other terms can be reliably estimated. Since the pioneering paper by Leopold (1951), many attempts have been made to derive water budgets from ancient lakes in the Southwestern U.S.A. and Australia (Coventry, 1976; Brakenridge, 1978). C. Washbourn-Kamau was the first to apply this palaeoclimatic method in eastern Africa (Washbourn, 1967b). As in all other studies to date, she made the simplifying assumption that E is a function of temperature alone (in this case mean annual temperature, T_{ann}). The drawback of this approach is that it leads to considerable ambiguity when considering high lake stands which occurred during periods of lower temperatures: in these cases, it is difficult to distinguish the effects of increased precipitation from those of reduced evaporation. This has led to a long, fruitless and often acrimonious dispute between the advocates of the "pluvial" and "minevaporal" hypotheses (Galloway, 1965, 1970; Morrison, 1968; Reeves, 1973; Dury, 1973; Coventry, 1976; Brackenridge, 1978). Fortunately this problem does not arise in the present case, except for phases Ziway-Shala II and III (chapter 8B) which could not be reliably modelled anyway because of the difficulty of reconstructing the corresponding lake areas with sufficient accuracy.

Although it may eventually prove possible to develop other types of models, based for example on energy-balance considerations (Reeves, 1968, p.141; J.E. Kutzbach, pers. comm., 1978), the present study follows Washbourn-Kamau's approach (Street, in press, b). A simple FORTRAN programme, BUDGET, was written to solve equation (1) for any desired combination of the other variables.

9C THE WATER-BALANCE MODEL

I. Input data

The principal steps in the reconstruction of past climatic conditions are as follows (Washbourn, 1967b):

1) Modern data on climate, hydrology and basin characteristics are used to derive present-day values for A_L, A_B, P_L, P_B, E and k, and to estimate the relationship between E and mean annual temperature, T_{ann}.

2) Past lake depth and surface area are obtained from geomorphological, stratigraphic and topographic information.

3) Independent climatic and environmental data are then used to estimate past values of E and k which can be substituted in equation (1). This is probably the stage at which errors are most easily introduced.

4) Equation (1) is solved for P_B.

The input data used in the present study are given in Table 9.1 and Figures 9.1, 9.2. Three time periods were modelled: the terminal Pleistocene lowstand; the early Holocene highstand (Ziway-Shala V); and the late Holocene highstand (Ziway-Shala VII). No attempt has been made to simulate the water-balance of lakes Ziway-Shala II and III, since the results obtained would be too sensitive to the exact values of T_{ann} adopted. For phase V, the calculated value of P_B refers to the situation when the lake stood at the lip of its outlet, and is therefore a conservative estimate of the precipitation required to cause the lake to overflow.

The most difficult variables to estimate realistically for past time periods are T_{ann}, E and k. The sources used to estimate T_{ann} are given in Table 9.1. There is now some consensus that temperatures at the last glacial maximum were lowered by about 4-7^{o}C in eastern Africa. What is less certain is the time of onset of post-glacial warming. For 13,000 - 14,000 BP, estimates have been made for a 3^{o}C lowering and for the full 6^{o}C lowering suggested by the Mt. Badda data (Table 8.3). By the early Holocene it is believed that temperatures had risen to, or even above, their present levels, but conservative values are adopted here.

Table 9.1: <u>Data used to estimate past precipitation in the Ziway-Shala Basin</u>

Item		Source of data
Area of catchment[a] (A_B)	14,700 km^2	1:250,000 U.S. Defense Mapping Agency Series 1501, sheets NB 37-2 and 37-3, NC 37-14 and 37-15.
Area of lakes in 1969[b] (A_L)	1,190 km^2	Ditto
Area of 1670 m lake[b] (A_L)	2,690 km^2	Transferred to 1:250,000 map base from 1971-72 D.O.S. aerial photographs (scale 1:40,000 to 1:56,000)
Area of 1595 - 1600 m lake[b] (A_L)	1,590 km^2	
Mean annual rainfall over the catchment (P_B)	957 mm	Area - weighted average[b] from map in Kingham (1975)
Mean annual rainfall over the present lakes (P_L) Hence, $P_L = 0.698\ P_B$	668 mm	Rainfall-altitude regression for all stations in the Ziway-Shala Basin (calculated from data in Kingham, 1975)
Mean annual evaporation from the present lakes (E)	2,000 mm	Modern water-balance data. Makin <u>et al</u>. (1976)
Modern runoff coefficient for the catchment (k)	0.123	Calculated as residual in modern water-balance
Modern runoff coeffcient for Meki Basin	0.162	United Nations (1973)
Modern relationship between rainfall and runoff		Bauduin and Dubreuil (1973) for the Webi Shebeli Basin
Change in evaporation per 1 ^{o}C change in mean annual temperature, T_{ann}	133 mm	Monthly Colorado pan and temperature data from the Awash Basin (United Nations, 1965, vol. 3) using a pan coefficient of 0.81 (Hounam, 1973)
Temperature lowering in eastern Africa at the last glacial maximum	4-7 ^{o}C	Van Zinderen Bakker and Coetzee (1972); McCullough and Smith (1976); Williams <u>et al.</u> (1978) and Table 8.3.
Temperatures in eastern Africa during the early Holocene (T_{ann})	At or above present	Richardson and Richardson (1972); Gasse and Descourtieux (in press)

a. Area measured using 100 km^2 grid.
b. Area measured by weighing traced outlines.

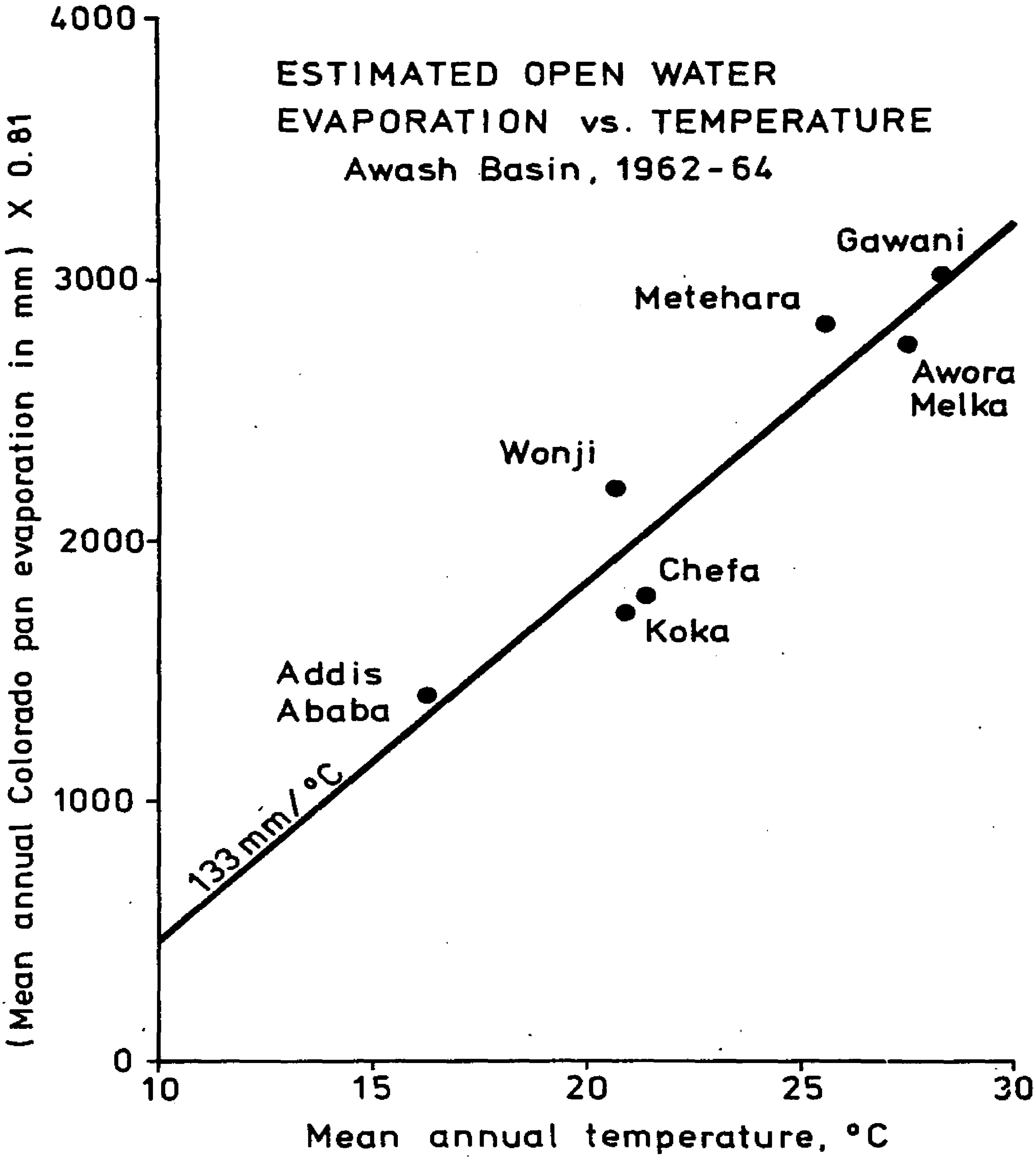


Fig. 9.1: Correlation between estimated mean annual open-water evaporation and temperature at stations in the Awash Basin.

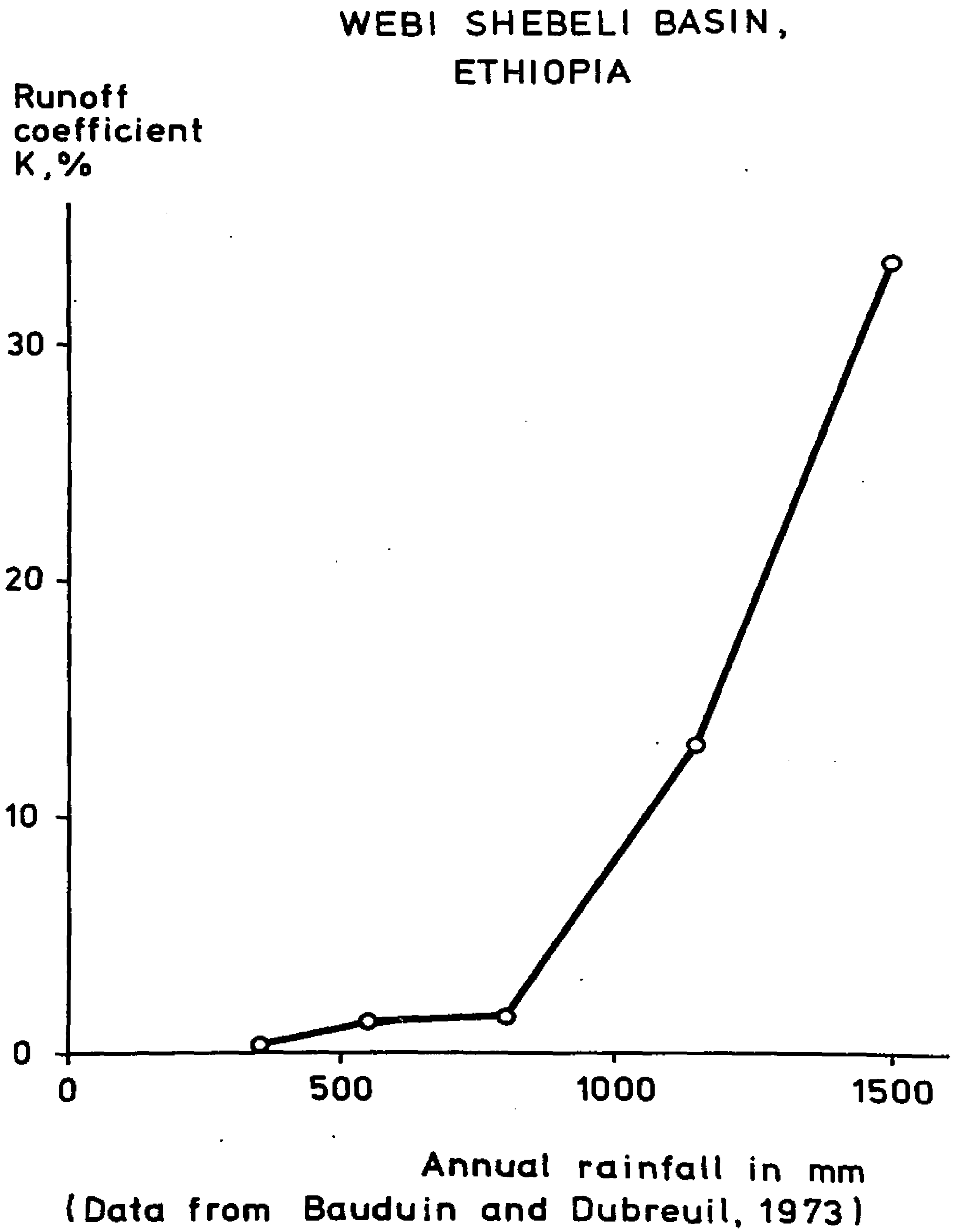


Fig. 9.2: Increase in the runoff coefficient (k) with mean annual rainfall in the Webi Shebeli Basin.

Previous studies in eastern Africa have generally interpolated open-water evaporation rates (E) from the graph complied by Langbein (1961) for lakes in the U.S. Some workers, for example Holdship (1976) have additionally made use of American data (Schumm, 1965) to derive runoff rates. The approach used here is strictly empirical. Rather than extrapolating relationships obtained in mid-latitude environments characterized by much higher relative humidity and seasonal contrast, modern evaporation-pan and temperature data from central and southern Ethiopia have been used to calculate values of E appropriate to past conditions. It has been shown by Dagg and Blackie (1970) that East African pan data yield estimates of open-water evaporation in good accord with the officially-recommended Penman formula (Dagg et al., 1970). The Colorado pan data used here consist of 26 months of carefully supervised observations by a United Nations team studying the adjacent Awash catchment (United Nations, 1965) (Fig. 9.1). Unfortunately, there is some uncertainty about the appropriate conversion factor for the sunken Colorado pan: the pan coefficient employed is the annual mean value recommended by W.M.O. (Hounam, 1973).

The annual mean evaporation gradient calculated from figure 9.1 is 133 mm/^{o}C. It varies seasonally from 63 to 211 mm/^{o}C. This figure can be compared with an average of 91 mm/^{o}C derived from data published by Bauduin and Dubreuil (1973) for a smaller number of stations in the neighbouring Webi Shebeli Basin. The difference between the two estimates is rather disturbing, but has less effect on the final results than the uncertainty in the choice of palaeotemperature values.

The runoff coefficient, k, may be estimated either from modern runoff data, or from the following equation:

$$k = \frac{P_B - U}{P_b} \qquad (2)$$

where U is mean annual basin loss (the depth of rainfall lost by evapotranspiration) (Dury, 1973, p. 3669). Since the most practical means of estimating U from T, the Thornthwaite method, is notoriously unreliable in

East Africa (Dagg and Blackie, 1970), the first solution was preferred.

Values of k for other East African lake basins of comparable size vary with climate and relief: from 0.0326 for Nakuru-Elmenteita (Washbourn, 1967b) to 0.40 for Lake Kivu (Hecky and Degens, 1973). The modern average for the Ziway-Shala Basin is 0.123. Figure 9.2 shows clearly that the runoff coefficient in the Webi Shebeli catchment varies exponentially with mean rainfall. This has the important consequence that any past fluctuation in precipitation would tend to be amplified by a corresponding change in k.

The pollen data shown in Figure 8.7 suggest that variations in vegetation and soil conditions may also have influenced runoff. The values selected for k (Table 9.2) attempt, albeit crudely, to incorporate these effects. During the terminal Pleistocene, the decrease in effective rainfall may have been partly compensated by a thinning-out of the vegetation cover. By the early Holocene, the montane forest had still not entirely recovered, but the effective rainfall had greatly increased and argillic horizons were developing in soils on the Rift floor surrounding the lakes (Table 8.1). The value chosen for k, 0.162, is representative of the steep and well-watered Meki and Upper Awash catchments at the present day. For the late Holocene, modern conditions have been assumed, although there is some uncertainty about the hydrological changes effected by forest clearance since 1850 BP (Figure 8.7).

II. Assumptions

In applying the Washbourn-Kamau method to the Ziway-Shala Basin, it is necessary to take account of certain explicit and implicit assumptions which may limit the value of the results. These are as follows:

a. Explicit assumptions

1) The basin has been topographically stable.
2) There are no signs of river capture.
3) The surface and subsurface catchments are coincident.
4) P_L is a known and constant function of P_B.
5) Seasonal variations in temperature have always been small, so that mean annual temperature, T_{ann}, can be used.
6) E is a known and constant function of T_{ann}.

Table 9.2: Palaeoprecipitation estimates for the Ziway-Shala Basin

Time Period	Lake Area (km^2)	Basin Open/ Closed	A_B/A_L	Palaeo-temperature (departure from present, ^oC)	Estimated Evaporation (mm)	Vegetation Cover (from Fig. 8.7)	k	% present precipitation
Present day	1190	Closed	11.4	0	2000		0.123	-
Late Holocene Maximum (phase Ziway-Shala VII) ca. 2000 BP	1590	Closed	8.25	0	2000	+ ?	0.123	122
Early Holocene Maximum (phase Ziway-Shala V) 8500-9400 BP	2690	Open	4.46	0 -2	2000 1734	- ?	0.162 0.162	147 128
Terminal Pleistocene Minimum 13,000 - 14,000 BP	1190	Closed	11.4	-3 -6	1600 1200	- - ?	0.100 0.100	91 68

\+ Greater than at present

\- Less than at present

7) Past values of k can be estimated with sufficient accuracy.

An additional assumption made here is that:

8) E can be reliably estimated using evaporation-pan data for stations at different elevations in the Awash catchment.

These assumptions have been discussed in chapter 2A and section I (above).

b. Implicit assumptions

It is further assumed that the following factors can be neglected, or have been adequately allowed for:

1) Changes in cloud cover and humidity through time (Kingham in Grove et al., 1975).
2) Changes in wind strength (Bowler, 1975).
3) Changes in the seasonality, intensity and form of precipitation (Rognon, 1976).
4) The effects of changing lake area on local precipitation (Hutchinson, 1973).
5) Changes in the storage of water in glacier ice (chapter 8CIIb).
6) Changes in flow routing and groundwater storage due to:
   - natural vegetation changes
   - pedogenesis
   - fluctuating groundwater levels (Brakenridge, 1978)
   - recent changes in land use (Pereira, 1962; Dagg and Blackie, 1965; Rapp, et al., 1972).
7) Changes in hot-spring activity (Hecky and Degens, 1973; Hecky, 1978).
8) The effects of thermal storage on lake-surface evaporation (Hounam, 1973).
9) The difference between evapotranspiration losses from permanent swamps and from open water (Bernatowicz et al., 1976; Balek, 1977).
10) The effects of changing salinity on lake-surface evaporation (Langbein, 1961).

The most important of these factors are probably 1, 2, 3 and 6. Kingham (op. cit.) has shown that fluctuations in the level of Lake Awasa since the middle 1950s can only be explained by changing evaporation losses, possibly coupled with a slight shift in the seasonal pattern of rainfall. This does not seem to be the case in the Ziway-Shala Basin, where variations in precipitation can adequately explain the recent lake-level record (chapter 5BIII), but long–term shifts in seasonality are quite conceivable, given that the basin lies near to the modern zone of transition between three different rainfall regimes (Bethke, 1976). This suggestion is reinforced by the greater extent of glaciation on the eastern side of Mt. Badda (Table 8.3) which may indicate a higher frequency of rain-bearing winds from the Indian Ocean (A. Hamilton, in litt. 1/11/78). Increased windiness at the glacial maximum has been inferred by many authors (Parkin and Shackleton, 1973; Bowles, 1975; Bowler, in press). Finally, geomorphic evidence from the Afar indicates that the rainfall regime there was more intense during the terminal Pleistocene, and less intense during the early Holocene, than it is today (Rognon, 1976).

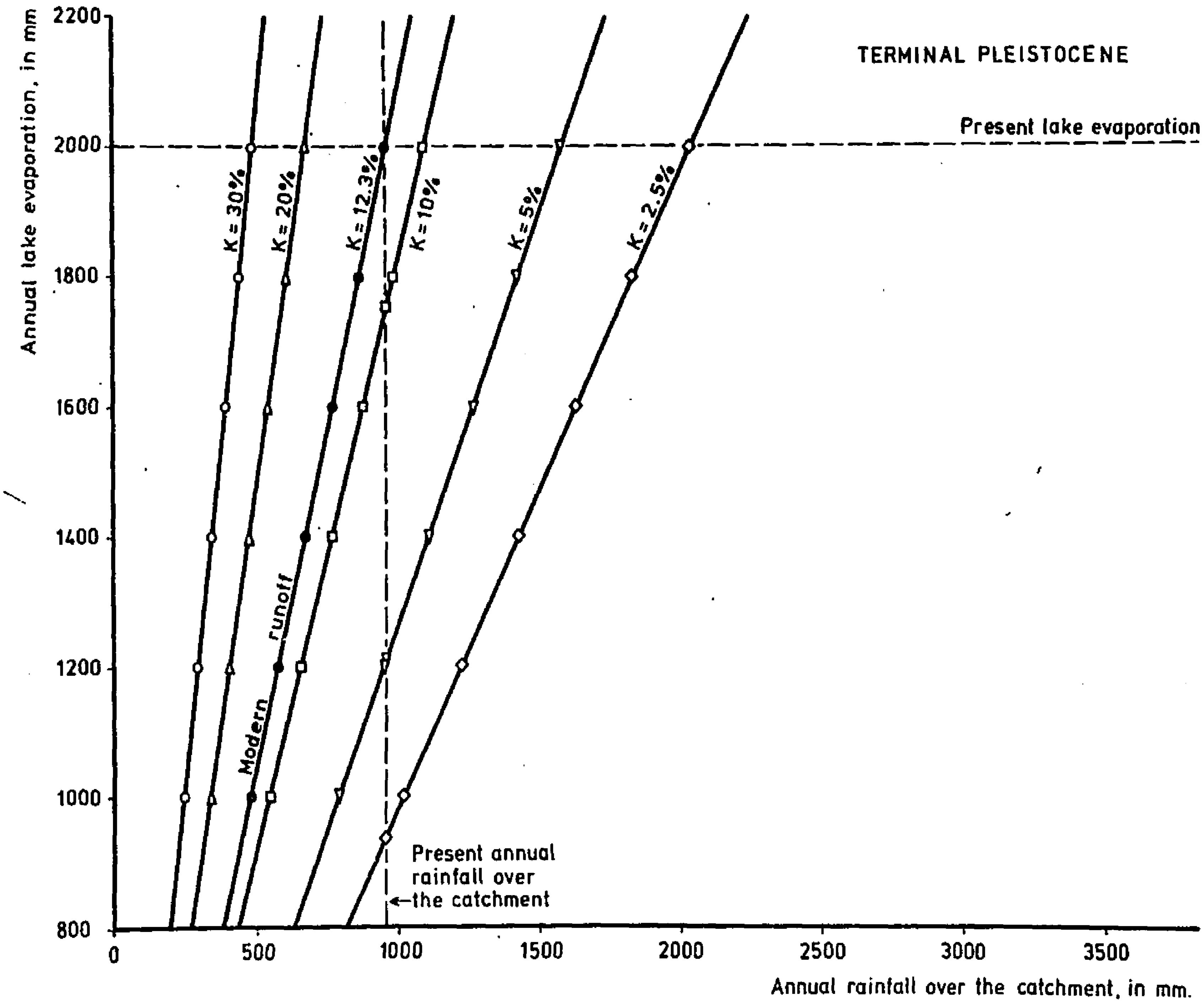
Changes in the proportion of rainfall reaching the lakes as surface and subsurface flow must have occurred, but are very hard to evaluate. Widespread fluctuations in groundwater levels in central Ethiopia have been inferred by a number of workers (United Nations, 1973; Gasse, Rognon and Street, in press). Higher water tables during the early and mid-Holocene probably increased the total area of swampland and the density of permanent drainage channels; which would tend to augment the amount of surface runoff and shorten the response time of the lakes.

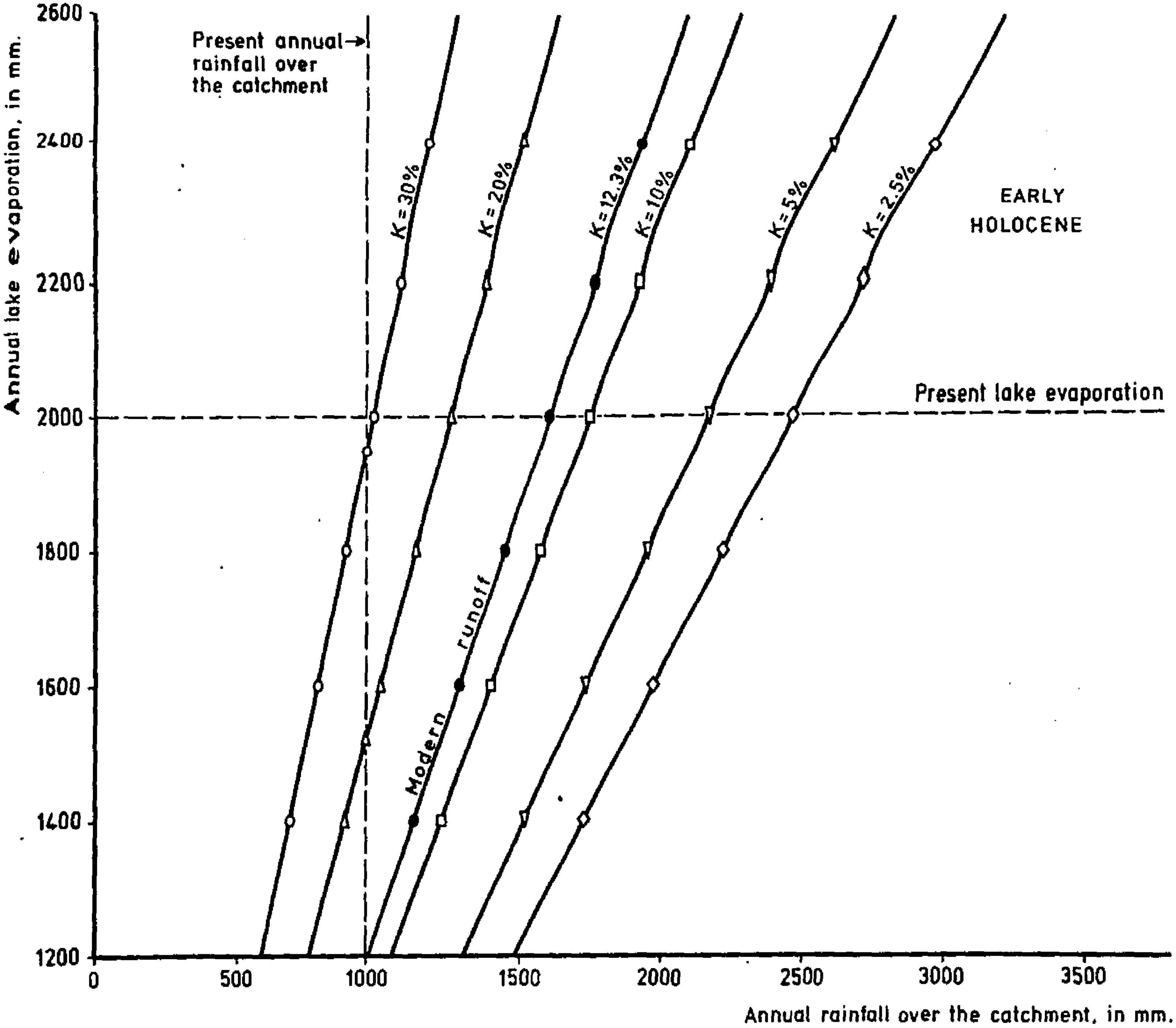
Despite all these reservations, the magnitude of the calculated changes in runoff is so great that large fluctuations in rainfall totals must be invoked to explain them.

III. Results

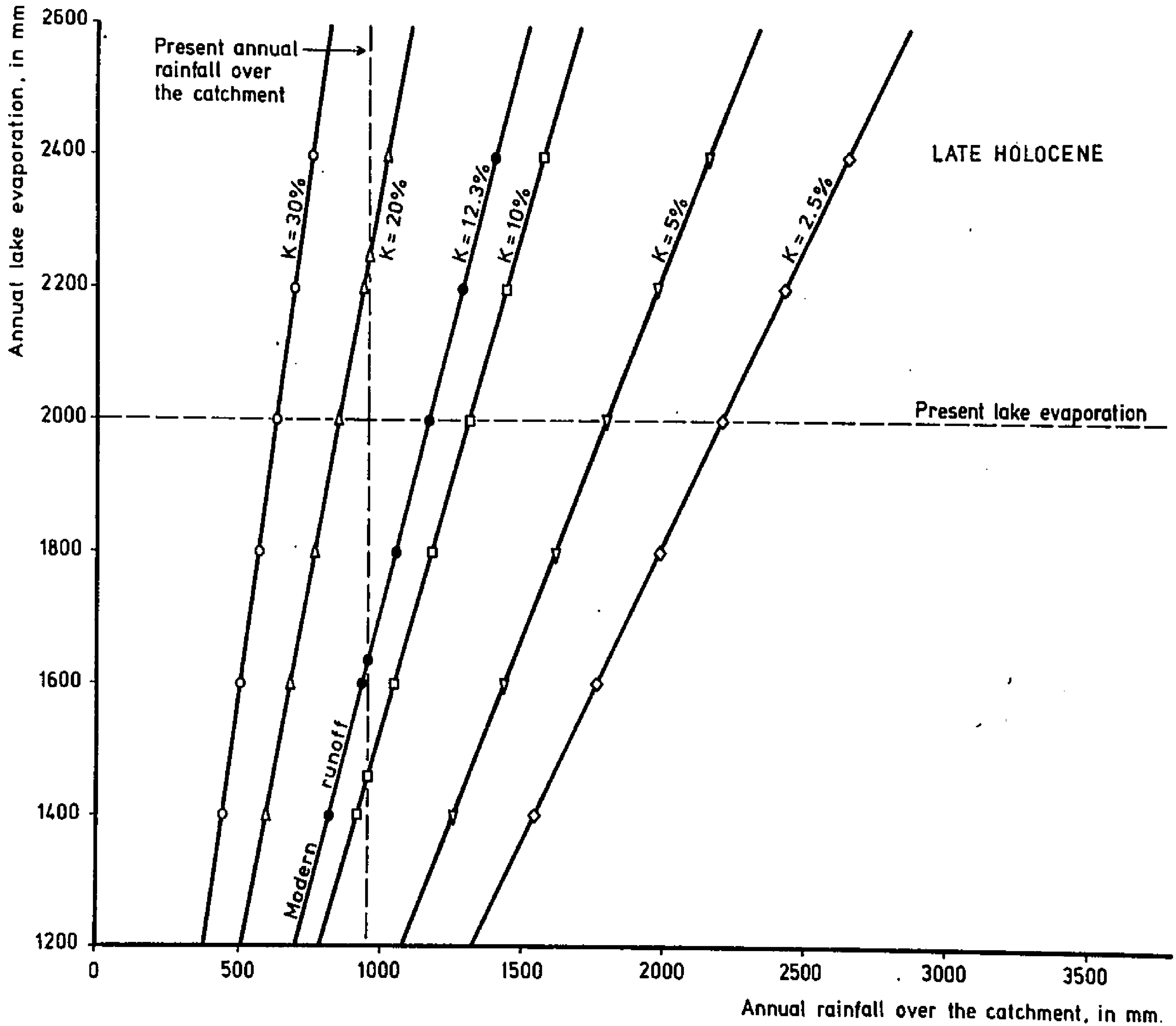
The computational results for the three selected time periods are shown in Figures 9.3 - 9.5. For the terminal Pleistocene case (Fig. 9.3), values of P_B much less than present are obtained unless k is drastically reduced, whereas in the early Holocene model (Fig. 9.4) only very low values of E or unrealistically high values of k would yield lower precipitation totals than today. The same conclusion applies to the mid-Holocene maximum. The additional losses through the Meki-Dubeta outlet would raise the calculated values of P_B still higher. The late Holocene model (Fig. 9.5) is similar to Figure 9.4, but as might be expected yields lower values of P_B.

Figs. 9.3 to 9.5: Terminal Pleistocene, early Holocene and late Holocene water-balance models. Combinations of annual rainfall and evaporation adequate to sustain the lakes, given specified values of the runoff coefficient (k).


TERMINAL PLEISTOCENE
Present lake evaporation
K = 30%
K = 20%
K = 12.3%
K = 10%
K = 5%
K = 2.5%
Modern runoff
Present annual rainfall over the catchment
Annual lake evaporation, in mm
Annual rainfall over the catchment, in mm.
2200
2000
1800
1600
1400
1200
1000
800
0
500
1000
1500
2000
2500
3000
3500



Present annual rainfall over the catchment
Annual lake evaporation, in mm.
2600
2400
2200
2000
1800
1600
1400
1200
0
500
1000
1500
2000
2500
3000
3500
Annual rainfall over the catchment, in mm.
K=30%
K=20%
K=12.3%
K=10%
K=5%
K=2.5%
EARLY HOLOCENE
Present lake evaporation
Modern runoff



Annual lake evaporation, in mm
2600
2400
2200
2000
1800
1600
1400
1200
0
500
1000
1500
2000
2500
3000
3500
Annual rainfall over the catchment, in mm.
Present annual rainfall over the catchment
LATE HOLOCENE
Present lake evaporation
K = 30%
K = 20%
K = 12.3%
K = 10%
K = 5%
K = 2.5%
Modern runoff

Table 9.2 gives the best estimates of P_B for each time period, given the previously selected values of E and k. These results are compared with evidence from other parts of East Africa in Table 10.2. The principal remaining area of uncertainty surrounds the conditions prevailing during phases Ziway-Shala II and III, when both lower temperatures and increased precipitation may have contributed to the observed lake-level maxima.

CHAPTER TEN

Comparison With Other Lakes In The Horn of Africa

10A SUMMARY

The Late Quaternary history of Lake Ziway-Shala shows a striking resemblance to the history of other lakes which are hydrologically dependent on the Ethiopian Highlands. This is apparent not only from its record of lake-level fluctuations, but also from the variations in sedimentation and diatom assemblages through time, which are closely analogous to those experienced by the Central Afar lakes, especially Abhé (Gasse and Street, 1978 a,b; Gasse, Rognon and Street, in press). This overall similarity is a clear indication of the regional extent of the climatic fluctuations which affected the Horn of Africa during Late Quaternary time. Lake Ziway-Shala is hydrologically akin to Lakes Abhé and Turkana in that it receives most of its inputs from large highland rivers rather than from direct rainfall or groundwater resurgence. This helps to explain its very similar pattern of lake-level response. The greatest differences exist between Lake Ziway-Shala and spring-fed lakes such as Asal and Afrera. The latter are situated at much lower altitudes. Their environmental history has been greatly influenced by the hydrogeological factors which determine the source, volume and chemistry of their subsurface inflows, and by the smoothed response of the aquifers to climatic change.

10B BRIEF DESCRIPTION OF OTHER LATE QUATERNARY LAKES IN THE HORN OF AFRICA

I. The Awash system

a. River-fed lakes

During humid episodes of the Late Quaternary, when it overflowed, Lake Ziway-Shala became the highest in a chain of lakes situated in the Main Ethiopian Rift along the course of the Awash. These included Lakes

Galila (Koka), Wonji and Gawani (Taieb, 1974), which developed by ponding of the river flow upstream of the modern gorge sections; and a large terminal lake (Abhé) (Gasse, 1975, 1977a; Gasse and Delibrias, 1977). At maximum the water-surface area of all these lakes totalled about 10,000 km^2 (Table 10.1). They received the runoff from a substantial part of the Ethiopian and Southeastern plateau margins (Figs. 1.1, 8.3).

b. Spring-fed lakes

At the present day, the Awash River undergoes considerable losses by effluent seepage between Koka (1589 m.a.s.l.) and Lake Abhé (240 m) (United Nations, 1965, 1973). The regional pattern of groundwater flow is directed NE into the Central Afar, towards the lowlying Dobi-Hanlé, Gaggade and Asal grabens. The present Lake Asal (-155 m) is also fed by seawater inflows from the nearby Gulf of Tadjurah (Gulf of Aden).

In the past, lakes developed in a number of fault troughs peripheral to the present Awash drainage. These grabens are now either dry or occupied by salt lakes. The palaeolakes included Besaka (present elevation ca. 955 m) (Williams and Dakin, 1976; Williams et al., 1977), Dobi-Hanlé and Asal. Lake Besaka was located close to the present river about 125 km east of Addis Ababa. Dobi-Hanlé and Asal were situated in the Central Afar between Lake Abhé and Djibouti (Fig. 1.1). Although direct rainfall and surface inputs were probably more important during lacustral phases than they are today, the geological evidence suggests that during maxima these lakes all received substantial inflows of groundwater originating in the Awash catchment.

II. The Afrera Basin

Lake Afrera (-80 m) lies in a small, closed depression in the Northern Afar, surrounded by volcanic ranges. At the present day it is fed almost entirely by resurgences of groundwater originating from the northern end of the Ethiopian Escarpment, making it hydrologically independent of the Central Afar lakes (United Nations, 1973).

Table 10.1: Geometry of Rift and Afar Lakes During Late Quaternary Highstands

	Water surface elevation (m.a.s.l.)	Increase in max. depth* (m)	Surface area (km^2)	Approx. increase in volume* (km^3)
LATE PLEISTOCENE				
Abhé I	> 400	>160	⩾ 5500	
Abhé II	> 390 to 400	>150 to 160	⩾5500	
Abhé III	> 410	> 170	⩾6000	
Ziway-Shala III	~1680?	~ 122?	~ 3000?	
EARLY HOLOCENE (ca. 9000 BP)				
Ziway-Shala	1670	112	2690	160
Abhé	400	160	5500	315
Dobi-Hanlé	300	200	1100	170
Asal	160 to 170	315 to 325	>1000	100
Afrera	-50 to -40	30 to 40	310	12
LATE HOLOCENE (ca. 2500-1500 BP)				
Ziway-Shala	1595	37	1590	23
Abhé	314	75	1100	55 to 60
Dobi-Hanlé	> 235	>135	>800	> 90

* relative to modern lake-level

(from Gasse and Street, 1978a, with additions)

III. The Abaya-Chamo-Chew Bahir-Turkana-White Nile system

The southern part of the Main Ethiopian Rift today forms two internal drainage basins; occupied by Lake Awasa and by the Abaya-Chamo-Chew Bahir chain of lakes respectively (Grove et al., 1975) (Table 2.3). Very little is known about the history of the Awasa catchment. The terminal playa of the Abaya system, Chew Bahir, appears to have overflowed during highstands into Lake Turkana (present elevation 375 m) (Fig. 1.1). At the present day, Lake Turkana is fed largely by the Omo River, which rises on the Ethiopian Plateau near Addis Ababa. During its early and mid-Holocene maxima, it also received the overspill from extensive lakes in the Chalbi Desert to the east (D. Phillipson, pers. comm. 1975) and the Suguta Valley to the south (Bishop, 1975). As a result, it expanded to cover an enormous area - roughly 38,500 km^2 compared with its present 7,500 km^2 (Yuretich, 1976) - and seems to have communicated in its turn with the White Nile via the Pibor-Sobat drainage (Butzer et al., 1972; Butzer, 1976). This means that the Ethiopian Plateau NW of Butajira (Fig. 2.1) was at times contributing runoff to the Awash River, Lake Ziway-Shala, and the White Nile.

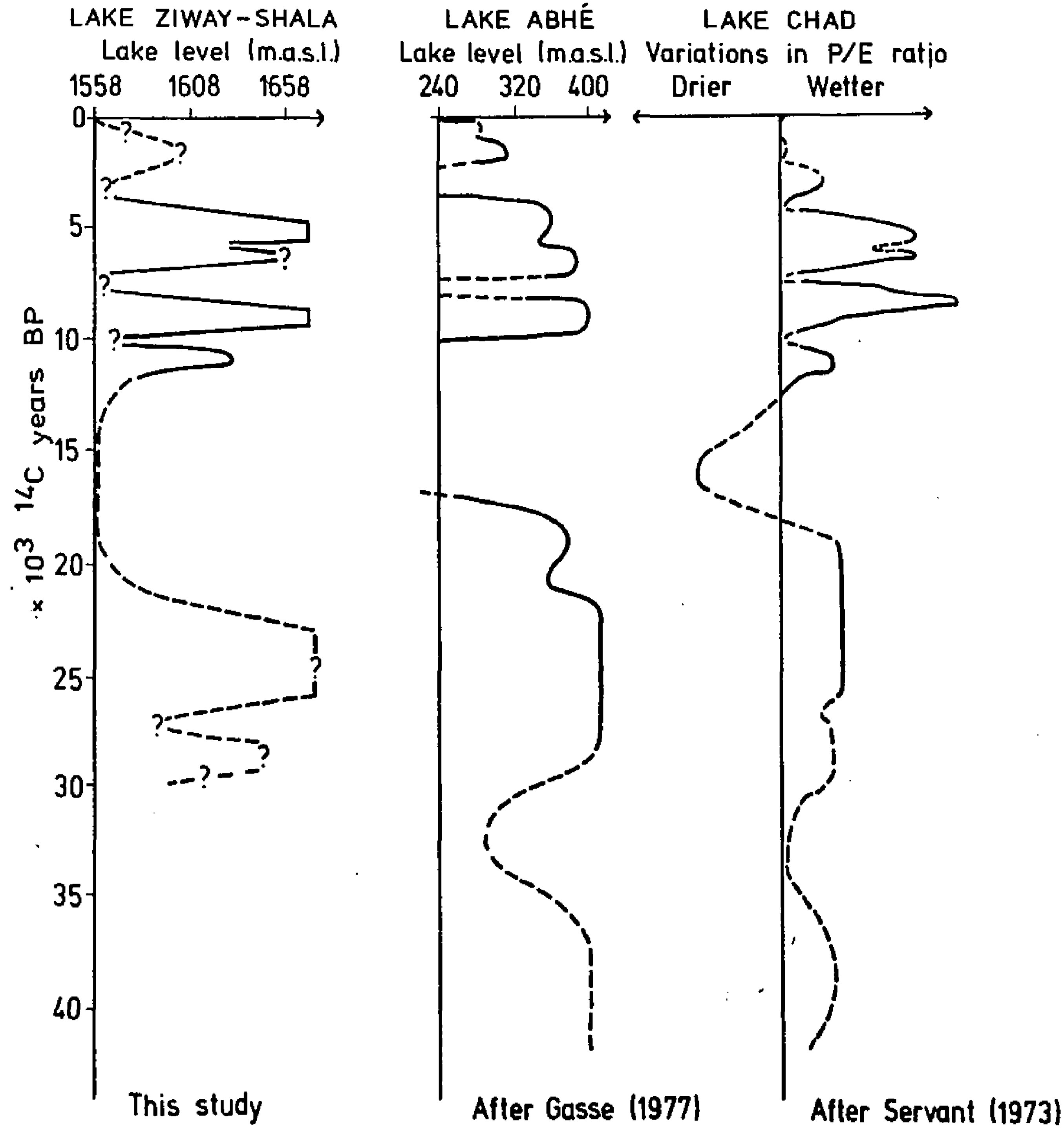
10C COMPARISON OF LAKE-LEVEL FLUCTUATIONS AND PALAEO-ENVIRONMENTS

In previous papers, Dr. Françoise Gasse and I have compared the history of water levels and sedimentation in the Ziway-Shala Basin with the record of a number of other lakes in Ethiopia and Djibouti (Gasse and Street, 1978 a,b; Gasse et al, in press). This section briefly summarizes our main findings, extending the discussion to include Lake Turkana (Figs. 10.1, 10.2). Figure 10.1 also compares the water-level curve for Lake Ziway-Shala with Lake Chad (13°N) (Servant, 1973), which is representative of the Southern Sahara (Faure, 1969; Chamard, 1972; Street and Grove, 1976; Servant et al., 1976; Rognon and Williams, 1977).

I. The complex Late Pleistocene lacustral interval

The most complete sequence for this period is derived from the Abhé Core (Gasse, 1975, 1977a; Gasse and Delibrias, 1977) (Fig. 10.1). Three

Fig. 10.1: Water-level curves for Lakes Ziway-Shala and Abhé since 40,000 BP, compared with past variations in the precipitation/evaporation ratio in the Chad Basin (derived from the fluctuations of Lake Chad).



LAKE ZIWAY-SHALA
Lake level (m.a.s.l.)
1558 1608 1658
LAKE ABHÉ
Lake level (m.a.s.l.)
240 320 400
LAKE CHAD
Variations in P/E ratio
Drier Wetter
0 5 10 15 20 25 30 35 40
× 10³ ¹⁴C years BP
This study
After Gasse (1977)
After Servant (1973)

distinct lacustral phases: Abhé I (100,000 - 70,000 BP?), Abhé II (>40,000 - 30,000 BP?) and Abhé III (>27,000 - 17,000 BP) have been inferred. These were separated by partial or complete desiccation of the lake. Abhé I has been dated by extrapolating the sedimentation rate from the upper part of the core. Abhé II lies at the limit of ^{14}C and the results obtained so far are inconsistent, making the duration of the Abhé II/Abhé III regression rather uncertain (Gasse and Delibrias, 1977). But the timing of phase Abhé III is now well established. Three lacustral episodes (Asal I-III) can also be distinguished stratigraphically in the Asal Basin, although they have not yet been dated. Only the second and third, however, are recorded in the Dobi-Hanlé graben, where the "Abhé III" maximum is dated 23,600 ± 650 BP (Gasse, 1975). No definite evidence relating to this time period is known from the other Afar basins, although undated fossiliferous clays of probable Late Pleistocene age occur around Lake Besaka (Williams and Dakin, 1976).

In the Omo Valley, at the north end of Lake Turkana, Butzer et al. (1969, 1972) have failed to find any trace of lacustrine sediments with ^{14}C ages falling between 37,000 and 10,000 BP. They interpret this hiatus to mean that lake level was low throughout this interval but their argument is not buttressed by any dates on nonlacustrine sediments. To the east of the present lake, erosion of Holocene lake beds is proceeding so fast under the present arid conditions that the lack of upper Late Pleistocene evidence in the Omo Valley is not at all surprising. In all probability, this controversy will only be resolved through the study of continuous cores from the centre of the lake.

As yet it would be premature to correlate phases Abhé I and Abhé II with events in the Ziway-Shala Basin; especially as phase Ziway-Shala II was preceded by an interlacustral interval lasting an estimated 40,000 - 50,000 years, which is not matched in the Abhé sequence (chapter 8BIIb). However, the available radiocarbon dates suggest that phases Ziway-Shala II and III both fall within the time span of phase Abhé III (Table 8.1; Fig. 10.1). This correlation is supported by the striking similarity in sedimentation and diatom floras between the two lakes (Gasse, 1975, p. 369; Gasse and Street, 1978a;

Gasse and Descourtieux, in press). The main lake-level maximum is dated 25,000 - 23,000 BP in the Abhé Basin, with a secondary maximum after 20,800 BP; and < 27,000 to ca. 21,000 BP in the Deka Wede area (chapter 6CII). But the dates quoted for the Ziway-Shala Basin should be regarded as provisional.[1] At present it appears highly unlikely that lacustral conditions persisted in the latter as late as 17,000 BP, when Lake Abhé finally dried up (Gasse, 1977a; Gasse and Delibrias, 1977).

As a whole, the stratigraphic evidence suggests that the Late Pleistocene lake in the Ziway-Shala Basin was less stable than Lake Abhé III, but that both were larger than during any subsequent highstand (Table 10.1). The ecological resemblance noted above seems to be a reflection of regional climatic and vegetation conditions, but was probably accentuated by the existence of a hydrographic connection (Gasse and Street, 1978a). Both lakes were dilute and probably cooler than at the present day. Both deposited thick clayey diatomites characterized by Melosira spp and/or Cyclotella ocellata (Gasse and Descourtieux, in press). Very similar floral assemblages are found in sediments of the same age in the Dobi-Hanlé graben. Dr. Gasse and I have inferred from the predominance of biogenic sedimentation that the Ethiopian Highlands were subject to a more regular, lower-intensity rainfall regime than today; although there are as yet no reliable water-balance estimates available. We also suggested that the slopes were protected by a denser vegetation cover. This geomorphic stability was not, however, as marked as during the earlier phase Abhé II (Gasse and Street, 1978a,b; Gasse et al, in press).

II. The "terminal Pleistocene" arid interval

The prolonged period of desiccation which followed has left extensive traces throughout the Afar and the Main Ethiopian Rift (Gasse et al., in press) It lasted from 17,000 to 10,000 BP in the Abhé Basin and from ca. 21,000 to ca. 11,500 BP in the Bulbula Plain. Lake Abhé dried up completely. Enormous

1. See footnote on p. 351.

alluvial fans consisting of angular cobbles and boulders formed at the foot of the major fault escarpments in all the Central Afar grabens, while sand dunes were active in the most arid areas. In the more humid Southern Afar and Main Rift, this interval is commonly represented by fine-grained colluvial or alluvial deposits; with the exception of the Bulbula Plain, where the Abernosa pumice member records a prolonged local phase of explosive volcanism. The lakes in the Ziway-Shala Basin seem to have been very low. This is inferred from the degree of drainage incision, since the minimum level of the terminal lakes Abiyata and Shala has not yet been determined.

According to the water-balance calculations presented in chapter 9, annual precipitation totals during this interval were some 9 - 32% lower than at present, depending on the palaeotemperature values adopted. Corresponding estimates of the reduction in East Africa range from 20 to 46% (Table 10.2). Professor Pierre Rognon argues that the rains in the Afar must have been more intense than today in order to produce sheetflood depths sufficient to transport huge volumes of coarse, bouldery material across very gentle pediment surfaces (Rognon, 1975a; Gasse et al., in press).

III. The complex Holocene lacustral interval

The re-expansion of the lakes at the beginning of the post-glacial is dated at different times in different basins. In the Ziway-Shala Basin, the first rise in lake level occurred between ca. 11,500 and 10,200 BP, whereas there is no evidence that Lakes Abhé, Turkana or Afrera responded before 10,000 BP (Fig. 10.2). The earliest dates on high lake-levels in the other spring-fed basins are 11,430 ± 380 BP (Besaka), 12,930 ± 750 BP and 10,870 ± 220 BP (Dobi-Hanlé) and 9920 ± 320 BP (Asal) Fontes et al., 1973; Gasse, 1975; Williams et al., 1977). Although it is possible to suggest that these older measured ages are an artifact of contamination by dead carbon in groundwater (Gasse et al, in press), the charcoal dates from the Ziway-Shala Basin suggest that a short wet phase did in fact occur before 10,000 BP. This fits in with a date of 11,070 ± 160 BP obtained by Williams et al. (1977) from shells in a spring tufa in the Southern Afar; implying that the aquifers there were being actively recharged from the Southeastern Plateau.

Table 10.2: Comparison of Palaeoprecipitation Estimates from Lakes in Eastern Africa

	Reference	Assumed Temperatures	% Present Precipitation
TERMINAL PLEISTOCENE			
Ziway-Shala	(this study)	$3^{o}C$ lower $6^{o}C$ lower	91 68
Nakuru-Elmenteita	1		<80
Kivu	2	$3^{o}C$ lower	54
Victoria	3	$5^{o}C$ lower	<71
EARLY HOLOCENE			
Ziway-Shala	(this study)	Present $2^{o}C$ lower	>147 >128
Nakuru-Elmenteita	4	$2\text{-}3^{o}C$ higher	>165
Naivasha	4	Present	$\gg 125$
Manyara	5	Present	133

References:

1. Washbourn (1967b).
2. Hecky and Degens (1973).
3. Harvey (1976).
4. Butzer *et al*. (1972).
5. Holdship (1976).

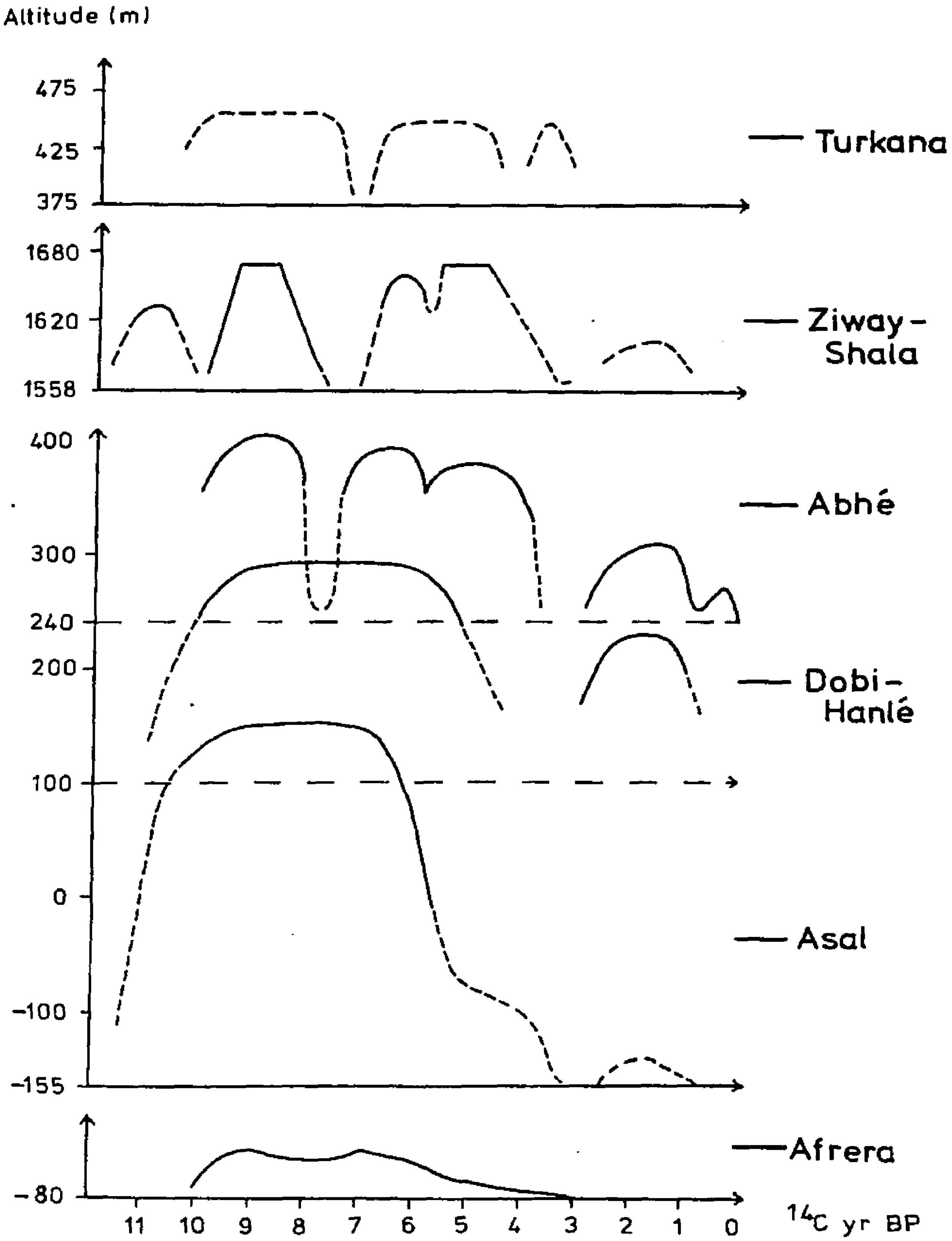


Fig. 10.2: Comparison of water-level curves for lakes fed directly or indirectly from the Ethiopian Highlands (data from Butzer et al., 1972 and Gasse and Street, 1978a, updated).

But here again the possibility of "zero error" cannot be eliminated. The rather equivocal evidence relating to soil, vegetation and runoff conditions in the Ziway-Shala Basin has already been described (chapter 8C). Clearly a great deal more research is needed in order to clarify the climatic and hydrological changes during the period 12,000-10,000 BP.

There is much less disagreement about the interval 10,000-4,000 BP. The large river-fed lakes Ziway-Shala, Abhé and Turkana experienced two major highstands (Abhé IV and V) interrupted by a very pronounced regression. This is dated 8500-6500 BP, 8400-7200 BP and 7900-7000 BP[1] respectively (Street, in press, a; Gasse and Street, 1978a; Butzer, 1976) (Fig. 10.2). The level of Lake Abhé apparently fell again for a brief period centred on 5900-5800 BP. This may reflect the same shortlived climatic deterioration as the Roricha fish bed, which has an age of approximately 5700 BP (chapter 6DIIb). The major regression marking the end of the mid-Holocene lacustral phase set in around 4800 BP in Lakes Ziway-Shala and Turkana and 4000 BP in Lake Abhé.

The groundwater-dependent lakes apparently remained high throughout the period 10,000 to 6,000 BP: there is no evidence for a regression between 8,000 and 7,000 BP except in the Afrera Basin, and even there it was greatly attenuated (Gasse, 1975; Williams and Dakin, 1976). This lack of sensitivity can be directly attributed to the damped response of the aquifers to climatic fluctuations of short duration (Gasse and Street, 1978a; Street, 1979). In these basins the onset of desiccation varied from 6,000 to 4,000 BP.

The interval from 10,000 to 4,000 BP was therefore characterized by increased runoff from both plateaus and from the Afar massifs (Gasse, 1975; Fontes and Pouchan, 1975); higher water tables (Williams and Dakin, 1976, p.56); and greatly enlarged lakes (Table 10.2). Montane forest gradually

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1. 400 years have been added back onto the Turkana dates to make them comparable with the ages for Ziway-Shala and Abhé, which are uncorrected for zero error (see Butzer <u>et al.</u>, 1969).

expanded on the Arussi and Balé Mountains (Hamilton, 1977a,b) (Fig. 8.7), while swamp soils developed in areas of poor drainage (Semmel, 1971; Williams and Dakin, 1976). During this period the lakes were characterized by high water temperatures and high planktonic and littoral productivity (Gasse et al., 1974; Gasse and Street, 1978a). Most were very dilute. Two exceptions were the small, spring-fed lakes Besaka and Afrera, which seem to have been rather brackish, although they supported thriving communities of freshwater mollusca (Gasse, 1975; Gasse et al., in press). Many of the larger lakes were fringed by extensive belts of thicket and reed swamp, which provided habitats for epiphytic diatoms, sponges, pulmonate snails, fish and freshwater crabs. In mid-basin, their sediments consisted essentially of calcitic or aragonitic marls or calcareous clays, which are associated with a high abundance of the planktonic diatom Stephanodiscus astraea and its varieties. This dominance of biochemical over clastic sedimentation fits in well with the general picture of increased vegetation cover and slope stabilization (Gasse and Street, 1978a).

The much greater extent of the lakes during this period of relatively high temperatures implies substantially greater rainfall and runoff. In chapter 9 the required increase in precipitation was calculated to be of the order of 28-47% (Table 9.2). This compares with published estimates of 25-65% for East Africa (Table 10.2).

During the late Holocene there is less similarity between the history of different basins (Fig. 10.2). This may reflect both a reduction in the scale of the underlying climatic patterns (Street and Grove, 1976) and also the increasing control exerted by local hydrogeological factors on water-level fluctuations and sedimentation (Gasse and Street, 1978a). Lake Turkana experienced a final transgression to ca. 445 m around 3650 $\pm$ 150 BP (Butzer et al., 1972). This is similar to many of the shallow East African lakes, which fell to very low levels, or even dried out, between 4500 and 3700 to 3000 BP, and then subsequently recovered (Richardson, 1972; Richardson and Richardson, 1972; Holdship, 1976; Viner, 1977). Many of the Saharan lakes such as Chad (Fig. 10.1) also experienced a minor maximum between 3500 and 3000 BP (Servant, 1973; Street and Grove, 1976). In Ethiopia and

Djibouti, on the other hand, there is no sign of even a partial recovery until after 2700 BP. In these areas the late Holocene lacustral phase is dated 2700 - 1000 BP (Abhé VI), 2500 - 1500 BP (Dobi-Hanlé) and < 2500 BP (Ziway-Shala VII). The corresponding fluctuation in the Asal Basin has not been dated, because the lake remained too saline for mollusca (Gasse, 1975). The other lakes were shallower (Fig. 10.2), more unstable and probably somewhat more alkaline than during their early and mid-Holocene maxima. Stable-isotope measurements by Fontes (Gasse et al., 1974; Fontes and Pouchan, 1975) indicate that Lake Abhé was once again receiving an influx of dilute runoff from the highlands. The sediments deposited during the late Holocene maximum are largely clastic in the basins fed by large rivers (Ziway-Shala, Abhé), but chemical or biochemical in groundwater-dependent basins such as Dobi-Hanlé and Asal (Gasse and Street, 1978a).

During the last millennium, conditions have been very unstable. The recent shorelines of Ziway-Shala, Abhé and Turkana record innumerable small fluctuations in lake-level, continuing right up to the present day.

10D CONCLUSIONS

The Late Quaternary history of the Ziway-Shala Basin, which has been reconstructed entirely from internal evidence, shows numerous parallels with the history of other lakes in the Horn of Africa, and more generally, with the whole of intertropical Africa between about 5^{o} and 15^{o} N (Chamard, 1972; Servant, 1973; Street and Grove, 1976; Rognon and Williams, 1977). The closest similarities are with large river-fed lakes such as Abhé, Turkana and Chad. Even among this group, Ziway-Shala was unusually sensitive. It occupied a particularly favourable altitudinal situation for recording past climatic fluctuations: high enough to register the slightest change in runoff from the montane areas, but low enough to remain without an outlet for long periods of time.

The relationship of high lake levels to mountain glaciation in eastern Africa is still not entirely clear, due to the lack of ^{14}C dates for the glacial maximum. But the evidence from Lake Ziway-Shala confirms that the third Late

Pleistocene lacustral interval previously recognized in the Abhé Basin (Abhé III, >27,000 - 17,000 BP) falls within the mid-Wisconsin (mid-Weichselian) interstadial of high latitudes (Dreimanis and Raukas, 1975). Furthermore, the beginning of the Holocene lacustral interval corresponded to the onset of post-glacial warming (Vincent, 1972). The intervening "terminal Pleistocene" and phase now appears to be the tropical counterpart of the glacial maximum in high latitudes, which was centred on 18,000 BP (CLIMAP Project Members 1976). This is supported by the peak of cold-loving diatoms in the Abhé and Abiyata Cores (Gasse and Delibrias, 1977; Gasse and Descourtieux, in press). The full implications of this cool, windy and arid episode for the vegetation, fauna and prehistory of the Ethiopian Highlands still remain to be explored.

GENERAL CONCLUSIONS

1) The four present-day alkaline lakes in the Ziway-Shala Basin are the shrunken remnants of a single lake which at times overflowed into the Awash River via the Meki-Dubeta col (ca. 1670 m). During its early and mid-Holocene maxima Lake Ziway-Shala (2690 km^2) was intermediate in area between the present "great lakes" Idi Amin Dada (Edward) and Kivu, and almost equalled Lake Turkana in volume.

2) The two most prominent shorelines in the Ziway-Shala Basin are situated at 1670 m and 1595-1600 m. The first corresponds to the early and mid-Holocene outlet level and the second to a late Holocene maximum which only united the three southern lakes. Neither shows any significant tectonic or isostatic deformation.

3) Deep exposures of lacustrine and intercalated subaerial deposits on the Rift floor confirm and extend the sequence of lake-level fluctuations derived from shoreline evidence. No one locality contains a complete succession, since the rapidity and large amplitude of the water-level oscillations often resulted in localized erosion or nondeposition. The composite sequence given in Tables 6.9 and 8.1 is based largely on the evidence from three key localities: the Deka Wede Valley, the Abiyata Core, and the Ajewa area.

4) Although lakes existed in the Ziway-Shala Basin as long ago as 1.05 m.y., their geometry can no longer be reconstructed. The Shala caldera was first flooded by a deep lake during the upper Middle Pleistocene or lower Late Pleistocene. At about this time, a shallow, brackish lake also formed to the east of the present Lake Ziway. Owing to later erosion and faulting, it is uncertain whether either of these events had any climatic significance. By the upper Late Pleistocene, however, the lake-level fluctuations can be confidently interpreted in climatic terms. Four major highstands and six minor ones have occurred since about 30,000 BP (Table 8.1). These can be grouped into a complex Late

Pleistocene lacustral interval (ca. 30,000 to ca. 21,000 BP) and a complex Holocene lacustral interval (ca. 11,500 to 4800 BP); separated by a prolonged period of aridity. Since 4800 BP, lake levels have remained low and fluctuating, apart from a brief late Holocene maximum after 2500 BP.

5) During highstands, Lake Ziway-Shala was very dilute, and ecologically similar to the contemporary Lake Abhé. Its deep-water deposits are mainly diatomites, diatomaceous muds or marls. The mollusc and fish faunas suggest comparisons with large freshwater lakes such as Mobutu Sese Seko (Albert) and Abaya, although they were less diverse.

6) During the last glaciation of the Arussi Mountains, the snowline was lowered by an estimated 900 m, corresponding to a temperature depression of 6°C. The extent of ice in the catchment was too small to have direct hydrological effects on the lakes. Deglacierization of the summits was complete before 11,500 BP. The post-glacial spread of montane forest reflects the advent of moister conditions, but gives no indications of the abrupt climatic oscillations recorded by the lakes.

7) Simple water-balance computations indicate that the Ziway-Shala Basin has been subject to variations in precipitation of the order of $\pm$ 25% over the last 20,000 years. It is not certain whether the Late Pleistocene highstands Ziway-Shala II and III were due to increased precipitation or to decreased temperatures or both.

8) The history of Lake Ziway-Shala has been very similar to the other lakes in the Horn of Africa and to a lesser extent, in the Southern Sahara. Its great sensitivity to climatic fluctuations reflects the large proportion of the catchment lying above 3000 m (17%), the dominance of surface runoff and evaporation in its water balance, and the trough-like topography of the Rift floor.

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APPENDIX 1

UTM grid references of sections and other localities cited in the text

Section/ locality no.	Name of area (if any)	No. Profiles measured (if > 1)	Grid ref.
0838-D1			DV 704086
D2			DV 750068
D3			DV 778036
D4	Meki Town		DV 804006
D5	Kenyi Gona		DU 701909
D6	Korke Adi		DU 761995
D7	Rama Gebriel		DU 722938
D8			DV 748070
D9			DV 786029
D10			DV 786030
D11			DV 734075
D12	Rama Gebriel		DU 727935
D13	E.H.A. Quarry		DV 842053
0738-A1	Gorgeza		DU 447436
0738-B1	Werja		DU 638773
B2	Boloko		DU 690672
B3	Kurkura		DU 613556
B7	Horocallo		DU 638494
B8	Bulbula Village		DU 609534
B14	Mudi		DU 485346
B15			DU 616591
B16	Kertefa		DU 702657
B17	"		DU 691647
B18	"		DU 687647
B19	"	3	DU 684645
B20/10C	Deka Wede	14	DU 659611
B21			DU 770534
B22			DU 778539
B24	Mt. Ilala		DU 776530
B26	Gerbi		DU 707731
B27	Kurkura		DU 627589
B28	Abelosa		DU 724685

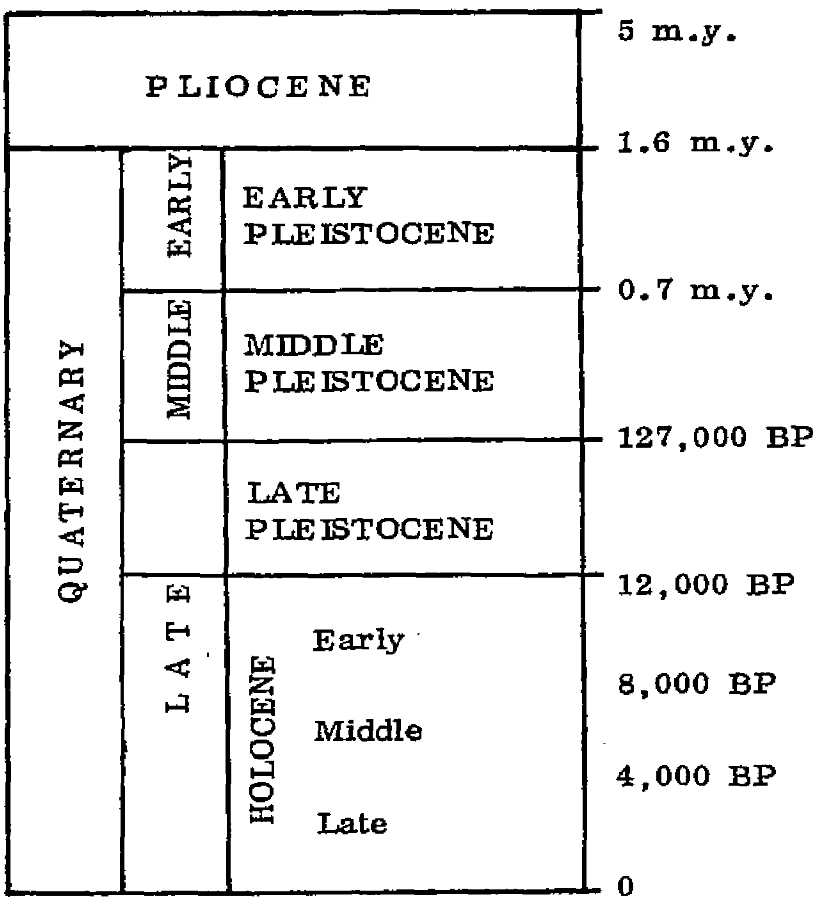
Section/ locality no.	Name of area (if any)	No. of Profiles measured (if >1)	Grid ref.
0738-B29	Awariftu		DU 774667
B30			DU 776497
B31			DU 802504
0738-C1	Rukesa	13	DU 406192
0738-D1	Mirrga graben		DU 642273
D2	Ajewa	96	DU 610265
D2-3			DU 596253
D3			DU 595248
D4	Aneno		DU 610254
D5	Roricha		DU 598240
D6			DU 596260
D9			
D11	Deka Halela	10	DU 588288
D12	" "	2	DU 583287
D13	Rebo	8	DU 606270
D17			DU 594288
0739-A3	Mt. Badda core (Bog 3)		EU 449720
Abiyata core			DU 562507

APPENDIX 2a

Subdivisions of the Quaternary

The time scale used in this study is based on the recommendations by Berggren and Van Couvering (1974), Butzer and Isaac (1975) and Haq et al. (1977). Since there is still no internationally agreed stratotype for the Pleistocene-Holocene boundary (Mercer, 1972; Hopkins, 1975), the base of the Holocene in Ethiopia is taken to be 12,000 BP, for convenience, following Gasse (1975, annexe VB).

The time scale adopted is given below:



Radiocarbon date list for the Ziway-Shala Basin and the Balé Mountains

All dates are based on the 5570-year half-life and are uncorrected for zero error or isotopic fractionation.
All SUA shell dates were measured on the inner 85-90% fraction.
* indicates small sample diluted with dead carbon.

Age (^{14}C yr BP)	Lab. No.	Locality	Section/ Profile No.	Lat.	Long.	Ref.	Material and Significance
28,200 ± 1000	GIF-4010	N.Bulbula Plain	0738-B26	7°54'N	38°44'E	1	Lacustrine shells (Corbicula). Inner portion of very robust valves from fine sand. Locality probably identical to SMU-61. Elevation: 1638 m (80 m.S.D.).
27,050 ± 1540*	SUA-588	Deka Wede V.	0738-B20/11	7°48'N	38°41'E	1	Dispersed charcoal at 1606 m (48 m.S.D.) in a buried soil. A horizon developed in fluvial gravel and immediately overlain by the basal marl of member Deka Wede 2a of the Bulbula formation. Associated with obsidian implements and bone refuse. Maximum age for the Ziway-Shala IIIA transgression.
26,780 ± 440	SMU-61	N.Bulbula Plain	0738-B26 (approx.)	7°53' 24"N	38°43' 48"E	4	Lacustrine shells (Corbicula), 2-4 m above Bulbula R. See GIF-4010.
24,000 ± 750	GIF-3988	Deka Wede V.	0738-B20/10C	7°48'N	38°41'N	1	Lacustrine marl (carbonate) from the base of member Deka Wede 2a of the Bulbula formation. Appears rather young by comparison with date from underlying soil (SUA-588). Elevation 1601 m (43 m.S.D.).

Age (^{14}C yr BP)	Lab. No.	Locality	Section/ Profile No.	Lat.	Long.	Ref.	Material and Significance
22,050 $\pm$ 650	GIF-3987	Deka Wede V.	0738-B20/10C	7°48'N	38°41'E	1	Lacustrine silt (carbonate) from the top of member Deka Wede 2b of the Bulbula formation. Elevation 1607 m (49 m.S.D.). Some sparry calcite present in voids.
14,400 $\pm$ 750	GaK-3748	N.E. Shala	0738-D2/67	7°28'N	38°39'E	2	Algal limestone (inorganic carbonate) from unit UM1 of the Ajewa formation, at 1630 m (72 m.S.D.). Date may be too young, due to contamination by younger carbon.
12,200 $\pm$ 200	GIF-3985	Deka Wede V.	0738-B20/10C	7°48'N	38°41'E	1	Lacustrine marl (carbonate) from member Deka Wede 3a of the Bulbula formation. Elevation 1615 m (57 m.S.D.). Too old by comparison with SUA-494.
11,870 $\pm$ 300*	SUA-494	Deka Wede V.	0738-B20/10C	7°48'N	38°41'E	1	Dispersed charcoal at 1615 m (57 m.S.D.) in a buried soil A horizon. Associated with LSA implements and bone refuse. Immediately underlies member Deka Wede 3a of the Bulbula formation. Maximum age for Ziway-Shala IV transgression. See GIF-3986.
11,800 $\pm$ 150	SUA-504	Deka Wede V.	0738-B20/12	7°48'N	38°41'E	1	Riverine shells (large, articulated valves of *Caelatura aegyptiaca* in growth position) from basal fill of deep channel incised into the Abernosa pumice member of the Bulbula formation, at 1602 m (44 m.S.D.). Cross-stratified sands containing shells are overlain by member Deka Wede 3.

Age (^{14}C yr BP)	Lab. No.	Locality	Section/ Profile No.	Lat.	Long.	Ref.	Material and Significance
11,510 ± 110	SMU-72	Wayso (Waso)	Locality 19	7°48'36" N	38°37'E	4	Charcoal from 80 cm below surface in red soil developed in pumiceous lapilli deposit on hillside, 9.6 km SW of Adamitulu. Dates episode of ash-fall and provides maximum age for soil development on ash.
11,500 ± 200	GIF-3891	Mt. Badda (Core from Bog 3)	0739-A3	7°52'N	39°22'E	1	Well decomposed mud peat. Basal date, 290-300 cm. Elevation of site: 4000 m.
10,450 ± 180	GIF-3986	Deka Wede V.	0738-B20/10C	7°48'N	38°41'E	1	Total organic matter from A horizon of buried soil (<2 mm fraction) immediately underlying member Deka Wede 3a of the Bulbula formation at 1615 m (57 m.S.D.). Too young by comparison with SUA-494.
10,370 ± 180	GIF-3969	Abiyata Core		7°42'N	38°36'E	1	Organic fraction of lake sediments, 8.25 - 8.30 m, 8.40 - 8.50 m and 8.55 m (amalgamated). Approximate elevation: 1577 m (19 m.S.D.). $\delta^{13}C$ not measured.
10,333 ± 90	SMU-86	Wayso (Waso)		Not publ.	Not publ.	5	Charcoal associated with artifacts in a reddish brown palaeosol developed in volcanic ash. See SMU-72 for comment.
10,220 ± 140	SUA-500	NE. Shala	0738-D2/64	7°28'N	38°39'E	1	Large lens of Acacia charcoal at 1629 m (71 m.S.D.), 25 cm below the top of unit UM3 of the Ajewa formation. Dates regression between lacustral phases Ziway-Shala V and VI.

Age (^{14}C yr BP)	Lab. No.	Locality	Section/ Profile No.	Lat.	Long.	Ref.	Material and Significance
10,010 ± 150	SUA-502	NE. Shala	0738-D2/71	7°28'N	38°39'E	1	Lacustrine shells (Melanoides tuberculata) from pebbly silts and sands in unit UM4 of the Ajewa formation at 1640 m (82 m.S.D.).
9950 ± 170	GIF-3968	Abiyata Core		7°42'N	38°36'E	1	Total organic matter, buried soil, 5.40 m. Approximate elevation 1580 m (22 m.S.D.). $\delta^{13}C$ -12.5 ‰
9810 ± 170	GIF-3966	Abiyata Core		7°42'N	38°36'E	1	Organic fraction of lake mud, 4.00 m. Approximate elevation 1581 m (23 m.S.D.). $\delta^{13}C$ -23.3 ‰ Stratigraphically inconsistent with GIF-3967.
9550 ± 170	GIF-3967	Abiyata Core		7°42'N	38°36'E	1	Organic fraction of lake mud, 4.90 m. Approximate elevation 1580 m (22 m.S.D.). $\delta^{13}C$ -18.2 ‰ Stratigraphically inconsistent with GIF-3966.
9480 ± 140	SUA-501	NE. Shala	0738-D2/66	7°28'N	38°39'E	1	Lacustrine shells (Melanoides tuberculata) from prominent shell bed near base of unit UM4 of the Ajewa formation, at 1633 m (75 m.S.D.).
9360 ± 210	Not publ.	C. Bulbula Plain (Kertefa)	0738-B17 (approx.)	7°51'N	38°44'E	3	Lacustrine shells, 60 cm from base of upper lacustrine beds.
9330 ± 100	SMU-64	W. Ziway (Terrace IV)	S. of Abosa	8°01'18" N	38°43'12" E	4	Lacustrine shells (Melanoides) from 2 m depth. Estimated elevation (from L.R.D. levelling): 1659 m (101 m.S.D.). Same sample as SMU-63.

Age (^{14}C yr BP)	Lab. No.	Locality	Section/ Profile No.	Lat.	Long.	Ref.	Material and Significance
9270 $\pm$ 85	Q-1358	N.Bulbula Plain	0738-B2	7°51'N	38°43'E	1	Mixed shell from lower shell bed, in lacustrine deposits at 1633 m (75 m.S.D.).
9220 $\pm$ 190	GaK-3386	NE. Shala	0738-D2/ 67-71 (est)	7°28'N	38°39'E	2	Lacustrine shells (Melanoides tuberculata) from prominent shell bed near base of unit UM4 of the Ajewa formation at 1642 m (84 m.S.D.). Locality given in Grove et al. (1975) is incorrect according to unpublished field report by Grove and Goudie.
9090 $\pm$ 100	SMU-63	W. Ziway (Terrace IV)	S. of Abosa	8°01'18" N	38°43'12" E	4	Lacustrine shells (Corbicula) from 2 m depth. Estimated elevation (from L.R.D. levelling): 1659 m (101 m.S.D.). Same sample as SMU-64.
8520 $\pm$ 200 *	GIF-3984	Deka Wede V	0738-B20/10C	7°48'N	38°41'E	1	Total organic matter from A horizon of buried soil (< 2 mm fraction) between members Deka Wede 3 and 4 of the Bulbula formation. Elevation: 1620 m (62 m.S.D.). Records the Ziway-Shala V/VI regression.
8320 $\pm$ 175	Q-1114	NE. Shala	0738-D2/74	7°28'N	38°39'N	2	Lacustrine shells (Melanoides tuberculata) from sands and gravels in unit UM4 of the Ajewa formation. Elevation: 1646 m (88 m.S.D.).

Age (^{14}C yr BP)	Lab. No.	Locality	Section/ Profile No.	Lat.	Long.	Ref.	Material and Significance
7970 ± 150	GIF-3965	Abiyata Core		7°42'N	38°36'E	1	Organic fraction of lake mud, 2.60 m. Approximate elevation: 1582 m (24 m S.D.). $\delta^{13}C$ -19.4‰
7920 ± 80	Q-1362	Balé Mts. (Danka V. core)	0739-D1	6°58'N	38°47'E	1	Organic fraction, silty and gravelly beds at base of core, 230-250 cm.
7880 ± 290	SMU-66	W. Ziway (Terrace III)	1.7 km NE of Abosa	8°2'6"N	38°44'6"E	4	Lacustrine shells (*Corbicula*) from depth of 0.5-1.0 m. Estimated altitude (from L.R.D. levelling) 1653 m (95 m S.D.). Same sample as SMU-68, 69. Authors suggest that *Corbicula* yields dates 2500 y too old but contamination by older reworked shell seems more likely.
7490 ± 140	GIF-4108	Mt. Badda (core from Bog 3)	0739-A3	7°52'N	39°22'E	1	Well decomposed mud peat, 232-250 cm. Elevation of site: 4000 m.
6500 ± 120	Q-1101	Mirrga graben (SW. Langano)	near 0738-D1	7°29'N	38°41'E	2	Lacustrine gastropod shells (probably *Melanoides tuberculata*) from beach deposits at 1651 m (93 m S.D.).
6150 ± 150 *	SUA-497	NE. Shala	0738-D2/54	7°28'N	38°39'E	1	Lacustrine shells (*Melanoides tuberculata*) from prominent shell bed near base of unit UM4 of the Ajewa formation, at 1617 m (59 m S.D.).

Age (^{14}C yr BP)	Lab. No.	Locality	Section/ Profile No.	Lat.	Long.	Ref.	Material and Significance
6060 ± 120	SUA-499	Deka Wede V.	0738-B20/10A	7°48'N	38°41'E	1	Lacustrine shells (Bellamya unicolor) associated with rolled fish bones in pebbly grit, probably a beach deposit, at 1643 m (85 m.S.D.). Dates member Deka Wede 4 of the Bulbula formation.
5940 ± 70	Q-1359	N. Bulbula Plain	0738-B2	7°51'N	38°43'E	1	Mixed shell from upper shell bed, in well-bedded lacustrine silts at 1640 m (82 m.S.D.).
5790 ± 115	SUA-503	Deka Wede V.	0738-B20/10 C	7°48'N	38°41'E	1	Lacustrine shells (Bellamya unicolor) from middle of three rich shell horizons in member Deka Wede 4 at 1622 m (64 m.S.D.). Shells and fish bones occur in pumice gravel interbedded with silt and waterlaid ash.
5750 ± 70	Q-1361	NE. Shala	0738-D2/11	7°28'N	38°38'E	1	Fishbones (collagen) from Roricha fish bed at 1567 m (9 m.S.D.). Base of unit D3 in the Ajewa formation. See Q-1369 and Q-1557.
5680 ± 55	Q-1557	NE. Shala	0738-D2-3	7°28'N	38°38'E	1	Mixed shell from Roricha fish bed, estimated altitude 1560 m (2 m.S.D.). Base of unit D3 in the Ajewa formation (see Q-1361, 1369).
5610 ± 100	GaK-3387	NE. Shala	0738-D2/55	7°28'N	38°39'E	2	Lacustrine shells (Melanoides tuberculata) from prominent shell bed near base of unit UM4 of the Ajewa formation at 1619 m (61 m.S.D.).

Age (^{14}C yr BP)	Lab. No.	Locality	Section/ Profile No.	Lat.	Long.	Ref.	Material and Significance
5570 ± 90	SMU-68	W. Ziway (Terrace III)	1.7 km NE. of Abosa	$8^{\circ}2'6''$N	$38^{\circ}44'6''$E	4	Lacustrine shells (Melanoides) from depth of 0.5-1.0 m. Estimated altitude (from L.R.D. levelling) 1653 m (95 m.S.D.). Same sample as SMU-66, 69.
5550 ± 120	GIF-4013	C. Bulbula Plain (Kurkura)	0738-B3	$7^{\circ}45'$N	$38^{\circ}39'$E	1	Lacustrine shells (Bulinus "truncatus") from bedded gravels in member Kurkura 6 of the Bulbula formation. Same sample as GIF-4012, 4014. Elevation: 1613 m (55 m.S.D.).
5530 ± 65	Q-1369	NE. Shala	0738-D2/11	$7^{\circ}28'$N	$38^{\circ}38'$E	1	Mixed shell from the Roricha fish bed at 1567 m (9 m.S.D.). Base of unit D3 in the Ajewa formation. See Q-1361 and Q-1557.
5450 ± 120	GIF-4012	C. Bulbula Plain (Kurkura)	0738-B3	$7^{\circ}45'$N	$38^{\circ}39'$E	1	Lacustrine shells (Bellamya unicolor) from bedded gravels in member Kurkura 6 of the Bulbula formation. Same sample as GIF-4013, 4014. Elevation: 1613 m (55 m. S.D.).
5370 ± 60	SMU-69	W. Ziway (Terrace III)	1.7 km NE of Abosa	$8^{\circ}2'6''$N	$38^{\circ}43'24''$E	4	Lacustrine shells (Bulinus) from depth of 0.5-1.0 m. Estimated altitude (from L.R.D. levelling) 1653 m (95 m.S.D.). Same sample as SMU-66, 68.
5300 ± 120	GIF-4014	C. Bulbula Plain (Kurkura)	0738-B3	$7^{\circ}45'$N	$38^{\circ}39'$E	1	Lacustrine shells (Melanoides tuberculata) from bedded gravels in member Kurkura 6 of the Bulbula formation. Same sample as GIF-4012, 4013. Elevation: 1613 m (55 m. S.D.).

Age (^{14}C yr BP)	Lab. No.	Locality	Section/ Profile No.	Lat.	Long.	Ref.	Material and Significance
5180 ± 55	Q-1360	Mirrga graben (SW. Langano)	0738-D1	$7^{\circ}29'$N	$38^{\circ}41'$E	1	Mixed shell from beach deposits at 1668.5 m (110.5 m.S.D.). Dates overflow level from Ziway-Shala Basin.
5050 ± 100	SMU-67	W. Ziway (Terrace IV?)	3.25 km W of Ziway Town	$7^{\circ}56'30''$ N	$38^{\circ}40'$ $30''$E	4	Lacustrine shells (Bulinus) from 10-30 cm depth. Estimated elevation (from L.R.D. levelling): 1664 m (106 m.S.D.). Same sample as SMU-65 (average 4930 ± 100).
4960 ± 140	Not publ.	C. Bulbula Plain (Kertefa)	0738-B18 (approx.)	$7^{\circ}51'$N	$38^{\circ}44'$E	3	Lacustrine shells from "terrace" cut into diatomaceous sediments. Elevation 1645 m (87 m.S.D.).
4800 ± 70	SMU-65	W. Ziway (Terrace IV?)	3.25 km W of Ziway Town	$7^{\circ}56'30''$ N	$38^{\circ}40'$ $30''$E	4	Lacustrine shells (Bulinus) from 10-30 cm depth. Estimated elevation (from L.R.D. levelling): 1664 m (106 m.S.D.). Same sample as SMU-67 (average 4930 ± 100).
3430 ± 120 *	GIF-4109	Mt. Badda (Core from Bog 3)	0739-A3	$7^{\circ}52'$N	$39^{\circ}22'$E	1	Well decomposed mud peat, 183-200 cm. Elevation of site: 4000 m.
2520 ± 100	GIF-3890	Mt. Badda (Core from Bog 3)	0739-A3	$7^{\circ}52'$N	$39^{\circ}22'$E	1	Well decomposed mud peat, 140-150 cm. Elevation of site: 4000 m.
2510 ± 100	GIF-4011	NE. Shala	0738-D2/39B	$7^{\circ}28'$N	$38^{\circ}38'$E	1	Large lens of Acacia charcoal from colluvial-alluvial deposits in unit DM5 of the Ajewa formation, at 1600 m (42 m.S.D.) Maximum age for Ziway-Shala VII transgression.

Age (^{14}C yr BP)	Lab. No.	Locality	Section/ Profile No.	Lat.	Long.	Ref.	Material and Significance
1540 $\pm$ 60	SMU-84	Macho (Site I)	9.3 km WSW of Adamitulu	$7^{\circ}50'$N	$38^{\circ}36'18''$E	4	Burnt tree stump in yellowish-brown soil in small erosional escarpment. Dates lapilli fall that burned trees.
770 $\pm$ 110	GIF-3889	Mt. Badda (core from Bog 3)	0739-A3	$7^{\circ}52'$N	$39^{\circ}22'$E	1	Well decomposed mud peat, 40-50 cm. Elevation of site: 4000 m.
230 $\pm$ 50	SMU-83	Macho (site II)	9.3 km WSW of Adamitulu	$7^{\circ}50'$N	$38^{\circ}36'18''$E	4	Charcoal from fire pit dug through thin pumiceous lapilli deposit overlying yellowish-brown soil on underlying ash. Minimum age for latest ash fall in Macho area.
149.7 $\pm$ 0.5% modern	SMU-60	L. Ziway (Ziway Town)		$7^{\circ}56'18''$N	$38^{\circ}43'24''$E	4	Living Bulinus "truncatus" shells. Indicates that no correction is necessary for dating fossil shells of this species.

References:

1. This study.
2. Grove et al. (1975).
3. Gèze (1975).
4. Haynes and Haas (1974).
5. Laury and Albritton (1975).

APPENDIX 2c

Calcite/aragonite determinations on dated shell samples

One very important source of error in dating subfossil shells by ^{14}C may arise through recrystallization of the shell material from aragonite to calcite. If any younger carbon is incorporated during recrystallization this will tend to yield measured ages which are systematically too young (Chappell and Polach, 1972).

The calcite content of shell samples submitted for dating was therefore compared as far as possible with modern shells of the same species in order to determine whether any recrystallization had occurred. Calcite percentages were determined by XRD using the method described by Griffin (1971) (Table 4.1).

The results obtained are given in the table below. The only samples giving cause for concern are 0738-D1 and 0738-B3. In both cases the observed calcite peak results from slight encrustation with pedogenic carbonate on the outside of the shells, which were otherwise fresh. This encrustation was carefully removed before dating by the laboratories concerned. It is still possible, however, that these dates are a few hundred years too young, due to contamination by younger carbon.

Calcite contents of modern and subfossil shell material

Age (^{14}C yr BP)	Locality	Species	Peak height ratio	% calcite
Modern	L. Ziway[1]	*Bulinus "truncatus"*	∞	0
"	"[1]	*Melanoides tuberculata*	∞	0
"	L. Tana at Gorgora[1]	*Bellamya unicolor*	11.0	1-2
"	L. Chad[2]	*Bellamya unicolor*	∞	0
"	L. Chad[2]	*Bulinus "truncatus"*	50	Trace (<0.5)
"	L. Chad[2]	*Melanoides tuberculata*	50	Trace (<0.5)

Age (^{14}C yr BP)	Locality	Species	Peak height ratio	% calcite
8320 ± 175	0738-D2/74	Melanoides tuberculata	28.2[a]	Trace (<0.5)
			12.8[b]	1
10,010 ± 150	0738-D2/71	Melanoides tuberculata	20.2[a]	Trace (<0.5)
			14.8[b]	1
9480 ± 140	0738-D2/66	Melanoides tuberculata	34.8[a]	Trace (<0.5)
			63.3[b]	Trace (<0.5)
6150 ± 150	0738-D2/54	Melanoides tuberculata	49.9[a]	Trace (<0.5)
			41.6[b]	Trace (<0.5)
5530 ± 65	0738-D2/11	Melanoides tuberculata	35.6[a]	Trace (<0.5)
		Valvata nilotica	31.7[a]	Trace (<0.5)
5680 ± 55	0738-D2-3	Melanoides tuberculata	36.4[a]	Trace (<0.5)
		Bulinus "truncatus"	∞[a]	0
5180 ± 55	0738-D1	Melanoides tuberculata	3.62[a]	8
			1.88[c]	14
		Bulinus "truncatus"	15.9[a]	1
			1.74[c]	15
6060 ± 120	0738-B20/10A	Bellamya unicolor	∞[a]	0
			91.9[b]	Trace (<0.5)
5790 ± 115	0738-B20/10C	Bellamya unicolor	∞[a]	0
			21.9[b]	Trace (<0.5)
11,800 ± 150	0738-B20/12	Caelatura aegyptiaca	38.7[a]	Trace (<0.5)
			∞[b]	0
5300 ± 120	0738-B3	Melanoides tuberculata	9.16[a]	2
			13.4[b]	1
5450 ± 120	"	Bellamya unicolor	46.0[a]	Trace (<0.5)
			11.6[b]	2
5550 ± 120	"	Bulinus "truncatus"	13.0[a]	1
			10.6[c]	2
(9360 ± 210)	0738-B17 (F. Geze)	Melanoides tuberculata	56.4[a]	Trace (<0.5)
			40.1[b]	Trace (<0.5)
(4960 ± 140)	0738-B18 (F. Geze)	Bellamya unicolor	∞[a]	0
			69.7[b]	Trace (<0.5)
		Melanoides tuberculata	∞[a]	0
			33.1[b]	Trace (<0.5)
		Caelatura aegyptiaca	58.5[a]	Trace (<0.5)
			50.8[b]	Trace (<0.5)
28,200 ± 1000	0738-B26	Corbicula sp.	36.5	Trace (<0.5)

KEY:

a	Freshest shells in sample
b	Most weathered shells in sample
c	Shells with carbonate encrustation
1	Sample provided by Dr. D.S. Brown
2	Sample provided by Mr. A.T. Grove

APPENDIX 3

Land snails from unit UM5 of the Ajewa formation (section 0738-D2)

P 80-82

Limicolariopsis ellisi Crowley and Pain
Cerastua sp.
Chlamydarion hians (Pfeiffer)

P 79

Limicolariopsis ellisi Crowley and Pain
Cerastua sp.
Chlamydarion hians (Pfeiffer)
Pupisoma steudneri (Jickeli)
Macroptychia sp. (immature)
Guppya rumrutiensis (Preston)
Thapsia ? cf. *abyssinica* (Jickeli)
Kaliella sp.
Truncatellina aff. *naivashaensis* (Preston)
Truncatellina sp.
Gastrocopta sp.
Pupilla fontana (Krauss)
Streptaxid

(Identifications by Dr. B. Verdcourt, Royal Botanic Gardens, Kew.)

APPENDIX 4a

Salinity and alkalinity classifications

The salinity and alkalinity classes for surface waters which are used in this study are taken from Gasse (1975, annexe IIIA-2) and Richardson et al. (1978). They both refer to $NaHCO_3$-dominated waters in which total dissolved solids (T.D.S.), alkalinity and pH are highly correlated.

Salinity classification

Class	T.D.S. (‰)
Hyperhaline	>35
Polyhaline	16 - 35
Mesohaline	2 - 16
Oligohaline	<2

Diatoms living preferentially in waters of any of these classes are distinguished by the ending -halobous, e.g. oligohalobous.

Alkalinity classification

Class	Alkalinity (meq/l)
Highest	>200
High	100 - 200
Medium-high	50 - 100
Medium	10 - 50
Medium-low	2 - 10
Low	1 - 2
Lowest	<1

APPENDIX 4b: Diatoms identified in lacustrine sediment samples from the Ziway-Shala Basin

Taxa	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
Achnanthes sp.									+								+		
A. clevei												+				+			
A. exigua v. heterovalvata															+	+	+		
A. lanceolata v. rostrata									+								+		
A. minutissima															+				+
Amphora coffaeformis											+								
A. ovalis			+	+		+											+		
var. libyca												+		+					
var. pediculus						+			+			+	+	+		+	+	+	D
A. veneta												+			+				
Anomoeoneis sphaerophora	+															+			
var. guntheri															+				
var. sculpta								+											
Caloneis sp.																			+
Campylodiscus clypeus	+					+		+			+								
Cocconeis sp.		+																	
C. diminuta																			+
C. placentula				+								+	+		+	+	+		+
var. euglypta									+							+			
C. thumensis										+						+	+	+	+
C. glomerata													?	?	?				
C. kutringiana			?	+	+					+									
C. meneghiniana			+			+					+	+			+			+	
C. ocellata		+	?		+	+	+		A	D			+	+		+	+	+	+
C. striata											+								
C. turgida			+			+			+	+									
C. ventricosa			+	+															
Cymatopleura solea				+									+						
Cymbella sp.		+							+										
C. affinis												+	+	+	+			+	+
C. cesatii														?					?
C. microcephala																			?
C. turgida				+									+	+	D	+			+
Epithemia sorex		+	+	+		+			+			+	+	+	D		+	+	+
var. grossestriata																+			
E. turgida										+			+	+			+		+
E. zebra	+	+	+	+			+			+	A			D	D	+	+	+	+
Fragilaria brevistriata				+		+	+		D	+			D	A	+	D	+	D	D
F. construens																+	+		
var. venter											+						+		
F. hungarica																		+	
F. intermedia															+			+	+

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
F. lapponica						+	+		+	+						+	+	+	
F. pinnata										+						+	+	+	+
F. subatomus																	+	+	
Gomphonema sp.		+													+				
G. dubravicense															+				
G. gracile														+					
G. parvulum v. micropus															+				
Gomphonitzschia beccari																	+		
Mastogloia elliptica				+									+	A	+		+	+	+
Melosira sp.									+	+									
M. agassizii		?		D			+	+	+	+							D		+
M. granulata						+	+					+					+	+	
var. angustissima						+				+					+				
f. curvata					+												+		
M. italica			?D	?															
M. nyassensis						+													
var. victoriae					+							[illegible]						+	+
Navicula spp. (indet.)										A						A			+
N. cryptocephala												+							
var. intermedia													+					+	
N. cuspidata													+						
N. minima																	+		
N. oblonga													+	+					
N. perlatoides													+					+	
N. pupula												+			+		+		+
N. radiosa												+	+	+	+				
N. rhynchocephala																			+
N. rotunda																			+
N. scutelloides								+		+									
N. seminuloides									+	+						+	+	?	+
N. seminulum						+													
N. subrotunda													+				+	+	
Neidium iridis			+																
Nitzschia spp. (indet.)	+									+							+	+	+
N. amphibia			+	+								+	+		A	+			+
N. bacata													?			?			
N. confinis													?						
N. frustulum										+	+								
var. perminuta										+	+		+		+	+	+	+	
var. perpusilla						+											+		
N. hungarica																	+		
N. lancettula						+	+			+						+	+	+	+

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
N. nyassensis																?			
N. tropica												+	?	+	+				+
Opephora martyi		+							+										+
Rhoicosphenia curvata												+					+		+
Rhopalodia gibba			+	+		+						+	A	D	+	+	+		+
R. gibberula	+							A			D		+		+	+			
var. debyi								+											
R. hirundiniformis								?											
R. musculus									+	+						+			
R. vermicularis						+	+	?	+	+						+	+	+	+
Stauroneis karstenii																	+		
Stephanodiscus astraea + v.						+	+		D	D			+			+	D	+	
S. dubius																+			
S. minutus														+		+	+		
Surirella ovalis	+																	+	
Synedra acus												+							
S. dorsiventralis																			+
S. rumpens													+		+			+	
S. ulna		+				+						+	+		A		+	+	+
Thalassiosira faurii											+	+	+	+	+		+		
T. rudolfi											+								

Identifications by Dr. Francoise Gasse, Ecole Normale Superieure, Fontenay-aux-Roses.

KEY

+ Present
A Abundant
D Dominant
? Identification uncertain

Sample details

1 0738-B20/11/3. Member 1b, Bulbula fm.
2 0738-B20/10C/40. Member 2a, Bulbula fm.
3 0738-B20/10C/22. Member 3a, Bulbula fm.
4 0738-B20/10C/20. Member 3b, Bulbula fm.
5 0738-D2/44/4.70 m. Unit DM3b, Ajewa fm. (diatomaceous mudstone)
6 0738-D2/67/0.25 m. Unit UM1, Ajewa fm.
7 0738-D2/67/0.70 m. Unit UM1, Ajewa fm.
8 0738-D5. Silts underlying Roricha fish bed
9 0738-C1/3. Diatomaceous mudstones (Holocene?)
10 0738-C1/3. Diatomaceous mudstones (Holocene?)
11 0838-D7. Lower Late Pleistocene ? diatomite
12 0738-D2/44. Unit DM3b, Ajewa fm. (fluffy white diatomite)
13 0738-D2/83/6.30 m. Unit U2b, Ajewa fm.
14 0738-D2/83/6.70 m. Unit U2b, Ajewa fm.
15 0738-B14. Fluffy diatomite
16 0738-B17. Kertefa diatomaceous silts
17 0738-B28. Kertefa-type diatomite
18 0838-D9. Kertefa-type diatomaceous silts
19 0838-D9. Kertefa-type diatomaceous silts

APPENDIX 5

Fish and crab remains from lacustrine sediments in the Ziway-Shala Basin

Ajewa section (0738-D2)

Roricha fish bed (unit D3)

P11	5 Tilapia cf. nilotica* 5 Tilapia sp.* 2 Cichlids, probably Tilapia* 6 Barbus cf. intermedius (large)* 2 Barbus sp. (small)* Barbus sp. Spines of Barbus cf. callensis (serrated) Synodontis sp. Otolith of ? Lates	(dated 5750 ± 70 BP and 5530 ± 65 BP)
P15-16	Cichlid*	

Unit D4

P14	Tilapia	
P15	Cichlid ?cichlid	

Unit UM4

P52-4	Probably cyprinid	
P54	Tilapia sp.	(dated 6150 ± 150 BP)
P66	Barbus cf. intermedius Barbus spines Cichlid spines	(dated 9480 ± 140 BP)
P74	Cichlid spine	(dated 8320 ± 175 BP)

Roricha fish bed further south (unit D3)

0738-D2-3	Cyprinid vertebrae	(dated 5680 ± 55 BP)
0738-D5	2 Tilapia sp.* 1 cichlid, probably Tilapia* 2 cichlids* Cichlid	

Deka Halela area (mid-Holocene?)

0738-D11	*Barbus* sp. (large)	
	Barbus sp. (small)	
	Large *Barbus* spine (serrated)	
	Tilapia sp.	
	Claw fragments of crab (not *Potamonautes antheus*)	
0738-D12	Probably cyprinid	

Rukesa section (0738-C1)

P7	Barbus spine (serrated) (probably reworked from mid-Holocene? shell beds upstream)	

Horocallo section (0738-B7)

	Tilapia sp.	

Kertefa area

0738-B17	*Barbus* spine	(dated 9360 $\pm$ 210 BP)
0738-B18	Cyprinid	(dated 4960 $\pm$ 140 BP)

Deka Wede section (0738-B20)

P10A	Indet.	(dated 6060 $\pm$ 120 BP)
P9	Indet.	

KEY: * Complete, articulated skeleton

(Identifications by Dr. K.E. Banister, British Museum (Natural History) and Mr. T.R. Williams, Dept. Zoology, University of Liverpool.)

APPENDIX 6a

Fossil bones from the palaeosol at 13.7 - 13.9 m in the type section of the Bulbula formation (0738-B20/10C)

1) Human (*Homo sapiens sapiens*)
   - cranial fragment (right parietal of adult) (Plate 56)

2) Baboon (*Papio* sp.)
   - right femur (Plate 56)
   - distal end left femur (Plate 56)

3) Monkey
   - ulna
   - incisor tooth

4) Medium-sized bovid

- 1 horn-core fragment - 1 upper premolar fragment - 2 lower jaw fragments	skull bones
- 3 fragments of phalanges - 3 fragments of metacarpus - 3 fragments of metatarsus - 1 distal sesamoid	foot bones

5) Larger bovid
   - 1 upper molar fragment

6) Other unidentified fragments

(Identifications by Dr. D. Geraads, Dept. of Geology, University of Addis Ababa, and R.L. Ciochon, Dept. Anthropology, University of California at Berkeley.)

APPENDIX 6b

Artifacts from soil profiles and nonlacustrine deposits in the Ziway-Shala Basin

0838-D1 - eroded slopes, above calcrete

MSA parti-bifacial point
MSA point
Broken Levallois point
Flake with multifacetted butt (age uncertain)
2 end scrapers (" ")
Side scraper (" ")
Convex single-sided scraper (" ")
Double-sided scraper (" ")
LSA blade
LSA point (backed blade)

0838-D7 - eroded slopes

Flake from prepared core

- surface

Flake
Chipped point
LSA point

0838-D8 - A horizon

LSA flakes

- slopes just below Bt horizon

LSA side/end scraper
LSA point

- gulley floors

9 scrapers
1 bladelet core
1 notched piece
14 blade fragments
7 flakes
1 chunk

- surface

LSA flakes
LSA scraper
Fragments

0838-D9 - surface

Scraper
Retouched blade fragment
Flakes
Fragments

0838-D10 - emerging from pebble gravel

LSA flake

- surface

LSA side/end scraper

0738-B1 - Cck1 horizon

LSA end scraper

- surface

3 LSA microblade cores
LSA end scraper
LSA distal end broken point
LSA ? end scraper

0738-B2 - eroded slopes (probably from colluvium dated between 9270 and 5940 BP

LSA flake
LSA flake fragment

0738-B20 - P11, in situ in soil beneath member Deka Wede 2a, dated 27,050 ± 1540 BP

Core trimming flake
Blade
(Remainder of collection lost during the Revolution)

- P13, in situ in channel sands at base of member 2a

LSA blade
Distal blade fragment

- P10C, in situ in soil beneath unit 3a, dated 11,870 ± 300 BP (charcoal) and 10,450 ± 180 BP (soil humus)

(associated with cranial fragment and baboon long bones (Plate 56)

32 LSA blade and flake fragments
5 LSA flakes
9 LSA blades
3 LSA chunks

(further along section)

Levallois point
LSA blade

3 blade fragments
LSA crescent (backed blade)
2 end scrapers
Scraper
Burin
(Large number of LSA flakes, blades and fragments lost during the Revolution)

0738-B24 - emerging from weathered (plateau" gravels

Large number of evolved Acheulian bifaces (obsidian) (lost during the Revolution) (Plate 4)

0738-D2 - P38-39, in situ, in unit DM4

Rolled MSA flake
Multifacetted LSA core
Flake fragment

- P39, in situ, in unit DM4

Backed piece (blade?)
Patinated flake

Artifacts not marked "lost during the Revolution" were deposited with the Ethiopian Ministry of Culture.

(Identifications by Professor J.D. Clark, H. Kurashina and S. Brandt, Dept. of Anthropology, University of California at Berkeley, and Dr. J. Chavaillon, C.N.R.S.)

APPENDIX 7

Selected sediment analyses from the Ziway-Shala Basin

a) Carbonate-free grain-size and carbonate analyses

Sample no.	Location etc.	Mean $M\phi$	Median $Md\phi$	Sorting $\sigma\phi$	Sorting Class	Skewness	Kurtosis	% Carbonate
Modern beach-ridge (B) and mudflat (M) samples								
M1(B)	NE. Shala	-0.319	-0.386	0.791	m.s.	1.25	10.4	n.a.
M2(B)	NE. Abiyata	-0.630	-0.185	1.74	p.s.	-0.312	2.58	n.a.
M3(B)	SW. Abiyata	2.04	2.07	0.602	m.w.s.	0.017	2.02	n.a.
M4(B)	N. Langano (O-itu)	-2.04	-2.10	0.808	m.s.	1.03	8.75	n.a.
M6(B)	SW. Langano	-0.063	0.212	1.07	p.s.	-1.78	6.84	n.a.
M7(B)	N. Langano (O-itu)	0.338	0.385	0.768	m.s.	0.065	4.67	n.a.
M8(B)	" " "	-1.80	-1.93	1.07	p.s.	1.03	4.34	n.a.
M9(B)	NE. Shala	-1.04	-1.09	0.694	m.w.s.	0.310	2.99	n.a.
M10(B)	N. Langano (O-itu)	0.697	0.630	0.579	m.w.s.	1.58	11.0	n.a.
M11(B)	SW. Langano	-2.46	-2.83	1.38	p.s.	0.878	2.79	n.a.
M12(B)	NE. Abiyata	0.053	0.003	1.29	p.s.	0.857	5.27	n.a.
M13(B)	SW. Ziway	0.675	0.522	1.20	p.s.	1.04	5.38	n.a.
M14(B)	SW. Ziway	0.031	-0.095	1.03	p.s.	1.25	7.20	n.a.
M21(M)	N. Abiyata	7.11	6.42	2.59	v.p.s.	0.508	2.06	10.0
M22(M)	N. Abiyata	7.16	6.19	2.76	v.p.s.	0.450	1.86	9.3
M23(M)	SW. Abiyata	4.34	3.70	2.55	v.p.s.	1.41	4.60	3.6
Modern fluviolacustrine samples (Bulbula delta)								
M5	Channel deposits	-0.617	-0.693	1.45	p.s.	0.484	4.54	n.a.
M15	" "	1.82	2.02	1.10	p.s.	-1.33	6.63	n.a.
M27	Abandoned channel fill	4.91	4.06	2.39	v.p.s.	1.38	4.16	

Sample no.	Location etc.		Mean $M\phi$	Median $Md\phi$	Sorting $\sigma\phi$	Sorting Class	Skewness	Kurtosis	% carbonate
	Modern gulley-floor samples (point-bar deposits)								
M16	0738-B19		-0.599	-1.05	1.78	p.s.	1.08	3.79	n.a.
M17	0738-B19		-1.41	-2.08	2.75	v.p.s.	1.61	5.24	n.a.
M18	0738-B19		-0.528	-0.513	1.83	p.s.	0.131	3.09	n.a.
M19	0738-B29		0.258	0.355	1.90	p.s.	-0.472	3.42	n.a.
	Modern bedload sample, Meki River								
M20	Nr. spillway		0.916	0.904	0.944	m.s.	-0.093	3.05	na
	Subrecent Gilbert-type delta samples, Ajewa section (unit D6)								
1.57 m	0738-D2/3	L(S)(foresets)	2.78	2.72	0.687	m.w.s.	0.135	4.24	5-10
1.90 m	0738-D2/3	L(S)(")	2.37	2.37	0.901	m.s.	0.180	2.94	5-10
0.65 m	0738-D2/3	L(S) (prodelta)	4.84	4.71	3.27	v.p.s.	0.245	2.41	5-10

Sample No.	Height (m)	Inferred origin	Mean $M\phi$	Median $Md\phi$	Sorting $\sigma\phi$	Sorting Class	Skewness	Kurtosis	% carbonate
	Section 0738-B20/10C								
1	22.3-24.3	S(C)	1.52	1.04	2.74	v.p.s.	0.621	2.87	1
2	22.0-22.3	L(D)	7.42	7.41	2.18	v.p.s.	0.035	2.31	14
3	21.7-22.0	L(S)	4.86	3.92	2.31	v.p.s.	0.888	3.73	8
4	21.55-21.7	L(S)	6.32	6.18	1.06	p.s.	0.423	2.66	1
5	21.4-21.55	L(G)	-0.728	-0.933	1.77	p.s.	1.30	5.08	1
6	21.25-21.4	A(L)	6.04	5.95	1.86	p.s.	-0.593	7.94	0
8	20.9-21.05	L(S)	5.30	5.60	1.70	p.s.	-2.14	9.20	0
9	20.65-20.9	L(G)	-0.174	-0.254	1.28	p.s.	0.435	4.27	0
11	20.3-20.6	L(G)	-0.889	-0.768	1.21	p.s.	0.673	6.99	0
12	19.9-20.3	L(G)	1.04	0.996	1.30	p.s.	0.565	4.35	0
13	19.6-19.9	L(D)	6.55	6.54	1.95	p.s.	0.020	3.15	5
14	19.1-19.6	P(A)	-1.38	-1.46	1.50	p.s.	0.889	5.19	0

Sample no.	Height (m)	Inferred origin	Mean MØ	Median MdØ	Sorting σØ	Sorting Class	Skewness	Kurtosis	% carbonate
15	18-7-19.1	S(P)	-0.080	-0.352	1.77	p.s.	0.675	3.06	0
16	18.2-18.7	S(P)	0.673	0.539	1.77	p.s.	-0.008	2.84	0
17	17.7-18.2	L(D)	6.65	6.80	2.21	v.p.s.	0.223	2.51	11
18	16.15-17.7	P(A)	-2.18	-2.16	1.98	p.s.	0.137	2.63	n.a.
19	15.55-16.15	S(F)	1.03	0.917	1.86	p.s.	-0.312	3.22	0
20	14.45-15.55	L(D)	6.44	6.68	2.37	v.p.s.	-1.06	5.81	49
21	14.2-14.45	F	-0.861	-1.56	2.73	v.p.s.	1.85	6.23	12
22	13.9-14.2	L(D)	7.21	7.45	2.59	v.p.s.	-1.33	6.33	79
23	13.7-13.9	S(C)	0.961	0.593	2.86	v.p.s.	0.641	3.04	0
24	13.1-13.7	S/C	1.08	0.758	2.74	v.p.s.	0.361	2.62	0
25	12.85-13.1	P(A)	-1.90	-2.27	1.93	p.s.	1.31	4.65	1
26	12.1-12.85	P(W)	-0.545	-0.867	1.85	p.s.	0.831	3.35	1
27	11.4-12.1	P(W)	-0.331	-0.911	2.19	v.p.s.	0.481	2.04	2
28	10.7-11.4	P(W)	0.718	0.882	2.42	v.p.s.	-0.156	1.79	1
29	10.1-10.7	P(A)	-1.69	-2.39	2.28	v.p.s.	1.02	3.29	1
30	9.1-10.1	P(W)	-0.655	-1.28	3.09	v.p.s.	0.668	2.79	0
31	8.9-9.1	A(A)	2.96	3.13	1.48	p.s.	0.007	3.74	2
32	8.1-8.9	P(A)	-2.15	-2.62	2.28	v.p.s.	1.13	3.98	0
33	7.7-8.1	P(W)	-0.352	-0.667	2.18	v.p.s.	0.557	2.49	0
34	6.45-7.7	P(A)	-2.27	-2.50	1.67	p.s.	1.63	7.29	0
35	5.75-6.45	L(D)	6.76	6.56	2.21	v.p.s.	0.381	2.78	16
36	5.05-5.75	L(S)	4.59	3.91	2.70	v.p.s.	0.240	2.72	3
37	4.05-5.05	L(S)	2.48	2.72	2.92	v.p.s.	0.422	4.53	2
38	3.25-4.05	L(S)	4.32	3.70	2.15	v.p.s.	0.792	3.92	17
39	2.6-3.25	L(D)	8.51	8.09	2.36	v.p.s.	-0.034	1.81	80
40	1.5-2.6	L(D)	7.72	7.54	1.87	p.s.	0.183	2.43	83
41	1.4-1.5	L(S)	4.07	3.55	1.80	p.s.	1.95	6.61	4
42	1.2-1.4	L(S)	5.24	5.04	1.87	p.s.	0.740	2.59	7
43	0.55-1.2	L(D)	6.17	6.22	2.48	v.p.s.	0.397	2.28	15
44	0.3-0.55	L(S)	4.55	3.70	2.12	v.p.s.	1.57	4.87	9
45	0-0.3	L(D)	6.29	6.25	2.22	v.p.s.	0.531	2.87	31

Sample no.	Height (m)	Sediment type	Mean $M\phi$	Median $Md\phi$	Sorting $\sigma\phi$	Sorting Class	Skewness	Kurtosis	% carbonate
Section 0738-B20/11									
1	5.25-5.7	S(F)	0.458	0.294	1.75	p.s.	0.410	2.95	0
2	4.35-5.25	F	-1.07	-1.09	1.49	p.s.	0.493	4.43	0
3	2.4-4.35	L(D)	6.13	6.19	1.97	p.s.	0.551	3.66	8
4	2.0-2.4	L(D)	7.10	7.20	1.88	p.s.	-0.137	3.65	1
5	1.85-2.0	L(S)	2.71	2.79	1.36	p.s.	0.013	3.53	0
6	1.55-1.85	L(S)	3.68	3.82	1.78	p.s.	-0.580	2.77	0
7	0.95-1.55	L(S)	4.25	5.10	1.99	p.s.	-1.363	4.64	0
8	0.6-0.95	L(S)	4.23	4.27	2.11	v.p.s.	-0.212	2.44	0
Section 0738-D2									
P54 (unit UM4)		Matrix of shell bed	2.30	2.66	1.37	p.s.	-0.700	2.33	n.a.
P59 (base UM4)		L(S) ("mud blanket")	5.33	6.09	2.22	v.p.s.	-0.527	3.11	n.a.
P57 (unit UM4)		L(D)	5.52	5.47	1.83	p.s.	0.548	3.55	n.a.
P74 (unit UM4)		L(S)	5.25	5.16	2.48	v.p.s.	0.303	3.94	7
P74 (unit UM4)		L(S) (foresets)	2.31	2.22	0.961	m.s.	0.493	2.79	11
P15 (unit D4)		L(S)	5.16	4.95	2.45	v.p.s.	0.999	3.53	14

KEY:

n.a.	not available
L(G)	Lacustrine gravels (beach or shoreface?)
L(S)	Lacustrine (shallow water?)
L(D)	Lacustrine (deeper water?)
A(L)	Water-laid ash
S	Soil (nature of sediment in brackets)
A(A)	Air-fall ash
P(A)	Air-fall pumice
P(W)	Weathered air-fall pumice?
C	Colluvium
F	Ephemeral - stream deposits

m.w.s.	moderately well sorted
m.s.	moderately sorted
p.s.	poorly sorted
v.p.s.	very poorly sorted

b) Bulk density measurements on lake sediments

Locality	Height (m)	Unit	Age	Sample no.	Bulk density (g/cm^3)
Shallow-water diatomites ("Kertefa" facies)					
i) Purest samples					
0838-D7			Lower Late Pleistocene?	1	0.45
0738-D2/44	3.60	DM3b (base)	Late Pleistocene?	2	0.38
0738-D2/83	6.10	U2b	Holocene?	3	0.39
0738-B14			Holocene?	4	0.30
"			"	5	0.17
ii) More silty variants					
0738-B28			Holocene?	6	0.58
0738-B17			Holocene?	7	0.57
0838-D9			Holocene?	8	0.63
Deep-water diatomites and diatomaceous mudstones ("Kurkurá" facies)					
0738-D2/25		DM1	Lower Late Pleistocene?	9	0.38
0738-D2/44	4.25	DM3b (top)	Late Pleistocene?	10	0.50
"	4.70	" "	" "	11	0.56
0738-B3	13.15	Kurkura 4	Late Pleistocene?	12	0.58
"	7.6	Kurkura 3	Late Pleistocene?	13	0.71
"	7.6	"	" "	14	0.72
0738-C1/3			Holocene?	15	0.64
Deep-water silts					
0738-D2/7	1.55	D3	mid-Holocene	16	0.93
"	2.40	"	" "	17	0.93

APPENDIX 8

Methods used to reconstruct the Late Pleistocene snowline in the Arussi Mountains (after Osmaston, 1975)

a) Höfer's mid-height method

This very simple method is based on the assumption that the firnline lies half-way between the altitudes of the top and bottom of a steady-state glacier. The palaeofirnline (F) is estimated from the equation:

$$F = \frac{T + B}{2}$$

where T is the altitude of the upper end (top) of the glacier, approximated by the summit elevation;

B is the altitude of the lower end (bottom) of the glacier, approximated by the lower limit of its terminal moraines.

This method has the advantage of being simple and of being usable in the absence of area data. It gives most accurate results for valley glaciers: for plateau glaciers it overestimates the area of the accumulation zone.

b) The general height method of Osmaston (1965, 1975)

A better estimate of the firnline can be made by taking some other fraction or "Hofer coefficient" of the height range, based on a least-squares fit to the field data. The relationship used is of the form:

$$F = B + H(T - B)$$

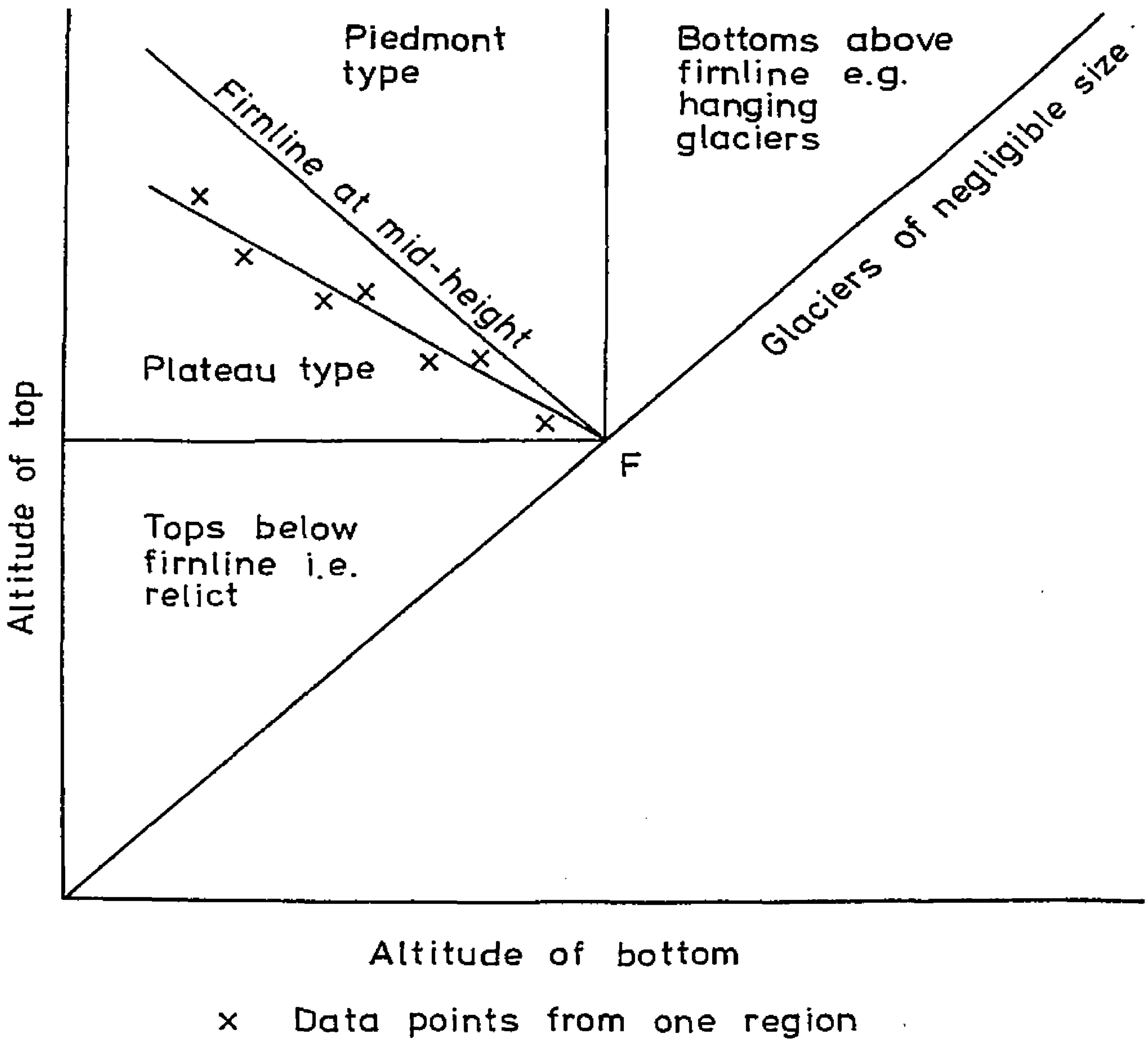
where usually $0.6 > H > 0.4$.

The basic general height graph shown below can be divided into different sectors, corresponding to different types of glaciers. F is the point at which the best-fit line through the data points cuts the diagonal corresponding to glaciers of zero size. The Höfer coefficient, H, is given by the equation

$$H = \frac{1}{1 - b}$$

where b is the slope of the regression line fitted to the data points (N.B. b is negative).

Plateau glaciers characteristically give values of $H > 0.5$, while piedmont glaciers give $H < 0.5$.



APPENDIX 9

Terminology used for lacustrine subenvironments

By analogy with marine coastlines, the lacustrine environment is divided here into three principal depth zones: the foreshore, which is generally characterized by swash-zone processes; the shoreface (or inshore zone) extending lakeward to the fair-weather wave base, which is permanently covered by water and over which beach sands and gravels actively oscillate with changing wave conditions; and the offshore zone to lakeward of the shoreface (Gary et al., 1974; Collinson et al., 1978).

The terms littoral and marginal are employed nonspecifically here for the shallow-water zone above wave base. Similarly, nearshore is a general term applicable to the inshore zone and the inner part of the offshore zone (Gary et al., 1974). It should be noted, however, that many authors use "littoral" to designate the depth zone in lakes in which aquatic macrophytes can grow, e.g. Ruttner (1963). Pelagic and profundal are used here synonymously to refer to the deeper parts of the lake ($\geqslant$10 - 20 m) which are characterized by fine-grained muds and by the absence of aquatic vegetation (Gary et al., 1974).

When referring to diatom assemblages, the term littoral includes both the epiphytic taxa which live on aquatic macrophytes and the benthic taxa living on the bottom in shallow water.

The terminology for lacustrine deltas follows Born (1972). The on-delta environment includes all the subaerial portions of the delta plain. The delta-front platform fringes the shoreline of the delta. The delta slope is analogous to the foresets of classic usage, and the pro-delta environment comprises the offshore areas distant from significant deltaic influence.

Table 8:1
Summary of Late Quaternary environmental changes in the Ziw

Lacustral phases	Estimated ages (^{14}C yr BP)	Lake levels (m.a.s.l.) (not drawn to scale)	Dominant diatom taxa (Abiyata Basin)	Molluscs identified	Ostracods identified	Fish and crabs identified	
ZIWAY — SHALA VIII	pre-1933 AD	1558 (Shala) 1564	?	All reworked	++		
	?	<1559					
ZIWAY — SHALA VII	<2500	ca. 1600	?	All reworked	++	*Tilapia* sp.	N
		<1569					
ZIWAY — SHALA VI	4800 5750BP?	1670 >1651	*Cyclotella ocellata* *Melosira granulata* var. *valida*	*Bellamya unicolor* *Valvata nilotica* *Melanoides tuberculata* *Lymnaea natalensis* *Bulinus "truncatus"* *Planorbis planorbis parenzani* *Ceratophallus bicarinatus* *C. blanfordi* *C. natalensis* *Biomphalaria pfeifferi/ barthi* *B. sudanica* *Burnupia* sp. *Caelatura aegyptiaca* *Corbicula pusilla* *Pisidium grovei* sp. n. *P. kenianum* *P. moitesserianum* *P. streetae* sp. n.	*Cyprideis* sp. (noded) *Ilyocypris* sp. *Metacypris* sp. *Darwinula* sp.	*Tilapia* cf. *nilotica* *Barbus* cf. *intermedius* *Barbus* sp. *Barbus* sp. (with serrated spines)? *B. paludinosus* *Synodontis* sp. ? *Lates* (crab)	
	6500		*Cyclotella meneghiniana* *Thalassiosira rudolfi* *Nitzschia* spp.				
	ca. 7000	<1561	*Rhopalodia gibberula* *Anomoeoneis sphaerophora* + var. *Nitzschia* spp. V. scarce				
ZIWAY — SHALA V	ca. 8000		*Cyclotella meneghiniana* *Thalassiosira rudolfi* *Nitzschia* spp.				
	8500 9400	1670	*Cyclotella ocellata* *Stephanodiscus astraea* + var. *Fragilaria brevistriata*	*Bellamya unicolor* *Valvata nilotica* *Melanoides tuberculata* *Lymnaea natalensis* *Bulinus "truncatus"* *Planorbis planorbis parenzani* *Ceratophallus natalensis* *Biomphalaria pfeifferi/ barthi* *Burnupia* sp. *Caelatura aegyptiaca* *C.* cf. *monceti* *Corbicula pusilla* *Sphaerium hartmanni* *Pisidium kenianum* *P. milium* *P. moitesserianum* *P. nitidum* *P. streetae* sp. n.	*Cyprideis* sp. (noded)	*Barbus* cf. *intermedius* *(B. paludinosus)* cichlid	
	9800		*Cyclotella meneghiniana* *Thalassiosira rudolfi* *T. faurii*				
	9950	<1579	*Cymbella pusilla*				
ZIWAY — SHALA IV	10,200 11,500	ca. 1632	*Cyclotella ocellata* *Melosira granulata* var. *valida* *Rhopalodia gibberula*	*Melanoides tuberculata* *Valvata nilotica* *Bulinus "truncatus"* Small planorbids		+	N
	ca. 13,000?	<<1594					
	? 21,000?	ca. 1641 C	?	V. rare — reworked? *(Valvata nilotica)* *Melanoides tuberculata* *(Lymnaea natalensis)* *(Bulinus "truncatus")* Small planorbids *Biomphalaria pfeifferi/ barthi* *Caelatura* sp. *Corbicula* sp. *(Pisidium margarithae* sp.n.) *P. moitesserianum* *(P. streetae* sp. n.) *(P. subtruncatum)*	+++	+	
ZIWAY…							

Table 8:1
Quaternary environmental changes in the Ziway-Shala Basin

...n)	Molluscs identified	Ostracods identified	Fish and crabs identified	Inferred characteristics of the surface waters: Chemistry	Temperature etc.	Sedimentation	Conditions in the lake catchment
	All reworked	++		Shala Ms, H?		S — small deltas and bouldery strandlines	Forest clearance
	All reworked	++	*Tilapia* sp.	Ms, M/MH?		A, S — beach ridges and small deltas, ostracod-rich sands and silts	1850 BP
						S — colluvial/alluvial loams	
						No deposits known	
...llata ...ulata	*Bellamya unicolor* *Valvata nilotica* *Melanoides tuberculata* *Lymnaea natalensis* *Bulinus "truncatus"* *Planorbis planorbis parenzani* *Ceratophallus bicarinatus* *C. blanfordi* *C. natalensis* *Biomphalaria pfeifferi/barthi* *B. sudanica* *Burnupia* sp. *Caelatura aegyptiaca* *Corbicula pusilla* *Pisidium grovei* sp. n. *P. kenianum* *P. moitesserianum* *P. streetae* sp. n.	*Cyprideis* sp. (noded) *Ilyocypris* sp. *Metacypris* sp. *Darwinula* sp.	*Tilapia* cf. *nilotica* *Barbus* cf. *intermedius* *Barbus* sp. *Barbus* sp. (with serrated spines)? *B. paludinosus* *Synodontis* sp. ? *Lates* (crab)	O, ML, >10 mg/l SiO_2	P/W <25-27°C Shala stratified before 5750 BP?	Z — shell beds B — shelly gravels and small deltas DW — marl, shell beds A — diatomaceous sands S — calcareous silts and shell beds, including Roricha fish bed Marginal diatomites?	Maximum extent of montane forest 6400 BP?
...neghiniana ...rudolfi				Ms, M		A — sands	
...bberula				Ms, MH		NB — colluvium DW — air-fall pumice, palaeosol A — pumice sands S — fluvial gravels	
...eghiniana ...rudolfi				Ms, M		A — silts and sands	
...llata ...us ...r. ...istriata	*Bellamya unicolor* *Valvata nilotica* *Melanoides tuberculata* *Lymnaea natalensis* *Bulinus "truncatus"* *Planorbis planorbis parenzani* *Ceratophallus natalensis* *Biomphalaria pfeifferi/barthi* *Burnupia* sp. *Caelatura aegyptiaca* *C.* cf. *monceti* *Corbicula pusilla* *Sphaerium hartmanni* *Pisidium kenianum* *P. milium* *P. moitesserianum* *P. nitidum* *P. streetae* sp. n.	*Cyprideis* sp. (noded)	*Barbus* cf. *intermedius* *(B. paludinosus)* cichlid	O, ML, <1 mg/l SiO_2	P/C	Z — shell beds DW — marl, shell beds A — calcareous, diatomaceous clays S — reworked shells Marginal diatomites?	Increasing extent of montane forest
...eghiniana ...rudolfi				Ms, M		A — calcareous, diatomaceous clays	Soil formation begins on Rift floor
...lla				Ms		DW — fluvial gravels A, S — palaeosol	Very dry. Montane forest locally present on mountains
...lata ...ulata ...berula	*Melanoides tuberculata* *Valvata nilotica* *Bulinus "truncatus"* Small planorbids		+	Ms, M/O, ML		DW — marl A — clayey silts S — swamp silts with shells	11,500 BP
						A — Abernosa pumice mbr. S — colluvial/alluvial loams, palaeosol	Ice caps completely melted
	V. rare — reworked? *(Valvata nilotica)* *Melanoides tuberculata* *(Lymnaea natalensis)* *(Bulinus "truncatus")* Small planorbids *Biomphalaria pfeifferi/barthi* *Caelatura* sp. *Corbicula* sp. *(Pisidium margarithae* sp.n.) *P. moitesserianum* *(P. streetae* sp. n.) *(P. subtruncatum)*	+++	+	Ms? ↑ O?	C? Little fringing swamp	K — Thin-bedded gravels and ostracod marls NB — shelly beach gravels K — calcareous silts and pumice gravels S — diatomaceous silts, shells, algal limestone?	
...llata) ...ulata	*Corbicula* sp.			O, ML, ...		DW — marls and calcareous ...	

	ca. 7000	<1561	*Rhopalodia gibberula* *Anomoeoneis sphaerophora* + var. *Nitzschia* spp. V. scarce				
	ca. 8000		*Cyclotella meneghiniana* *Thalassiosira rudolfi* *Nitzschia* spp.				
ZIWAY — SHALA V	8500 9400	1670	*Cyclotella ocellata* *Stephanodiscus astraea* + var. *Fragilaria brevistriata*	*Bellamya unicolor* *Valvata nilotica* *Melanoides tuberculata* *Lymnaea natalensis* *Bulinus "truncatus"* *Planorbis planorbis parenzani* *Ceratophallus natalensis* *Biomphalaria pfeifferi/ barthi* *Burnupia* sp. *Caelatura aegyptiaca* *C.* cf. *monceti* *Corbicula pusilla* *Sphaerium hartmanni* *Pisidium kenianum* *P. milium* *P. moitesserianum* *P. nitidum* *P. streetae* sp. n.	*Cyprideis* sp. (noded)	*Barbus* cf. *intermedius* (*B. paludinosus*) cichlid	C
	9800		*Cyclotella meneghiniana* *Thalassiosira rudolfi* *T. faurii*				
	9950	<1579	*Cymbella pusilla*				
ZIWAY — SHALA IV	10,200 11,500	ca. 1632	*Cyclotella ocellata* *Melosira granulata* var. *valida* *Rhopalodia gibberula*	*Melanoides tuberculata* *Valvata nilotica* *Bulinus "truncatus"* Small planorbids		+	Ms
	ca. 13,000?	<<1594					
ZIWAY — SHALA III	?	ca. 1641 C	?	V. rare — reworked? (*Valvata nilotica*) *Melanoides tuberculata* (*Lymnaea natalensis*) (*Bulinus "truncatus"*) Small planorbids *Biomphalaria pfeifferi/ barthi* *Caelatura* sp. *Corbicula* sp. (*Pisidium margarithae* sp. n.) *P. moitesserianum* (*P. streetae* sp. n.) (*P. subtruncatum*)	+++	+	
	21,000?	>>1622 B ? >>1617 A	(*Cyclotella ocellata*) (*Melosira granulata* var. *valida*, var. *jonensis*) (*Fragilaria brevistriata*)	*Corbicula* sp.			O
	24,000? 26,500?		(*Cyclotella ocellata*) (*Melosira granulata*) (*Thalassiosira rudolfi*) (*Synedra ulna*)	*Melanoides tuberculata* *Valvata nilotica* *Bulinus "truncatus"* Small planorbids *Caelatura* sp. *Corbicula* sp.			Ms
		<1591	(*Cyclotella meneghiniana* var. *pumila*) (*Rhopalodia gibberula*)				
ZIWAY — SHALA II	28,000? 30,000?	>1638 <1602 >1609	?	*Bulinus "truncatus"* *Corbicula* sp. Small planorbids	++		
		<1584					
ZIWAY — SHALA I	LOWER LATE PLEISTOCENE ?	>>1602	?				

Key

+	Present	O	Oligohaline (<2 g/l)	ML	Medium-low alkalinity (2-10 meq/l)	P	Pres
++	Common	Ms	Mesohaline (2-16 g/l)	M	Medium alkalinity (10-50 meq/l)	C	Sligh
+++	Abundant			MH	Medium-high alkalinity (50-100 meq/l)	W	Sligh
( )	Not dated (inferred correlation)			H	High alkalinity (100-200 meq/l)		

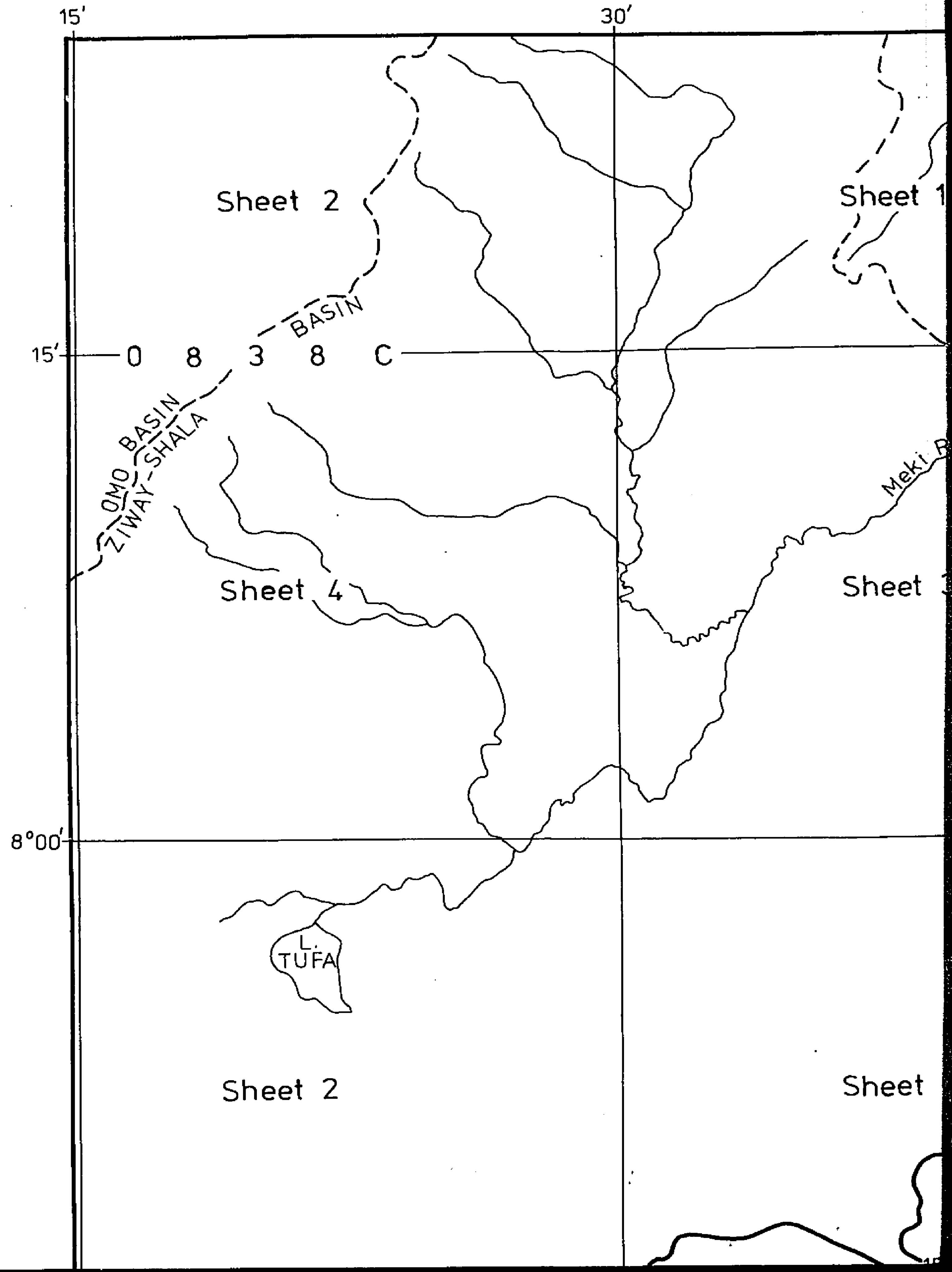
berula				Ms, MH		NB — colluvium DW — air-fall pumice, palaeosol A — pumice sands S — fluvial gravels	
eghiniana *udolfi*				Ms, M		A — silts and sands	
ata *s* *striata*	*Bellamya unicolor* *Valvata nilotica* *Melanoides tuberculata* *Lymnaea natalensis* *Bulinus "truncatus"* *Planorbis planorbis parenzani* *Ceratophallus natalensis* *Biomphalaria pfeifferi/barthi* *Burnupia* sp. *Caelatura aegyptiaca* *C.* cf. *monceti* *Corbicula pusilla* *Sphaerium hartmanni* *Pisidium kenianum* *P. milium* *P. moitesserianum* *P. nitidum* *P. streetae* sp. n.	*Cyprideis* sp. (noded)	*Barbus* cf. *intermedius* *(B. paludinosus)* cichlid	O, ML, <1 mg/l SiO_2	P/C	Z — shell beds DW — marl, shell beds A — calcareous, diatomaceous clays S — reworked shells Marginal diatomites?	Increasing extent of montane forest
neghiniana *rudolfi*				Ms, M		A — calcareous, diatomaceous clays	Soil formation begins on Rift floor
illa				Ms		DW — fluvial gravels A, S — palaeosol	Very dry. Montane forest locally present on mountains 11,500 BP
llata *ulata* *bberula*	*Melanoides tuberculata* *Valvata nilotica* *Bulinus "truncatus"* Small planorbids		+	Ms, M/O, ML		DW — marl A — clayey silts S — swamp silts with shells	
						A — Abernosa pumice mbr. S — colluvial/alluvial loams, palaeosol	Ice caps completely melted
	V. rare — reworked? *(Valvata nilotica)* *Melanoides tuberculata* *(Lymnaea natalensis)* *(Bulinus "truncatus")* Small planorbids *Biomphalaria pfeifferi/barthi* *Caelatura* sp. *Corbicula* sp. *(Pisidium margarithae* sp.n.) *P. moitesserianum* *(P. streetae* sp. n.) *(P. subtruncatum)*	+++	+	Ms? ↑ O?	C? Little fringing swamp	K — Thin-bedded gravels and ostracod marls NB — shelly beach gravels K — calcareous silts and pumice gravels S — diatomaceous silts, shells, algal limestone?	
llata) *ulata* *s)* *vistriata)*	*Corbicula* sp.			O, ML, >10 mg/l SiO_2		DW — marls and calcareous sands K, A — noncalcareous diatomites S — diatomaceous mudstones Algal limestones?	Boraluku glaciation??.
llata) *ulata)* *rudolfi)* *)*	*Melanoides tuberculata* *Valvata nilotica* *Bulinus "truncatus"* Small planorbids *Caelatura* sp. *Corbicula* sp.			Ms, M/O, ML		DW — shelly, fluviodeltaic sands K — calcareous diatomites	
neghiniana *bberula)*				M, Ms?		DW — fluvial gravels, palaeosol K — shelly pumice sands A — pumice sands?	
	Bulinus "truncatus" *Corbicula* sp. Small planorbids	++		O/Ms?		DW — sands with ostracods K — impure diatomite A — diatomite? S — marginal diatomite	
						S — deltaic sands	
						B — alluvial-fan gravels A — pumice sands and gravels with diatoms S — colluvium and fluvial gravels	
				?	Shala permanently stratified?	S — laminated diatomite	

g/l)
6 g/l)

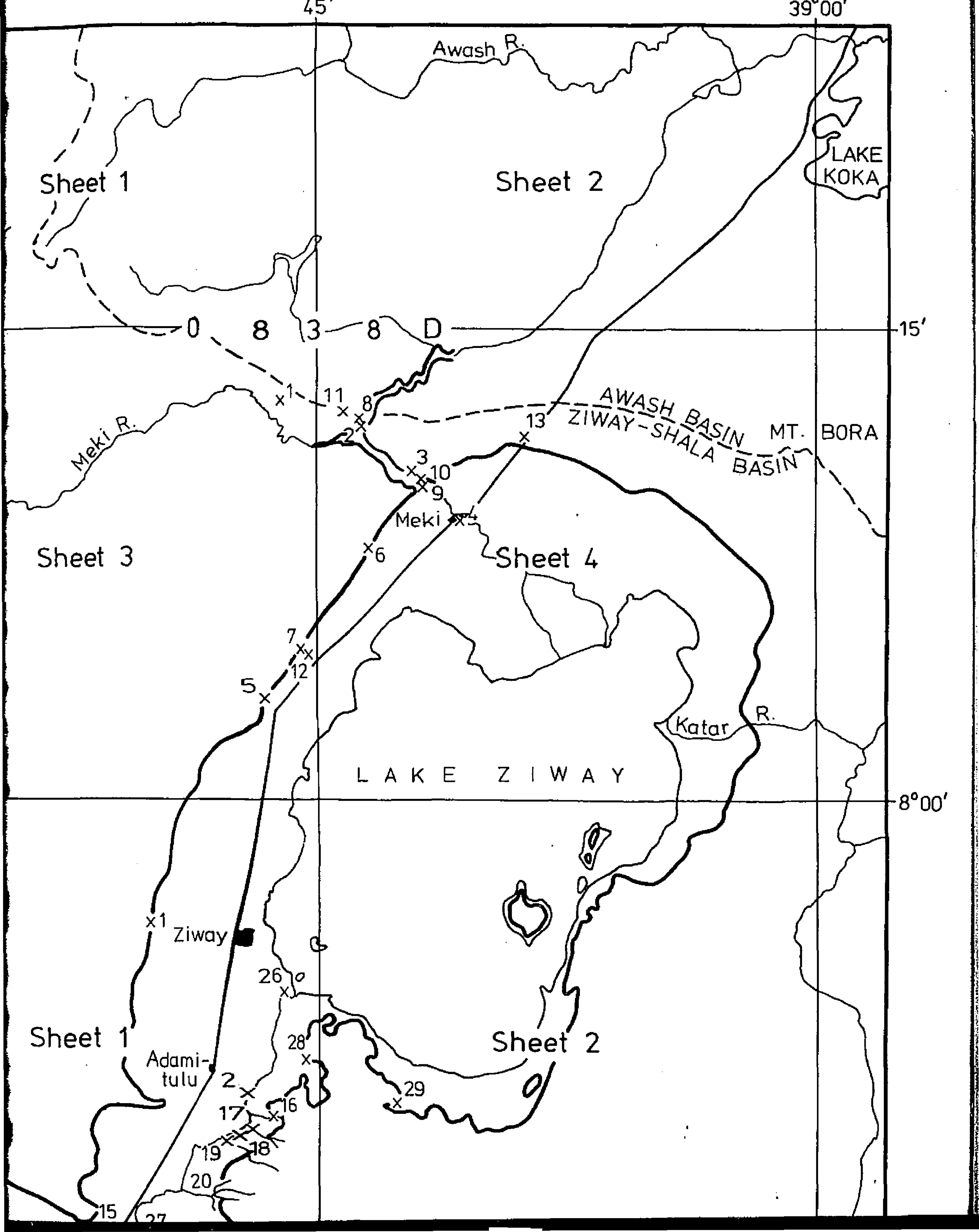
ML Medium-low alkalinity (2-10 meq/l)
M Medium alkalinity (10-50 meq/l)
MH Medium-high alkalinity (50-100 meq/l)
H High alkalinity (100-200 meq/l)

P Present-day range
C Slightly colder
W Slightly warmer

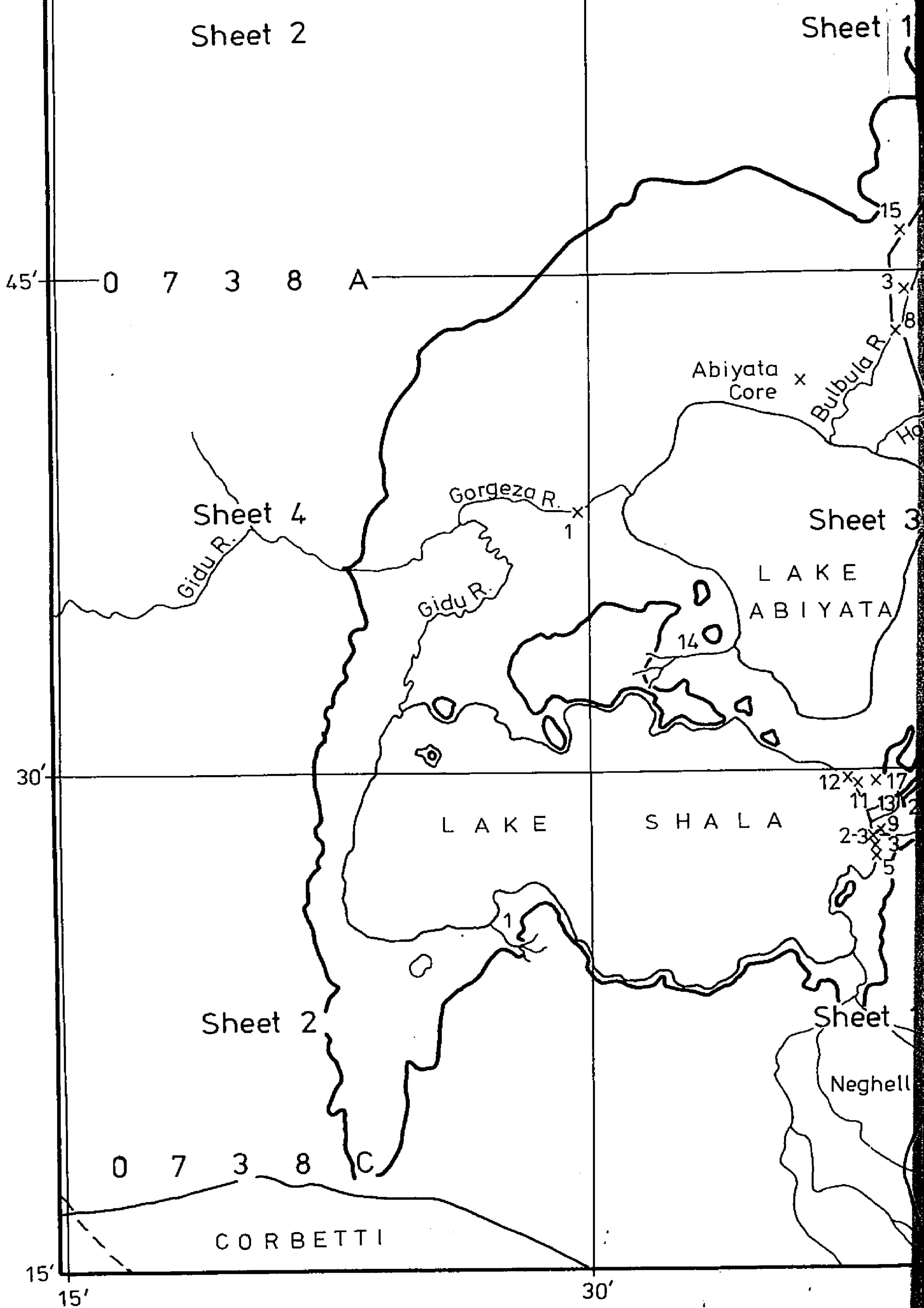
Z Ziway
NB Northern Bulbula Plain
B Bulbula Plain
DW Deka Wede Valley
K Kurkura Area
A Abiyata Core
S Shala Basin



15′
30′
Sheet 2
Sheet 1
BASIN
15′
0 8 3 8 C
OMO BASIN
ZIWAY-SHALA
Meki R
Sheet 4
Sheet 3
8°00′
L. TUFA
Sheet 2
Sheet



45'
39°00'
Awash R.
LAKE KOKA
Sheet 1
Sheet 2
0 8 3 8 D
15'
Meki R.
AWASH BASIN
ZIWAY-SHALA BASIN
MT. BORA
Meki
Sheet 3
Sheet 4
Katar R.
LAKE ZIWAY
8°00'
Ziway
Sheet 1
Adami-tulu
Sheet 2



Sheet 2
Sheet 1
15
3
8
45'
0 7 3 8 A
Abiyata
Core
Bulbula R.
Ho
Gorgeza R.
1
Sheet 4
Sheet 3
Gidu R.
Gidu R.
LAKE
ABIYATA
14
30'
12
17
11
13
9
2-3
3
5
LAKE SHALA
1
Sheet 2
Sheet 1
Neghell
0 7 3 8 C
CORBETTI
15'
15'
30'

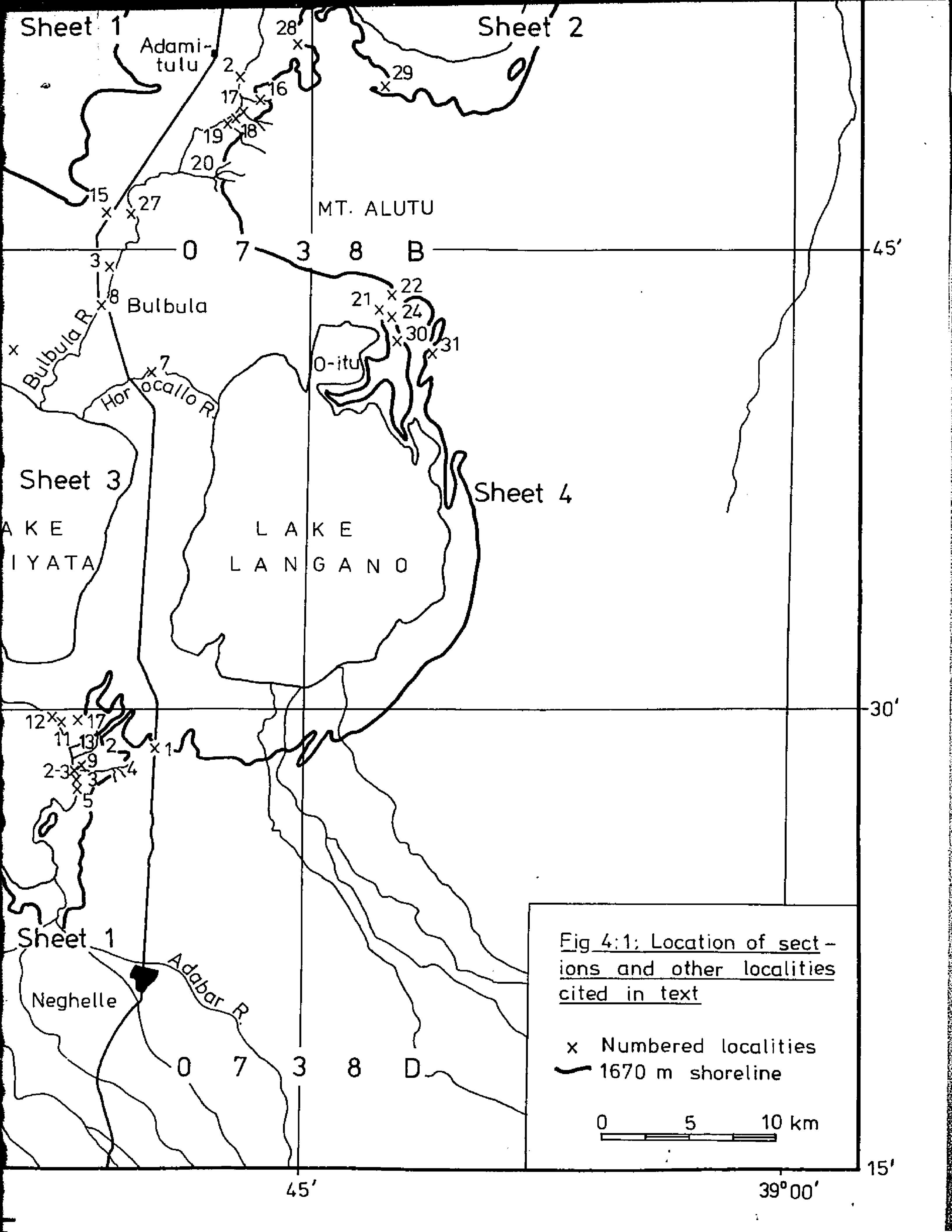
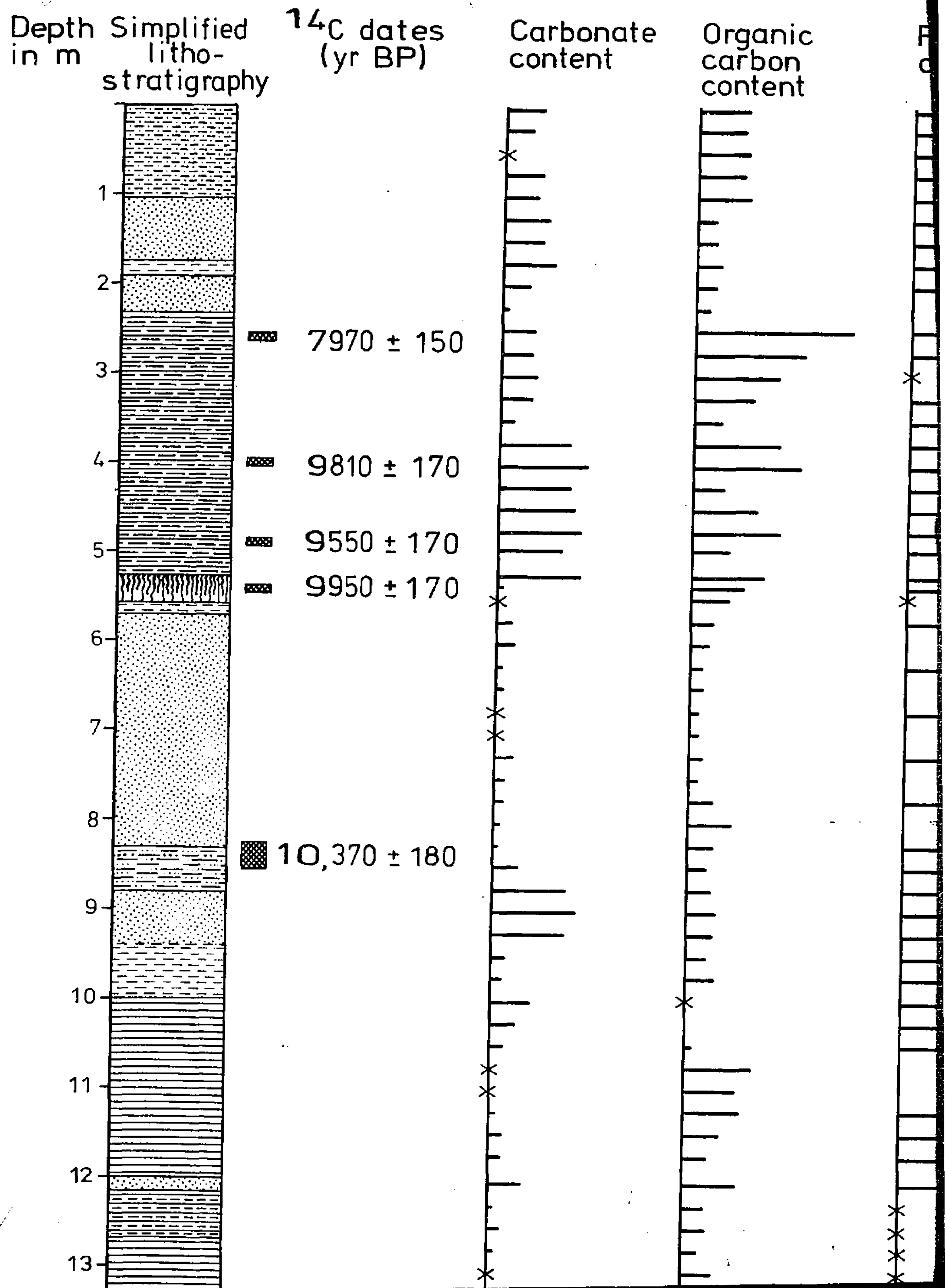
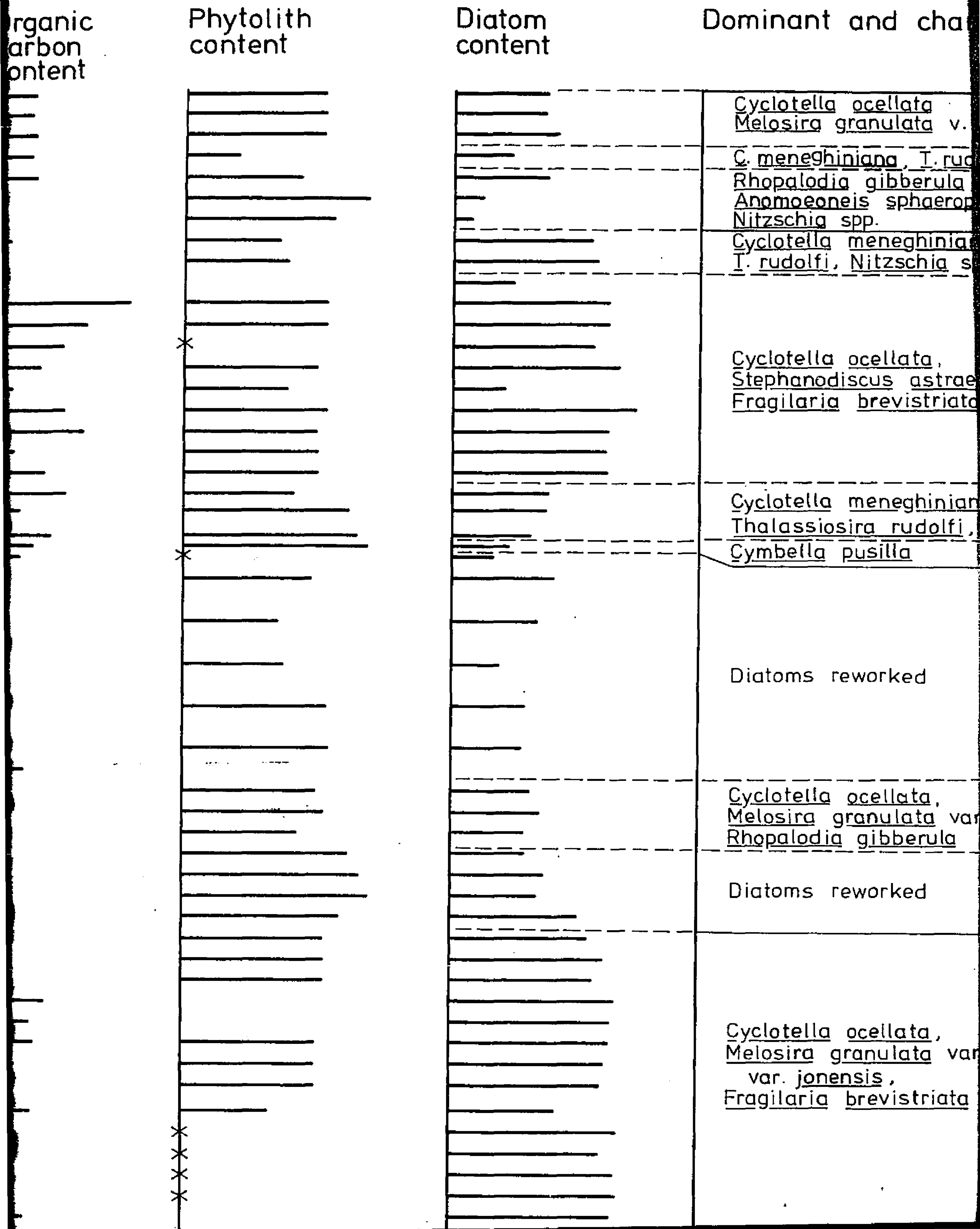


Fig 4:1; Location of sections and other localities cited in text

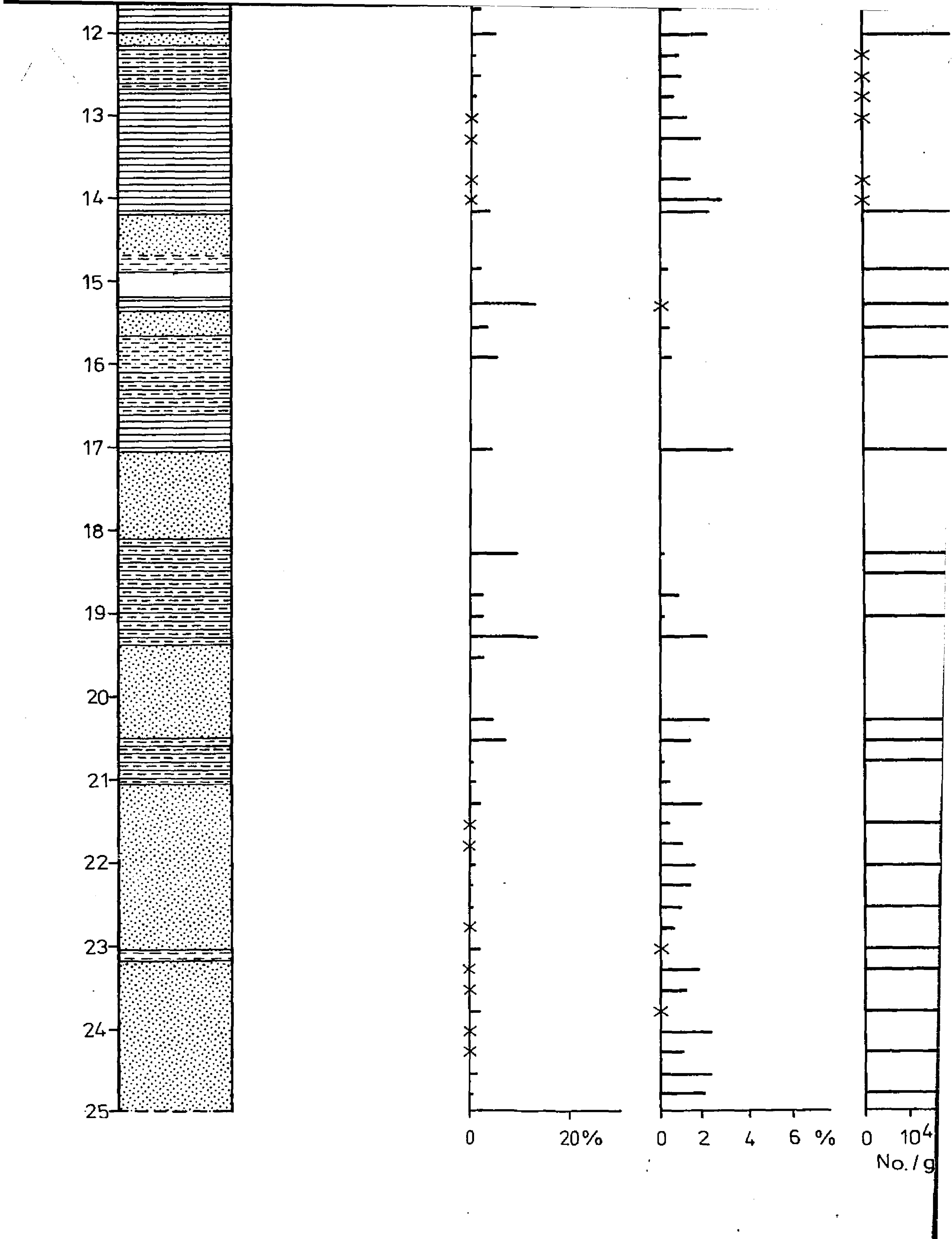


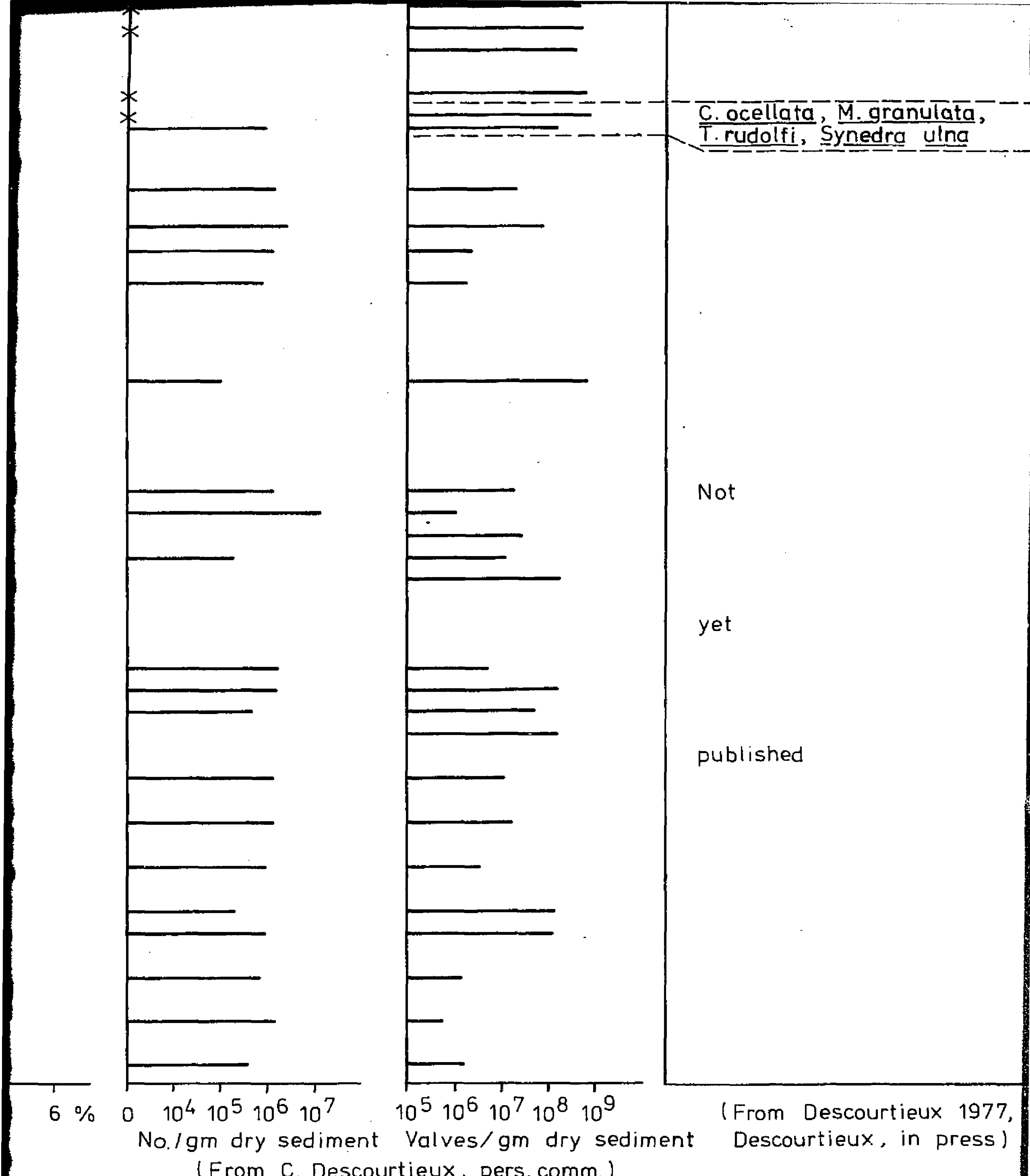
Depth in m
Simplified litho-stratigraphy
^{14}C dates (yr BP)
Carbonate content
Organic carbon content
7970 ± 150
9810 ± 170
9550 ± 170
9950 ± 170
10,370 ± 180
1
2
3
4
5
6
7
8
9
10
11
12
13



rganic
arbon
ontent
Phytolith
content
Diatom
content
Dominant and cha
Cyclotella ocellata
Melosira granulata v.
C. meneghiniana, T. rud
Rhopalodia gibberula
Anomoeoneis sphaerop
Nitzschia spp.
Cyclotella meneghinia
T. rudolfi, Nitzschia s
Cyclotella ocellata,
Stephanodiscus astrae
Fragilaria brevistriata
Cyclotella meneghinian
Thalassiosira rudolfi,
Cymbella pusilla
Diatoms reworked
Cyclotella ocellata,
Melosira granulata var
Rhopalodia gibberula
Diatoms reworked
Cyclotella ocellata,
Melosira granulata var
var. jonensis,
Fragilaria brevistriata

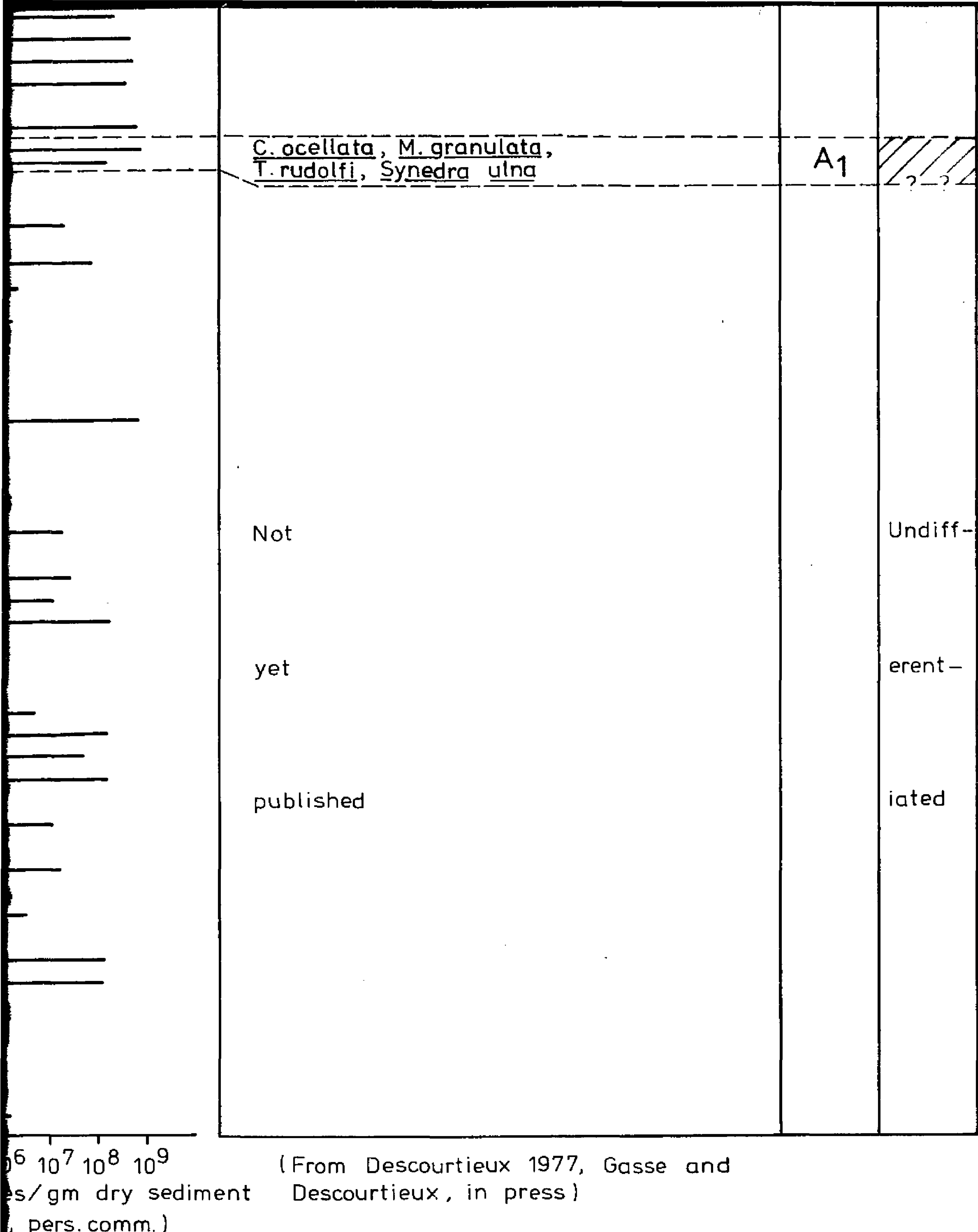
...om ...ent	Dominant and characteristic diatoms	Diatom zones	Lacustral phases
	Cyclotella ocellata *Melosira granulata* v. *valida*	D_3	IV
	C. meneghiniana, *T. rudolfi*, *Nitzschia* spp.	D2	
	Rhopalodia gibberula, *Anomoeoneis sphaerophora* + var., *Nitzschia* spp.	D_1	
	Cyclotella meneghiniana, *T. rudolfi*, *Nitzschia* spp.	C_4	
	Cyclotella ocellata, *Stephanodiscus astraea* + var., *Fragilaria brevistriata*	C_3	III
	Cyclotella meneghiniana, *Thalassiosira rudolfi*, *T. faurii*,	C_2	
	Cymbella pusilla	C1	
	Diatoms reworked		
	Cyclotella ocellata, *Melosira granulata* var. *valida*, *Rhopalodia gibberula*	B	II
	Diatoms reworked		
	Cyclotella ocellata, *Melosira granulata* var. *valida*, var. *jonensis*, *Fragilaria brevistriata*	A_2	I


12
13
14
15
16
17
18
19
20
21
22
23
24
25
0
20%
0
2
4
6 %
0
10⁴
No. / g



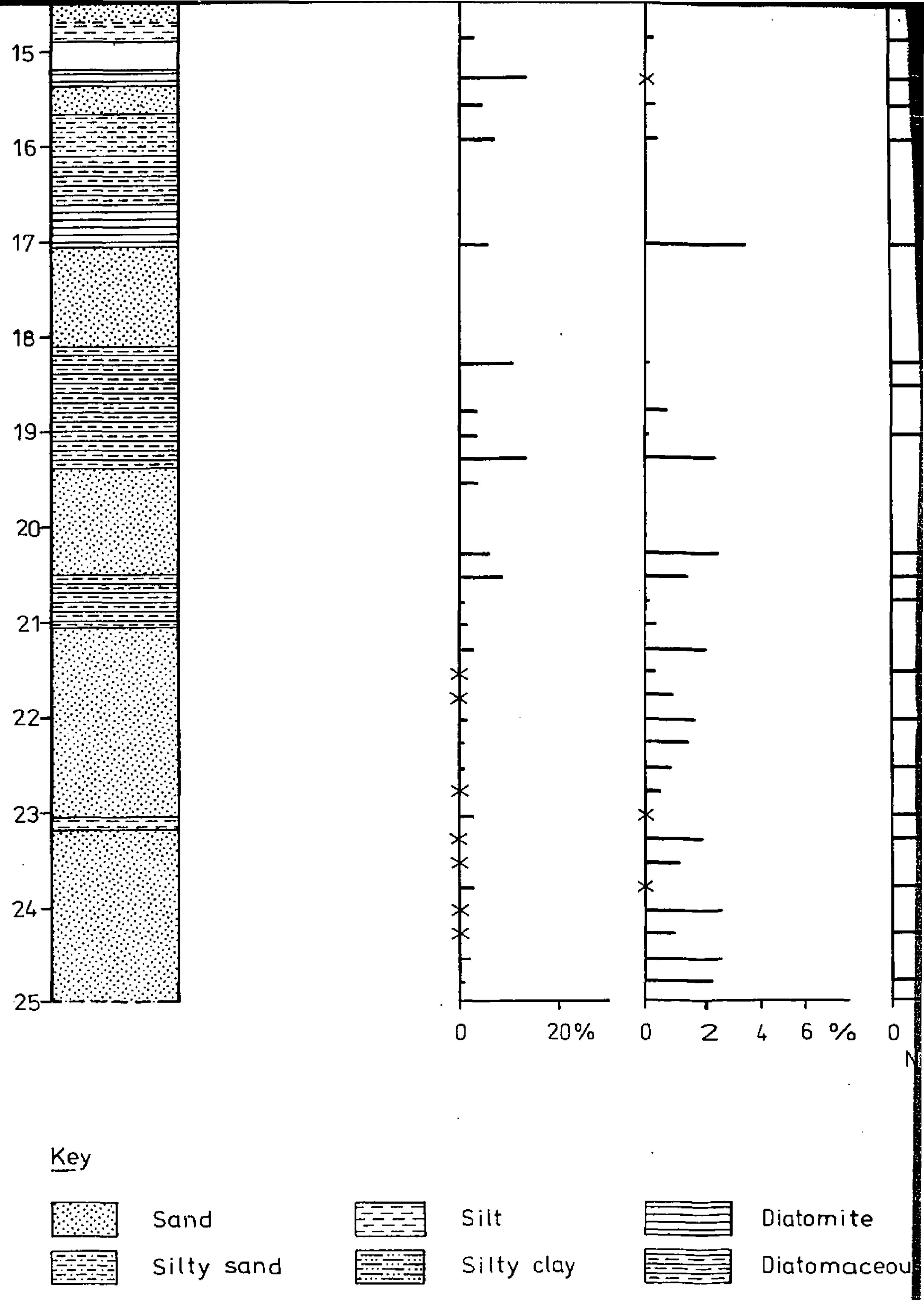
Diatomite

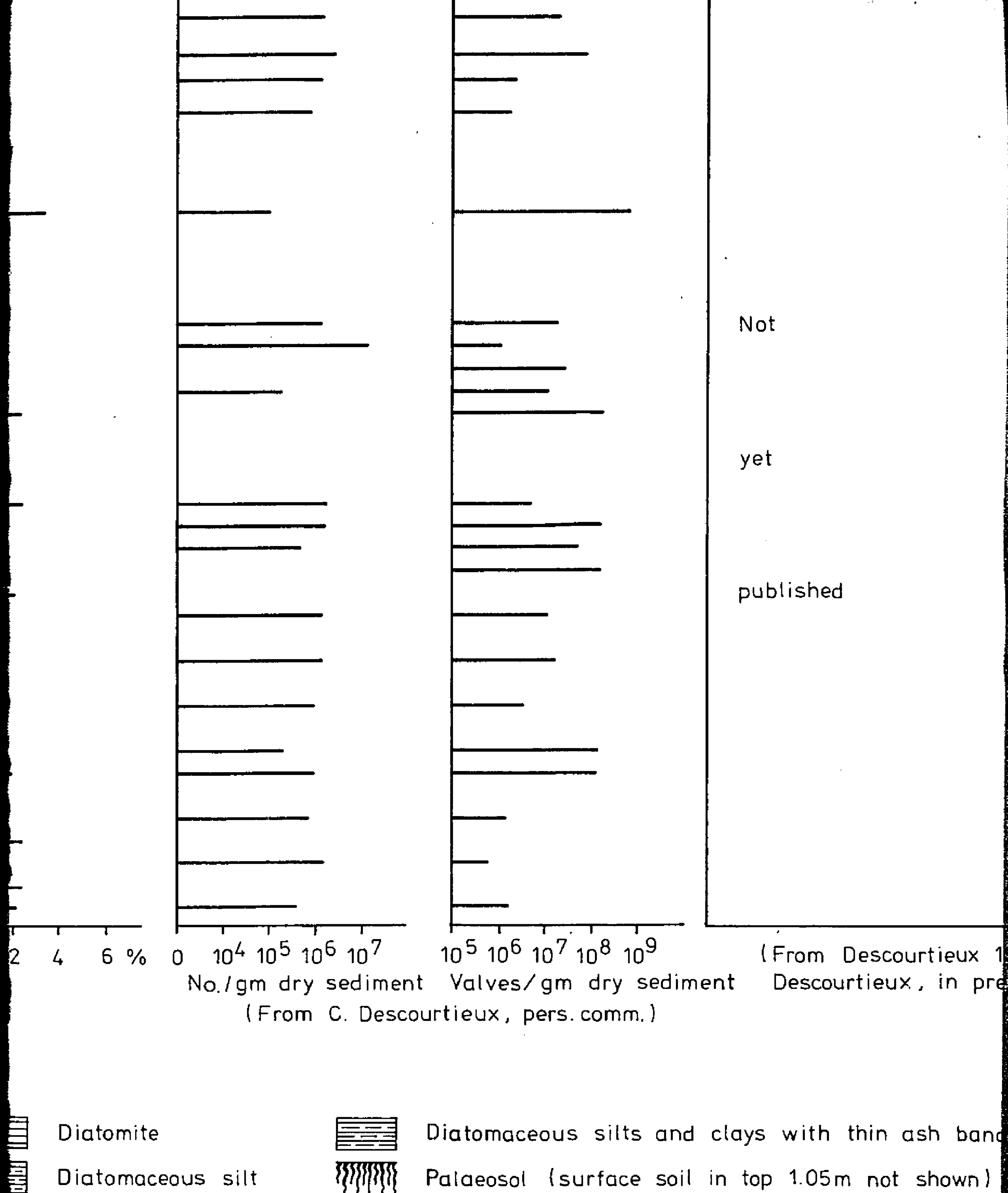
Diatomaceous silts and clays with thin ash bands



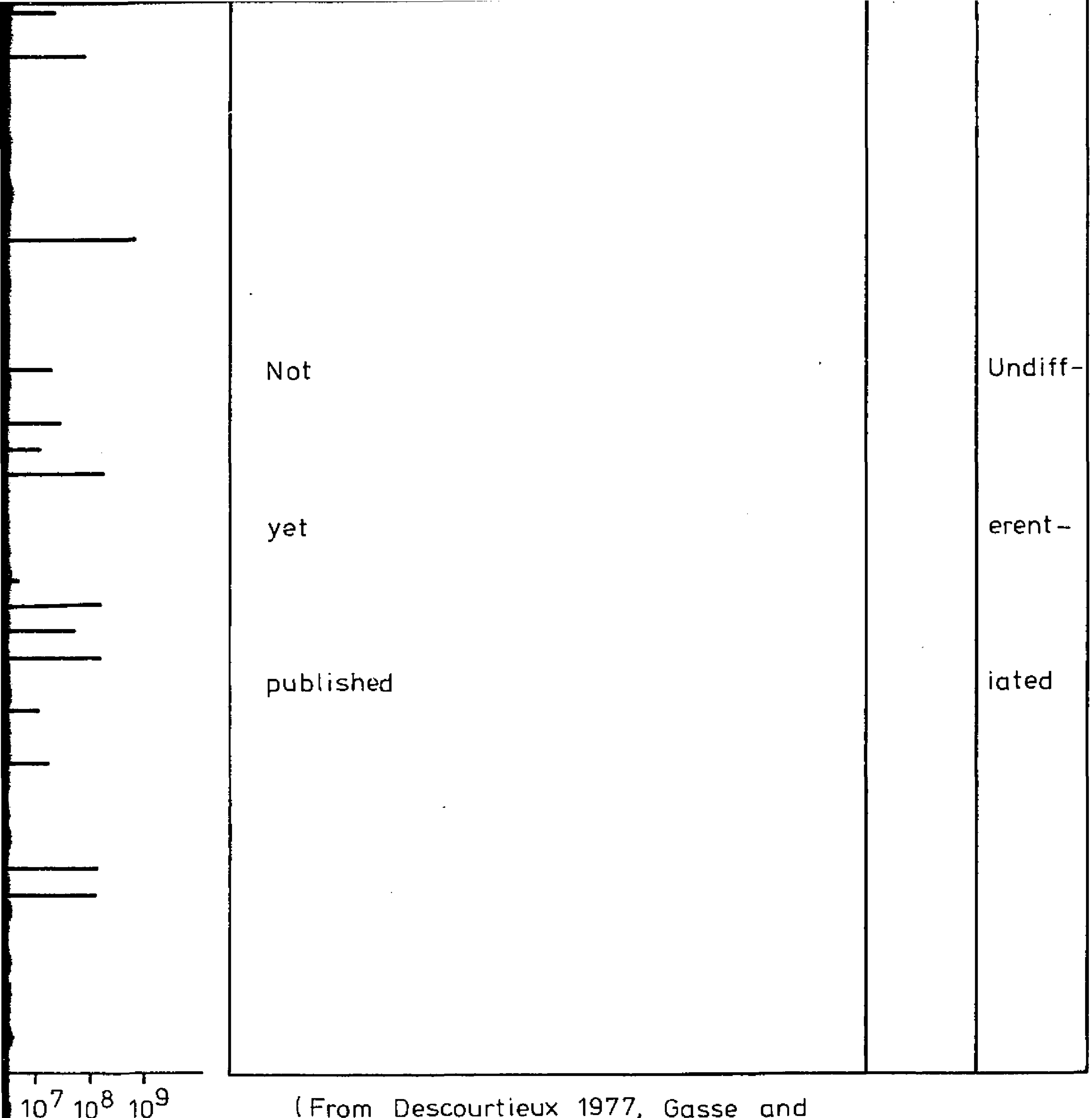
(From Descourtieux 1977, Gasse and Descourtieux, in press)

aceous silts and clays with thin ash bands Core disturbed


15
16
17
18
19
20
21
22
23
24
25
0
20%
0
2
4
6
%
0
Key
Sand
Silty sand
Silt
Silty clay
Diatomite
Diatomaceou



Not
yet
published
2 4 6 %
0 10^4 10^5 10^6 10^7
No./gm dry sediment
10^5 10^6 10^7 10^8 10^9
Valves/gm dry sediment
(From C. Descourtieux, pers. comm.)
(From Descourtieux 1
Descourtieux, in pre
Diatomite
Diatomaceous silt
Diatomaceous silts and clays with thin ash band
Palaeosol (surface soil in top 1.05m not shown)



/gm dry sediment

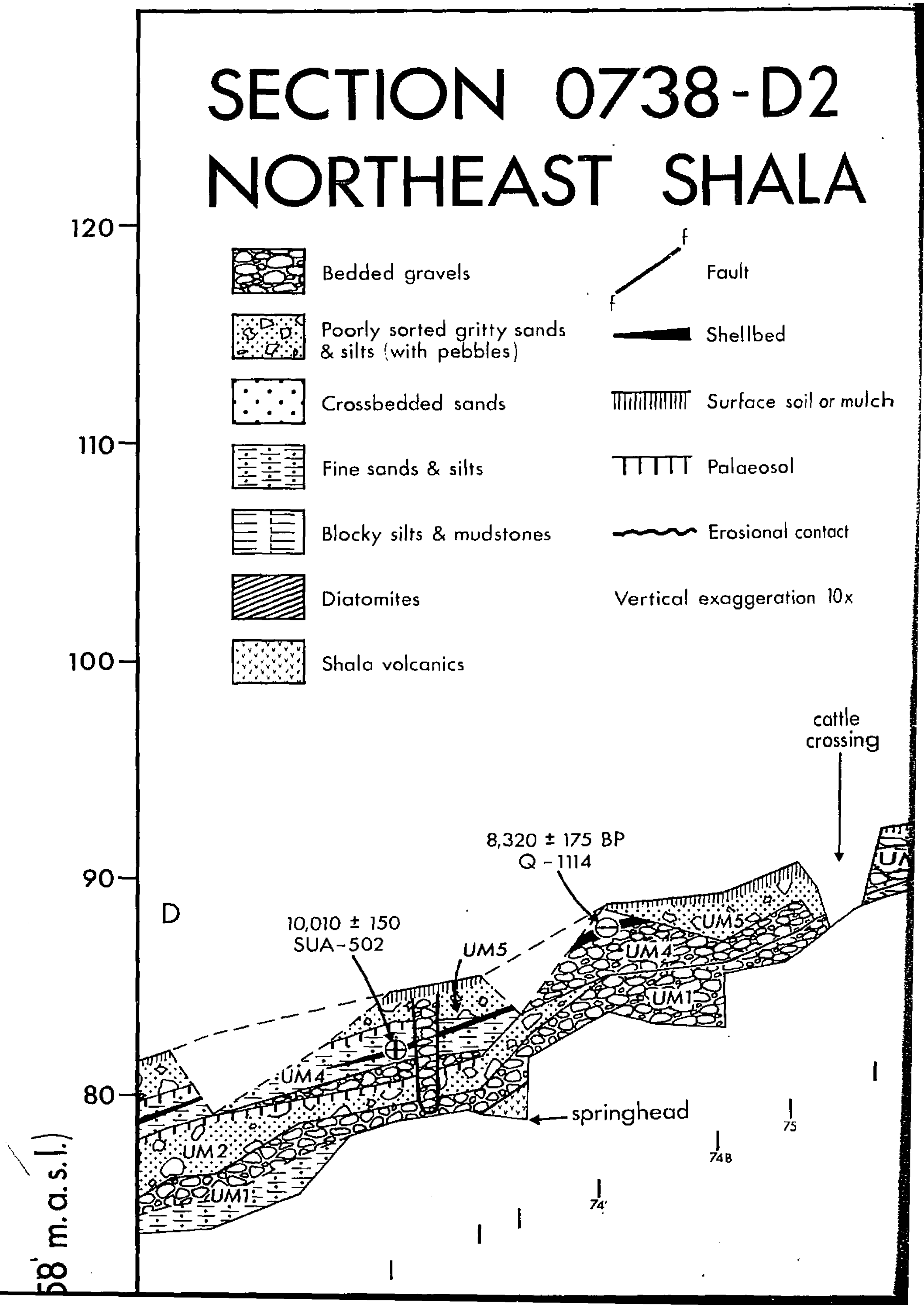
pers. comm.)

(From Descourtieux 1977, Gasse and Descourtieux, in press)

eous silts and clays with thin ash bands

Core disturbed

(surface soil in top 1.05m not shown)


SECTION 0738-D2
NORTHEAST SHALA
Bedded gravels
Poorly sorted gritty sands & silts (with pebbles)
Crossbedded sands
Fine sands & silts
Blocky silts & mudstones
Diatomites
Shala volcanics
Fault
Shellbed
Surface soil or mulch
Palaeosol
Erosional contact
Vertical exaggeration 10x
120
110
100
90
80
m.a.s.l.)
D
10,010 ± 150
SUA-502
UM5
8,320 ± 175 BP
Q-1114
cattle crossing
UM5
UM4
UM1
UM4
UM2
UM1
springhead
75
74B
74'

-D2

HALA

ault

ellbed

rface soil or mulch

laeosol

osional contact

ggeration 10x

Dating materials

▲ Carbonized wood

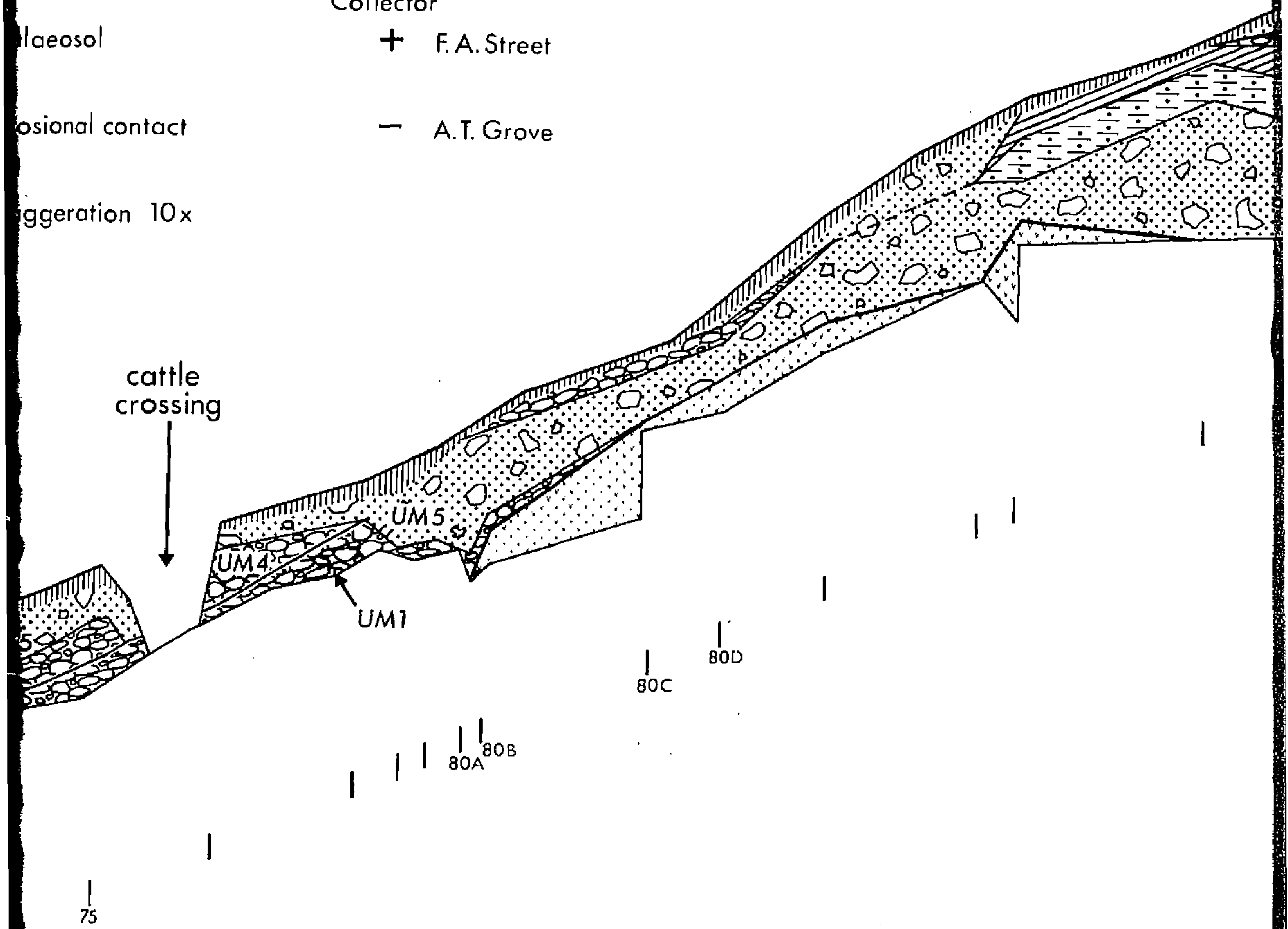
○ Shells

□ Algal limestone

Collector

+ F. A. Street

— A. T. Grove



-D2

HALA

ault

ellbed

urface soil or mulch

laeosol

osional contact

ggeration 10x

Dating materials

▲ Carbonized wood

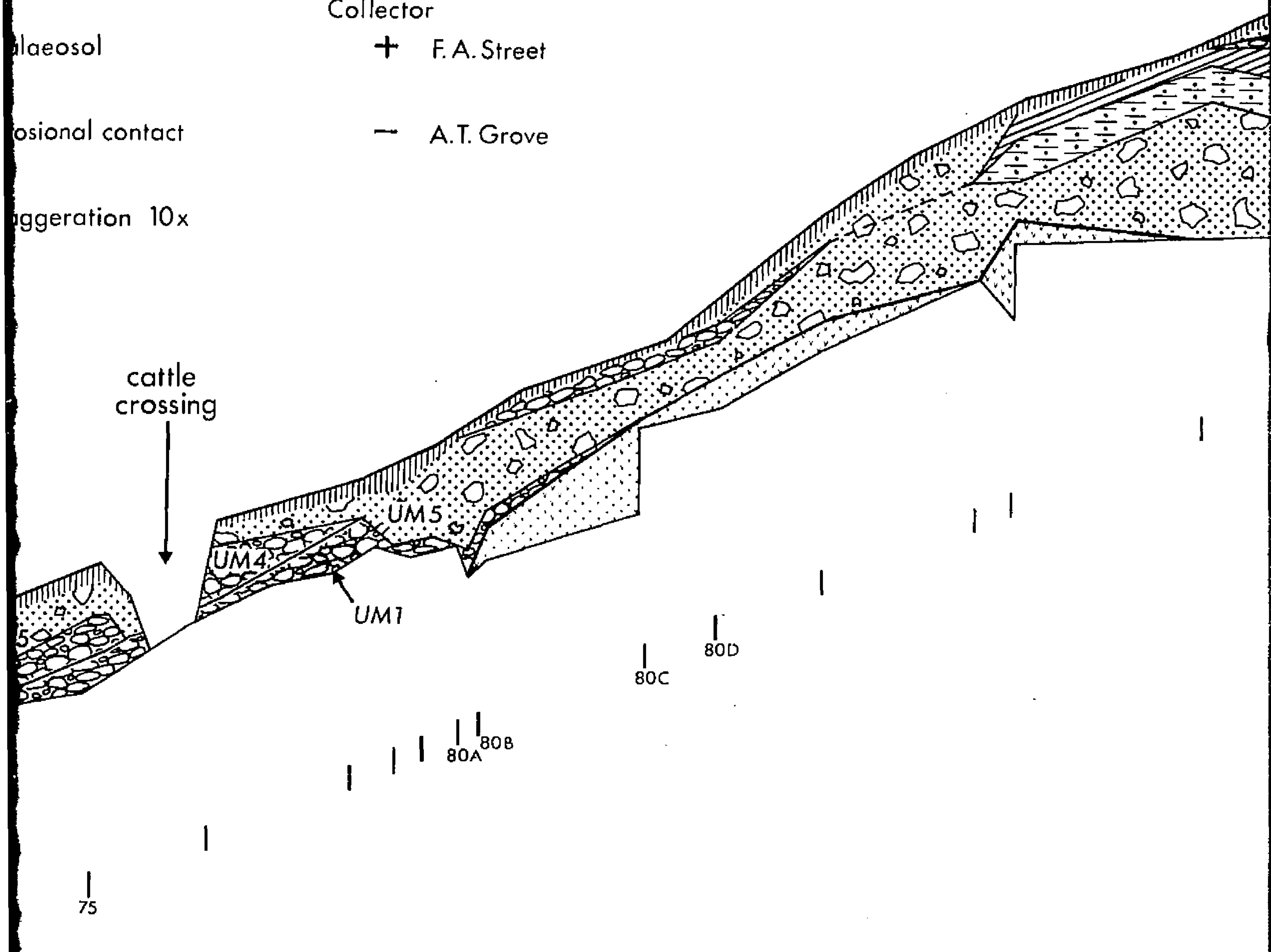
○ Shells

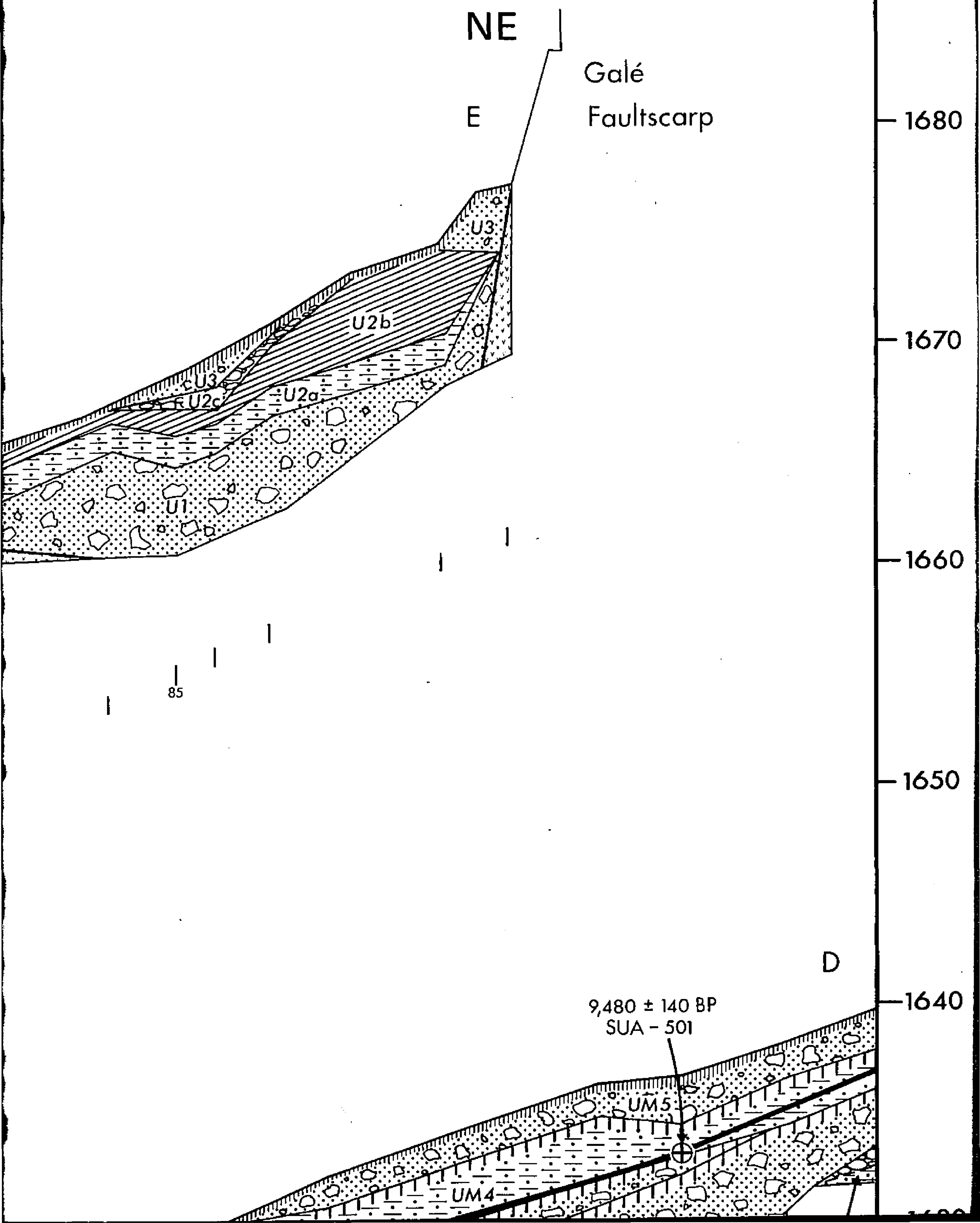
□ Algal limestone

Collector

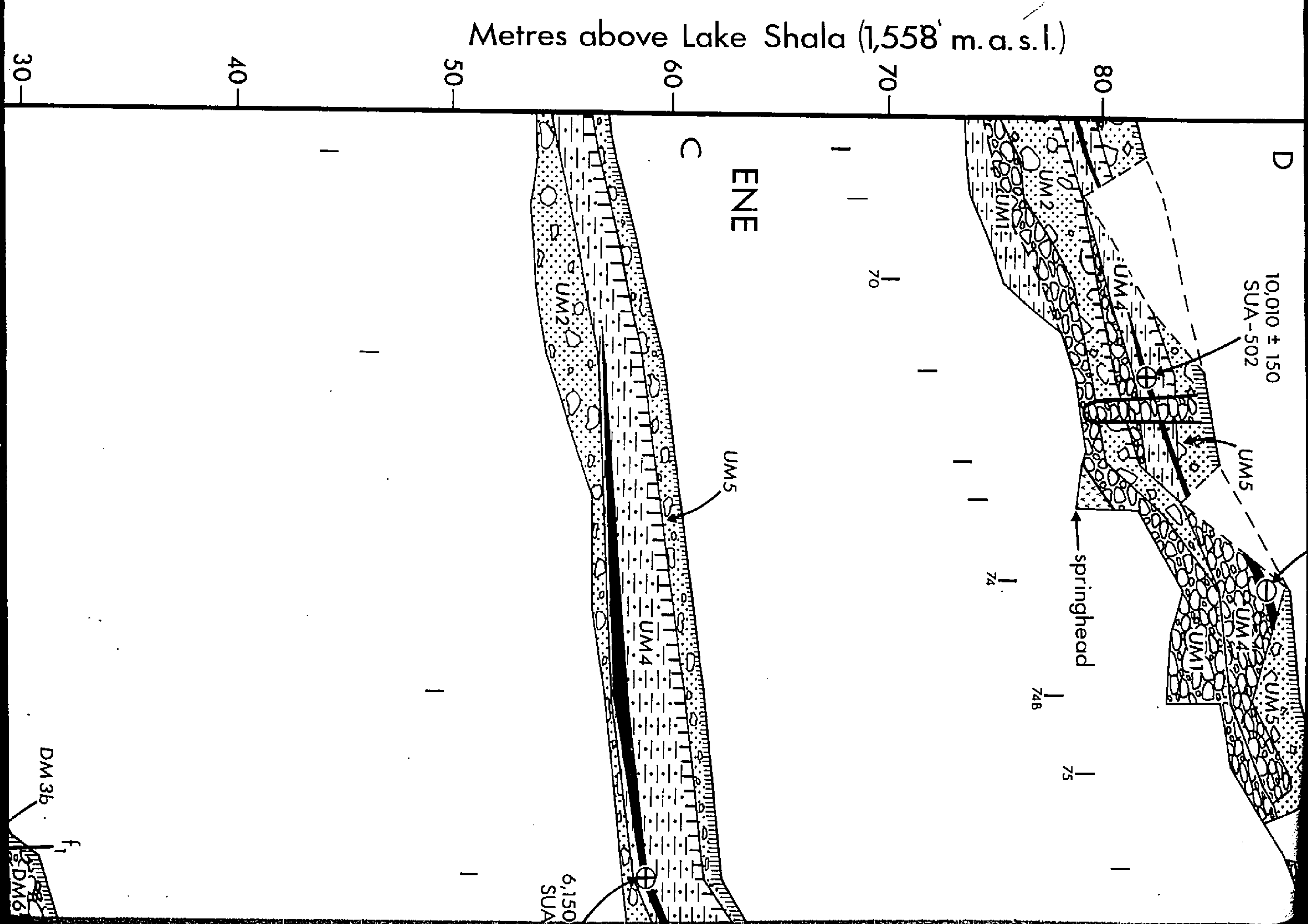
+ F. A. Street

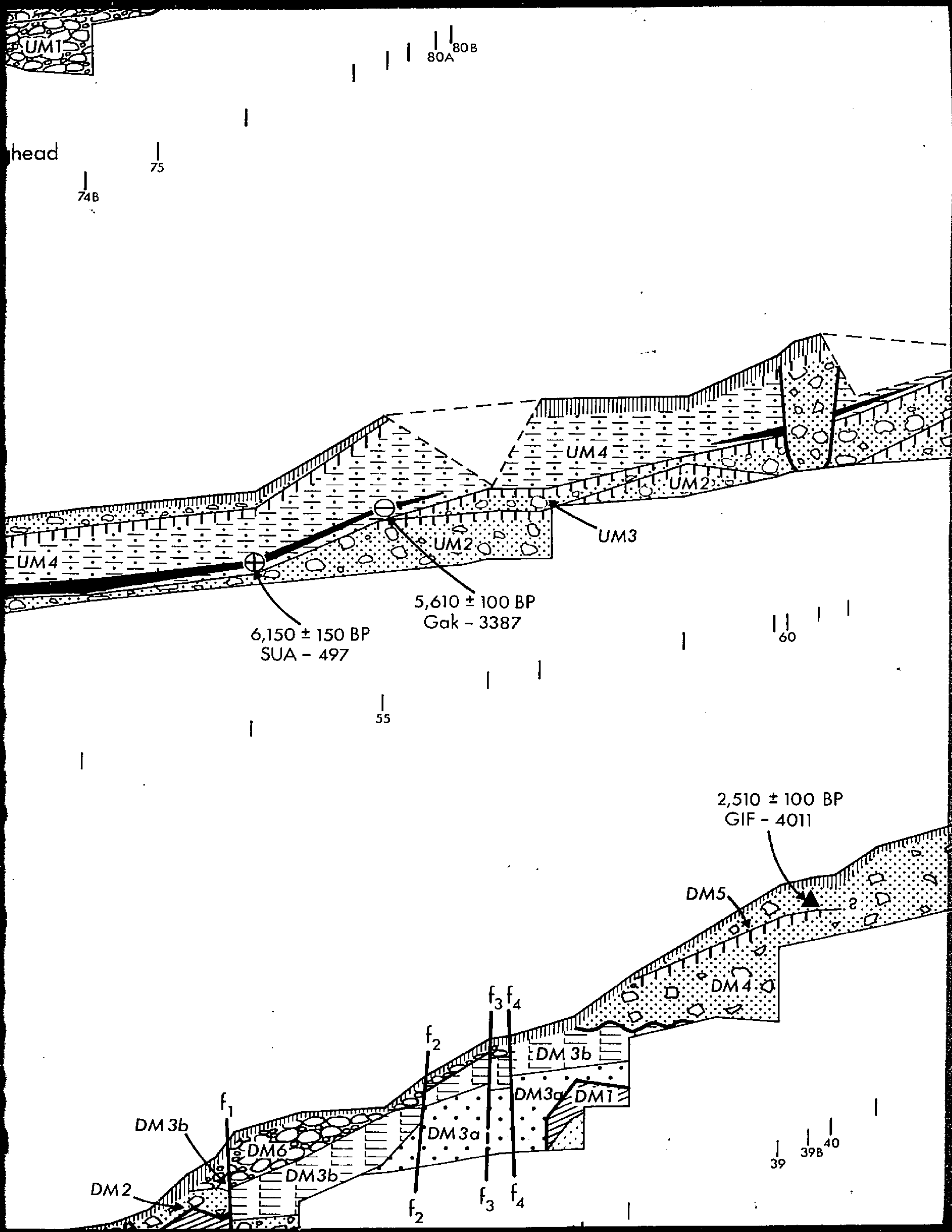
– A. T. Grove



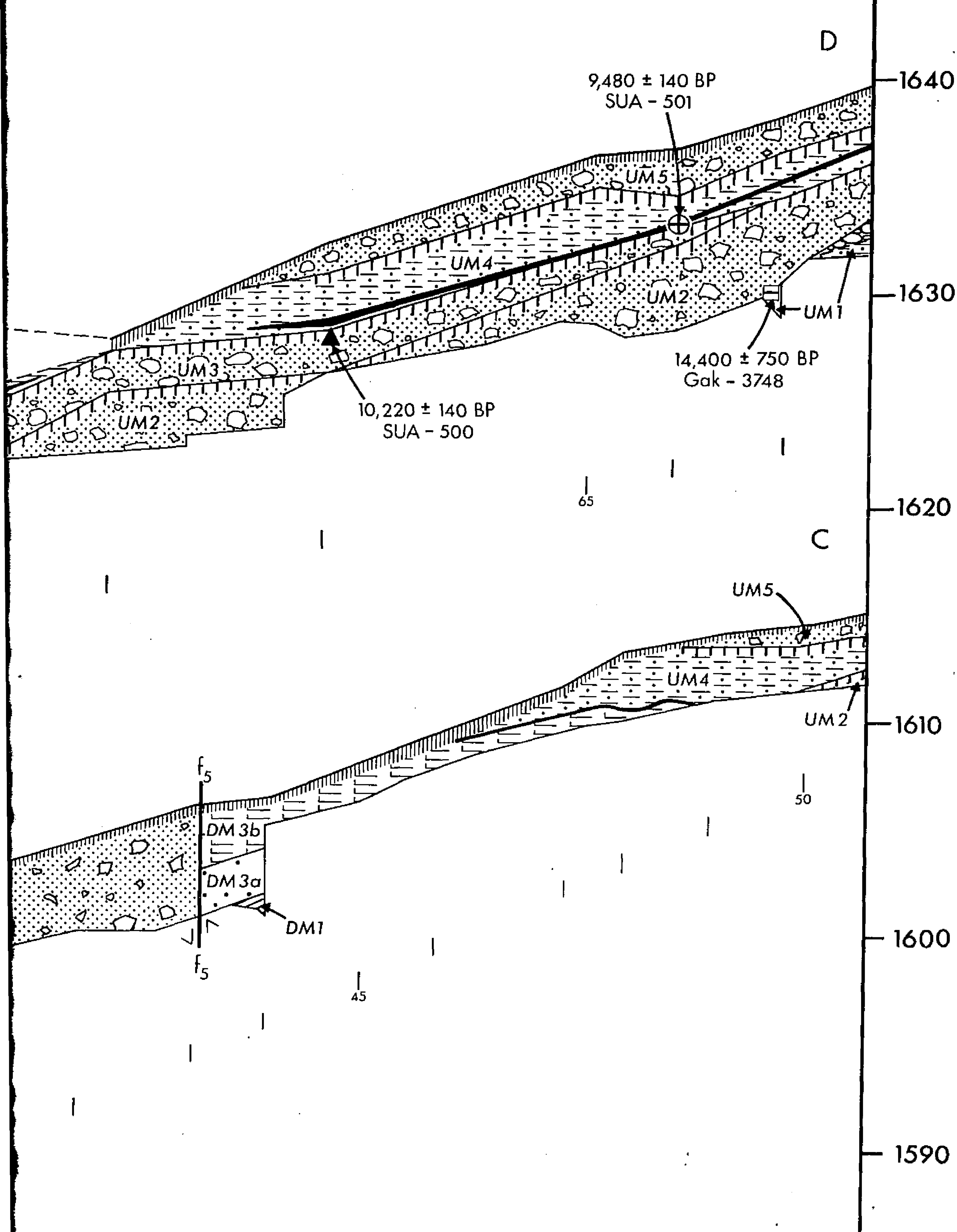


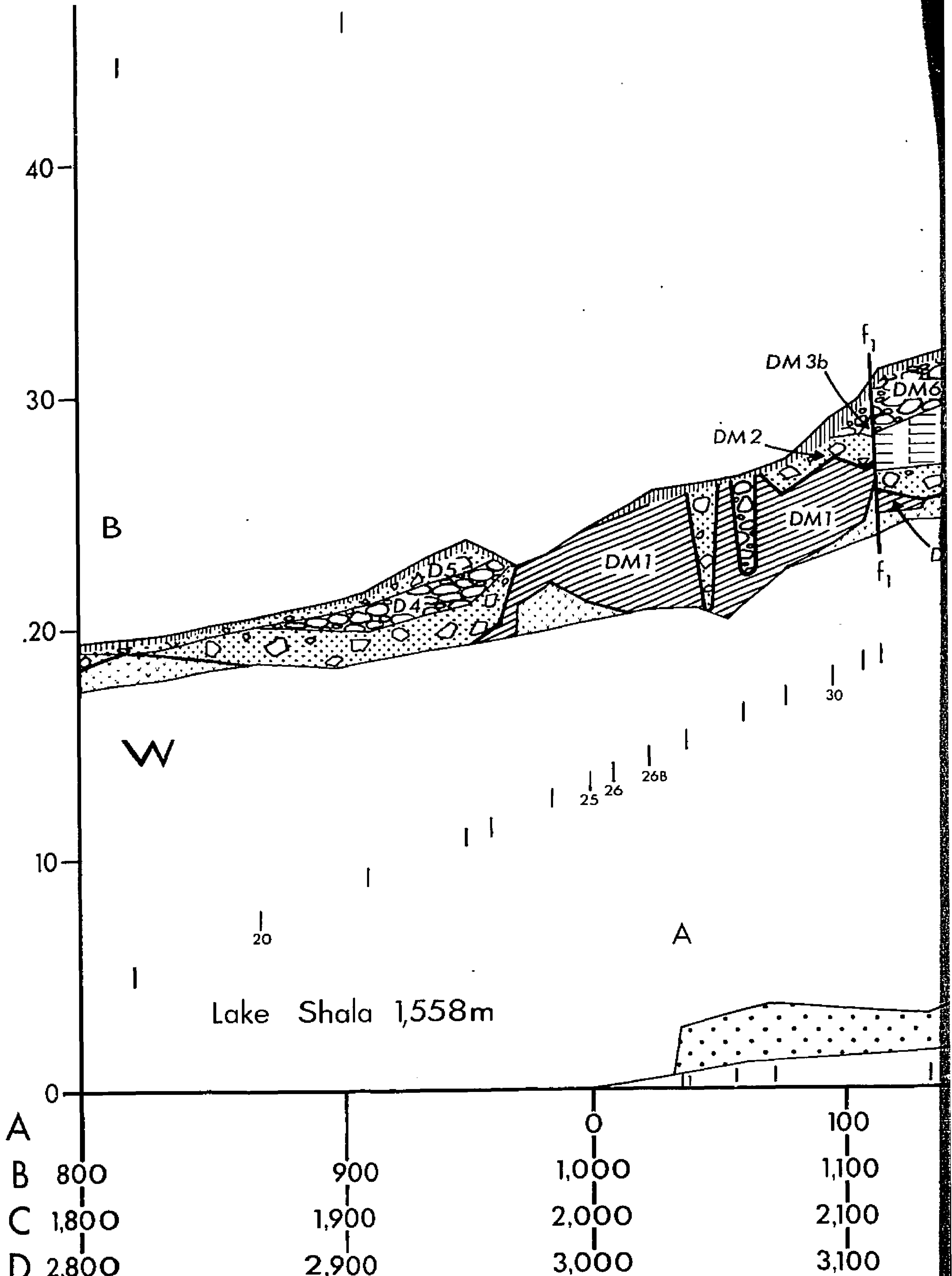
NE
Galé
E
Faultscarp
1680
1670
1660
1650
1640
U3
U2b
U3
U2c
U2a
U1
85
D
9,480 ± 140 BP
SUA - 501
UM5
UM4

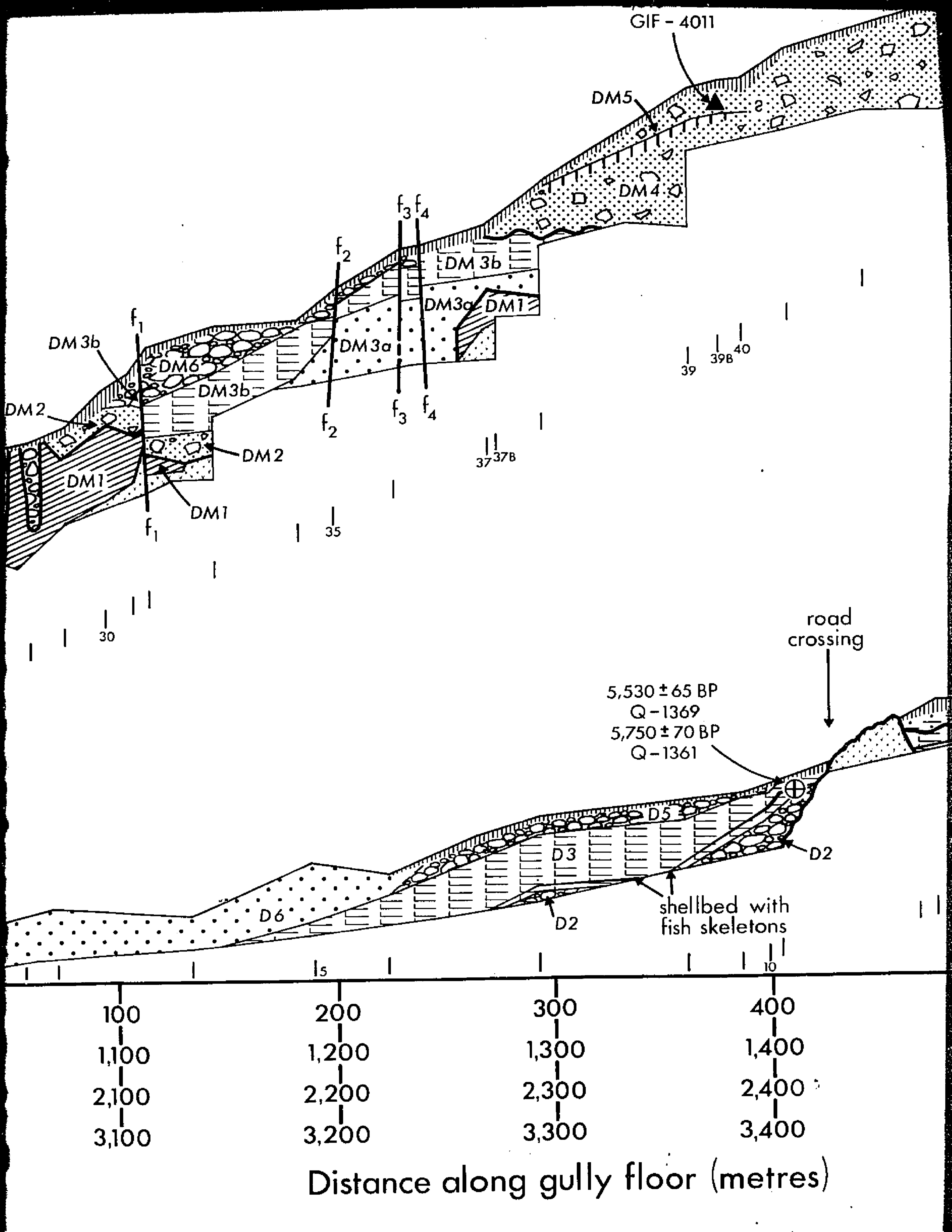

Metres above Lake Shala (1,558' m.a.s.l.)
30
40
50
60
70
80
C
ENE
D
UM2
UM4
UM5
UM1
springhead
10,010 ± 150
SUA-502
6,150
SUA
DM 3b
DM6
70
74
74B
75



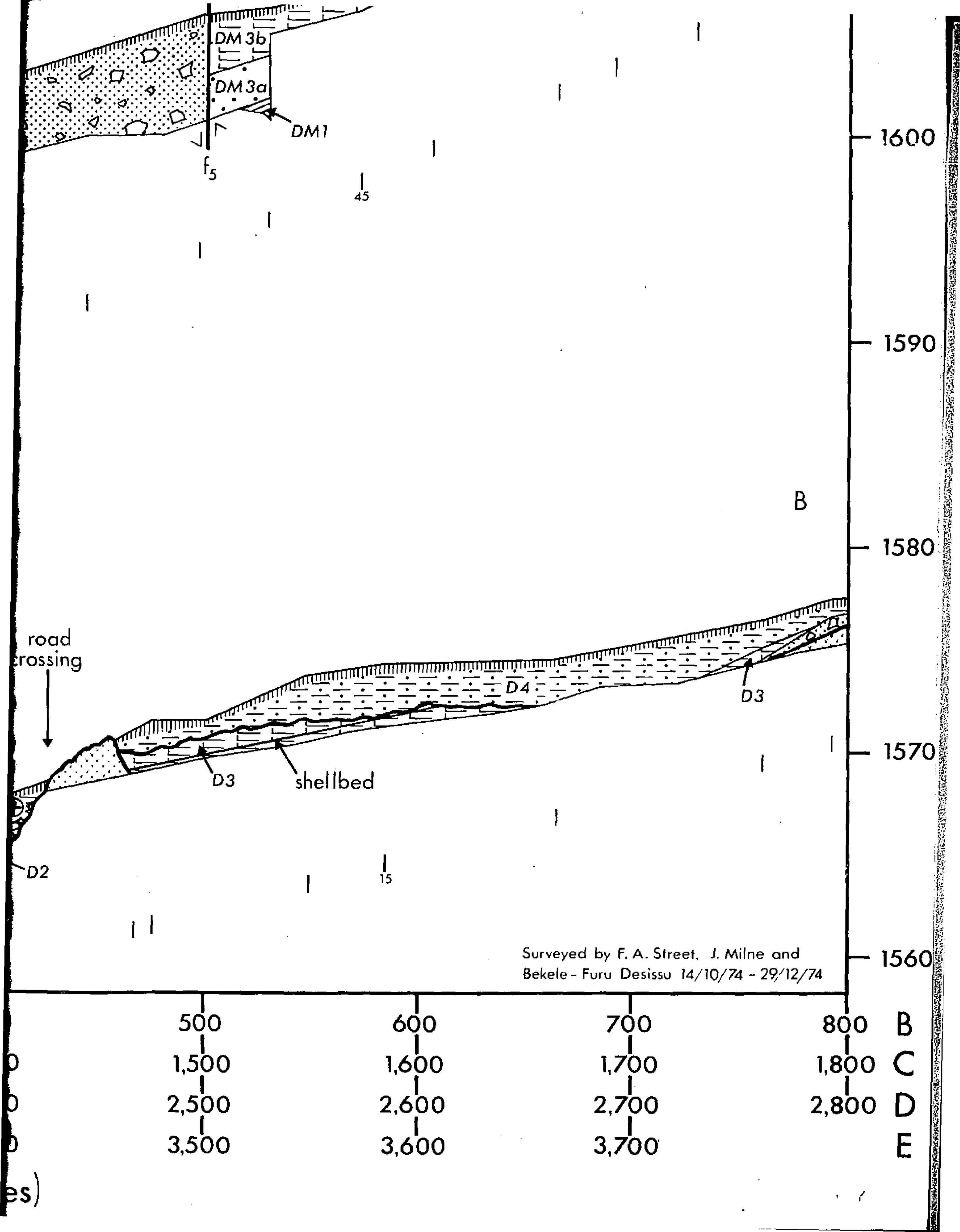
UM1
80A
80B
head
75
74B
UM4
UM2
UM3
5,610 ± 100 BP
Gak - 3387
6,150 ± 150 BP
SUA - 497
60
55
2,510 ± 100 BP
GIF - 4011
DM5
DM4
DM 3b
DM3a
DM1
DM 3a
DM6
DM3b
DM2
f1
f2
f3
f4
39
39B
40


D
9,480 ± 140 BP
SUA - 501
UM5
UM4
UM3
UM2
UM2
UM1
14,400 ± 750 BP
Gak - 3748
10,220 ± 140 BP
SUA - 500
65
C
UM5
UM4
UM2
50
f5
DM 3b
DM 3a
DM1
f5
45
1640
1630
1620
1610
1600
1590


40
30
.20
10
0
B
W
A
f1
DM 3b
DM6
DM 2
DM1
DM1
D5
D4
30
26B
25
26
20
Lake Shala 1,558m
A 0 100
B 800 900 1,000 1,100
C 1,800 1,900 2,000 2,100
D 2,800 2,900 3,000 3,100



GIF - 4011
DM5
DM 4
DM 3b
DM3a
DM1
DM 3a
DM 3b
DM6
DM3b
DM 2
DM 2
DM1
DM1
f1
f2
f3
f4
39
39B
40
37
37B
35
30
road crossing
5,530 ± 65 BP
Q - 1369
5,750 ± 70 BP
Q - 1361
D5
D3
D2
D6
D2
shellbed with fish skeletons
5
10
100
1,100
2,100
3,100
200
1,200
2,200
3,200
300
1,300
2,300
3,300
400
1,400
2,400
3,400
Distance along gully floor (metres)


DM 3b
DM 3a
DM1
f5
45
1600
1590
B
1580
road crossing
D4
D3
D3
shellbed
1570
D2
15
Surveyed by F. A. Street, J. Milne and Bekele - Furu Desissu 14/10/74 - 29/12/74
1560
500
600
700
800
B
1,500
1,600
1,700
1,800
C
2,500
2,600
2,700
2,800
D
3,500
3,600
3,700
E
es)