LATE PLIOCENE-QUATERNARY HISTORY OF THE NORTHWESTERN INDIAN OCEAN: AN ORGANIC GEOCHEMISTRY PERSPECTIVE

ALFRED NZIBAVUGA NYARUBAKULA MUZUKA, B.Sc (Geology)

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BY

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Department of Earth Sciences
Memorial University of Newfoundland.
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To my wife Mwajuma for her sacrifices and courage to withstand the hardship and loneliness during the trying life of our temporary separation.

ABSTRACT

The residue organic carbon (OC) content and stable isotopic compositions of OC and nitrogen for sediment samples from the Ocean Drilling Program (ODP) Leg 117, Sites 724 and 725, have been used to establish the history of the monsoon wind regime and its effect on coastal upwelling and primary productivity, in the northern Indian Ocean since the late Pliocene. Generally, residue OC content at Site 724 is greater than 2%; whereas at Site 725 it is less than 1%. Sediments deposited prior to the onset of the Pleistocene contain larger amounts of OC that have relatively depleted stable carbon and nitrogen isotope values. High primary productivity, reduced microbial activity as a result of the intrusion of highly saline waters from the Red Sea and the Persian Gulf, and/or enhanced preservation, produced these larger amounts of OC and relative depletion.

The amount of organic matter preserved near the Pliocene-Pleistocene boundary is low due to decreased coastal upwelling, productivity and weak southwest monsoon winds. The lower amounts of organic matter preserved during the Pleistocene-Holocene period are generally associated with more positive stable nitrogen isotope values, owing to an intensification of the oxygen minimum zone and denitrification since the early Pleistocene. A correlation of the residue OC content with oxygen isotope stages, at Site 724, indicates slightly better preservation during interglacial periods due to increased coastal upwelling, primary productivity and/or preservation. Generally, stable nitrogen isotope compositions are relatively enriched at Site 725 than at Site 724. This may be attributed to the stable oxygen minimum zone and intense denitrification at shallower depths. A correlation of the OC isotope values with oxygen isotope stages indicates an enrichment in ^{13}C during glacial periods.

Based on stable isotope ratios, terrestrial input of organic matter into the Oman Margin is minimal owing to the lack of major rivers and low productivity in the source regions.

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1. INTRODUCTION

The total amount of organic matter (OM) leaving the euphotic zone and becoming incorporated into sediments differs from location to another worldwide. This variable preservation is the result of differences in oceanic circulation, rates of alteration and diagenetic processes, water column productivity, and/or terrestrial inputs. Apart from regions of upwelling, sedimentary organic carbon (OC) content in the major ocean basins is low. Owing to high nutrient contents in upwelled waters, areas influenced by upwelling; such as the northern Indian Ocean, are associated with enhanced primary productivity and relatively high OC content in the underlying sediments. Coastal upwelling in the northwestern Indian Ocean occurs as a response to monsoon wind regimes. Coastal upwelling, productivity and fluctuations of the oxygen minimum zone (OMZ) in the northwest Indian Ocean are dependent on the intensity and direction of the monsoon winds. Intense southwest (SW) monsoon winds, which cause coastal upwelling, are associated with high primary productivity, especially off the coasts of Somalia and Arabian peninsula. The OM produced during these upwelling cycles causes a high demand for oxygen during water column diagenesis, leading to a development of the OMZ at intermediate water depths.

The samples used in this study were collected during the Ocean Drilling Program (ODP) Leg 117, in the northwestern Indian Ocean (Arabian Sea). One of the primary objectives of Leg 117 was to obtain continuous high resolution data on the onset and variability of the monsoon winds and its effect on the climate of the surrounding continents, oceanic circulation, productivity and the OMZ as recorded in the sediments deposited under different depositional environments. Furthermore, the information obtained was to be used to test the validity of the advanced hypotheses of monsoon forcing factors including changes in solar radiation and uplift of the Himalayan Mountains (see section 1.4).

In this study, an attempt is made to establish the variability of the monsoon wind regime, since the late Pliocene as recorded at two distinct water depths. A second objective is to establish the sources of OM as well as changes in the intensity of coastal upwelling, primary productivity and the fluctuation and effect of the OMZ on preservation and diagenesis of OM. The stable isotopic compositions of organic carbon and nitrogen and variations in OC content for sediments recovered from the Oman Margin are used to address these objectives.

1.1. ODP LEG 117 IN THE NORTHWESTERN INDIAN OCEAN.

Leg 117 coring commenced on the 19th of August, 1987 and ended on the 18th of October, 1987. A total of 25 holes were drilled at 12 sites (Fig. 1), which included a solitary location on the Indus Fan (Site 720), and multiple locations at the Owen Ridge (Sites 721, 722 and 731) and the Oman Margin (Sites 723 through 730). Only sediment samples from the Oman Margin, Sites 724 and 725 are used in this study. These two sites are located at a water depths of 592.8 and 311.5 respectively.

1.1.1. Northwestern Indian Ocean Setting.

The Arabian Sea is bordered by India to the east, Africa and Arabian Peninsula to the west and Iran and Pakistan to the north (Fig. 1). It is connected to the inland marginal seas of the Red Sea and the Persian Gulf by sills at the Straits of Bab el Mandeb and Hormuz, respectively. The continental margin of the Arabian Sea is wider on the eastern side (western India) than in the western side (Oman Margin). On the southeast Arabian side, the continental margin is narrow, steep and almost featureless between 18° and 22.5°N. A relatively wide continental shelf exists between Ras al Madrakah and Sharbithat (Prell et al., 1989). Tectonically, the continental margin off Oman is passive (Whitmarsh, 1979; White, 1984).

There are four major depositional basins in the northern Indian Ocean, namely: the Indus Fan, the Owen Ridge, the Owen Basin and the Oman Margin.

1.1.2. Leg 117 Drilling Results.

The samples used in this study are only from Oman Margin ODP Sites 724 and 725, therefore, drilling results for these two sites are discussed in detail. A brief summary of drilling results for Owen Ridge and other Oman Margin sites relevant to this study are also provided.

1.1.2.1. Site 724.

Site 724 is located at latitude 18°27.713'N and longitude 57°47.147'E (Fig. 1). This Site is located near the center of the upper slope basin at a water depth of 592.8m in the middle of the modern OMZ. The upper slope basin is about 5 km wide and is bounded by a fault block on the west and a basement ridge on the east. The basement is thought to be ophiolitic in nature and overlain by a syncline-shaped sediment cover which is thickest in the center of the basin.

The cores recovered from Site 724, range from early Pliocene to Holocene in age. The sediment unit recovered can be subdivided into two facies. The first facies (Facies I), which is mainly calcareous clayey silt beds is black (5Y 2.5/2), very dark grey (5Y 3/1), olive grey (5Y 4/2, 5/2) and olive (5Y 5/3) in colour. Smear slide results indicate that the average mineralogical composition is approximately 25% detrital calcite, 10% quartz, 20% nannofossils and 2% foraminifera (Prell, et al., 1989). Contacts between beds are generally burrow mottled. Only a 1 to 1.5m thick layer of finely bedded calcareous clayey silt at a depth of 16 meters below sea floor (mbsf) did not have burrow mottles. At Site 724C, facies I extends from 0 to about 170 mbsf.

The second facies (Facies II) observed is a laminated diatomaceous clayey silt. This facies is composed of laminae which are dark olive grey (5Y 3/2), olive grey (5Y 4/2), and olive (5Y 5/4). The mean mineralogical composition of laminae is approximately 46% diatom, 33% clay, 7% detrital calcite, 6% quartz and 2% nannofossils (Prell et al., 1989).

From the surface downcore to about 161m, planktonic and benthic foraminifers and calcareous nannofossils are abundant and moderately well-preserved. Below this depth, to about 225m, there is a decrease in the abundance of both planktonic foraminifera and nannofossils. The distribution and preservation of siliceous microfossils shows an inverse relationship to calcareous nannofossil abundance.

1.1.2.2. *Site 725.*

Site 725 is located at latitude 18°29.200'N, longitude 57°42.030'E and at a water depth of 311.5m (Fig. 1). This site is located on the landward edge of the upper slope basin at the transition zone of the continental shelf and continental slope. The sediments recovered range from Pleistocene to Holocene in age. Based on a visual core description, compositional smear slide results and OM and carbonate content, the sediments recovered at this site are subdivided into three lithologic units (Prell et al., 1989). Generally, the sediments are composed of a calcitic marl-nannofossil ooze, calcitic marl-calcareous ooze, interbedded clayey silt, laminated diatomaceous mud and calcitic-sand-silt or clay (Prell et al., 1989).

At Site 725C, unit I extends from 0 to 120 mbsf, and is composed of foraminifera rich calcitic sandy silt to calcitic marl-calcareous ooze, calcitic-marl-nannofossil ooze and nannofossil rich sand, silt or clay. Furthermore, this unit is slightly to moderately bioturbated. The colour of the sediments in unit I darken downcore. The colour changes from olive (5Y 4/3), olive grey (5Y 4/2) near the surface to olive (5Y 4/3), olive grey (5Y 4/2) and dark greenish grey (10Y 5/2) at intermediate levels, then to olive (5Y 4/3), olive grey (5Y 4/2), dark olive grey (5Y 4/1, 5Y 3/2) and very dark grey (5Y 3/1) lower in the section. Based on smear slide results, this unit contains 15% to 50% inorganic calcite, 5% to 20% quartz, 10% to 45% clays, 5 to 50% nannofossils and 2% to 20% foraminifers (Prell et al., 1989).

Unit II, which range from 120 to 145.4 mbsf, is composed of interbedded calcitic sand silty-clay and laminated diatomaceous beds. Based on smear slide results, diatomaceous laminae contain 10% to 30% diatoms, 2% to 5% sponge spicules, 10% to 35% nannofossils 1% to 5% foraminifers 25% to 35% clays, 15% to 25% inorganic calcite, and 3% to 20% quartz (Prell et al., 1989). The calcitic silty-clay component is essentially barren of biogenic material, and contain 20% to 50% inorganic calcite, 10% to 45% clay, 20% to 30% quartz, and 0% to 5% siliceous biogenic material (Prell et al., 1989). The light-coloured laminae contain an average of 55% diatoms, 30% clay, 3% inorganic calcite, and 2% to 5% each of quartz, nannofossils, and foraminifers (Prell et al., 1989).

Unit III, which range from 145.4 to 162.8 mbsf, is similar to unit I, and composed of a calcitic marly nannofossil ooze as a dominant component, and a nannofossil rich calcitic sand silt-clay as a minor component. Based on smear slide results, mineralogical composition of the marly nannofossil oozes component is 25% to 45% nannofossils, 25% to 35% inorganic calcite, 20% to 25% clay, 5% to 11% quartz, and 2% to 5% each of foraminifers and accessory minerals (Prell et al., 1989). The sand silt-clay component contain an average of 25% quartz, 25% clay, 35% inorganic calcite, 10% biogenic carbonate, and 5% accessory minerals (Prell et al., 1989). This unit is slightly bioturbated.

1.1.3. Other Sites

1.1.3.1 Owen Ridge Sites

The sections recovered from the Owen Ridge Sites (721, 722, and 731) range in age from the Oligocene to the Holocene. The cores recovered from the Owen Ridge sites have been divided into four major lithologic units.

Unit I is mainly composed of alternating light and dark layers of foraminifer-bearing, foraminifer-nannofossil ooze and nannofossil ooze of Holocene to late Miocene age. Changes in colour, bulk density, carbonate content and magnetic susceptibility of the sediments in this unit are cyclic. At the boundary between lithologic Unit I and Unit II on Site 721, planktonic foraminifera, characteristic of upwelling, such as Globigerina bulloides, Globorotalia menardii and G. tumida and the Neogloboquadrina lineage of N. acistaensis and N. dutertrei, were observed.

Unit II ranges from late to middle Miocene and is composed of late Miocene foraminifer-nannofossil chalk and ooze that grades downsection to a siliceous nannofossil chalk ooze of middle Miocene age. The abundance of biogenic opal in this unit is high owing to better preservation of both diatoms and radiolarians. Microfossil assemblages generally are characterized by the lack of typical tropical species and the dominance of cool-water assemblages.

Unit III is composed of a nannofossil/chalk of early to middle Miocene age. Preservation of siliceous components is poor in carbonate rich chalks (60%-90% $CaCO_3$). However, preservation of silica increases along with upwelling fauna of planktonic foraminifera in the uppermost part of the middle Miocene.

Unit IV is of late Oligocene(?) to middle Miocene age and is characterized by alternating coarse-grained sand and silt turbidites and mud turbidites with no overall large scale fining upward trend. Preservation of both planktonic calcareous and opaline tests is poor and benthic foraminifera from upper and middle bathyal depths are absent.

1.1.3.2. Oman Margin Sites

The sediments recovered from the Oman Margin sites range from Eocene(?) to Holocene in age. Lithologically, the middle Miocene to Holocene sediments, which have been influenced by upwelling are dominated by green to olive calcareous oozes to clayey silts. These upwelling-influenced sediments are underlain by shallow water foraminifera limestones of Eocene(?) age. Laminated opal-rich facies, which are confined to the late Pliocene and early Pleistocene were observed at Sites 723 and 724 located at the center of modern OMZ and at Sites 725 and 726 located at shallower depths. Microfossil assemblages change from being more siliceous in the Miocene and Pliocene to more calcareous in the Pleistocene. Other features observed include phosphorite-rich lag deposits at Site 726 and extensive microbial sulphate reduction at Sites 723, 724 and 725 fuelled by the additional supply of sulphate from subsurface sources (Prell et al., 1989). The observed sulphate reduction is beyond the depth range usually found in deep-sea sediments (Prell et al., 1989).

1.2. COASTAL UPWELLING

Coastal upwelling is an oceanic surface circulation pattern which occurs along some continental margins. It occurs as a result of the Ekman transport process, which carries water away from the coast in the northern or southern hemisphere when wind direction is to the left or right of the respective coast. Upwelling creates the upward movement of cold water from intermediate depths into the photic zone, which are often rich in nutrients such as nitrate, phosphate and silica.

Upwelling zones are characterized by low sea surface temperature (SST), low salinity, low oxygen content, high primary productivity and assemblages of cold water species (Diester-Haass, 1978; Baturin, 1983; Labracherie et al., 1983). Lowered SST and high primary production in the euphotic zone are the main features that produces the characteristics of sediments underlying upwelling regions and differentiate them from adjacent regions not affected by upwelling. Sediments underlying upwelling zones contain high OM, high biogenic silica, phosphate, abundant cold water foraminifera and diatom species such as <u>Globigerina bulloides</u> and <u>Neogloboquadrina dutertrei</u>, and varves sometimes. Moreover, high productivity of OM in the surface waters is reflected in the stable isotopic compositions of carbon, nitrogen and oxygen and sedimentary enrichments in barium, minor metals (Cu, Zn, Ni) and uranium (Diester-Haass, 1978; Baturin, 1983). Thus, these signals, when preserved in the sediments, can be used to infer variations in coastal upwelling.

1.2.1. Coastal upwelling in the Indian Ocean.

Unlike the other four major coastal upwelling centers, which are located on eastern boundary currents (off the western coasts of the continents), coastal upwelling in the Indian Ocean occurs along the western boundary current (Somali Boundary Current). Seasonal coastal upwelling off the coast of Somalia, southeast Arabia and off the west coast of India occur in response to the monsoonal wind regime (see section 1.4.0).

Intense winds blow nearly parallel to the coasts of Somalia and southeast Arabia during the SW monsoon. Since these northern hemisphere coasts are to the left of the wind direction, surface waters from the coast are transported offshore and are replaced by deep, cool, nutrient-rich waters (Prell and Streeter, 1982; Prell et al., 1989). During the NE monsoon, the Ekman process transports offshore surface waters toward the coast (downwelling) leading to a collapse of upwelling along the continental margin (Prell and Streeter, 1982; Zahn and Pedersen, in press).

Upwelling off the Somalia coast is intense between 5°N and 11°N and is located within the Somali Boundary Current. Sea surface temperatures as low as 14°C have been recorded during a period of intense coastal upwelling (Prell and Streeter, 1982). Concentrations of nutrients, such as PO_4^{-3} and NO_3^- , are more than $10 \mu g$-at L^{-1} during such periods (Wyrtki, 1973).

In the Arabian Sea, upwelling occurs east of 55°E (Wyrtki, 1973). According to Currie et al., (1973) and Swallow, (1984), upwelling extends at least 400 km offshore and parallels the coast for over 1000 km. However, as noted by Currie et al., (1973), there is substantial spatial variability in the upwelling along the Arabian coast. Similar conditions have been reported in the Peru upwelling region by Reimers and Suess, (1983). Intense upwelling occurs in the vicinity of Muria Island and Ras Madrakah. The concentrations of nutrients off the southeast Arabian coast are higher than in the Somali upwelling area and a large area is affected by high concentrations of both PO_4^{-3} and NO_3^- (Wyrtki, 1973). The coldest SST occurs between 17°N and 22°N and temperatures as low as 18°C have been recorded (Prell and Streeter, 1982).

The intensity of SW monsoon winds has changed since its establishment in the middle Miocene and coastal upwelling has changed accordingly. Variations in the intensity of upwelling may be inferred from changes in upwelling characteristics (see section 1.2.1). In the Indian Ocean, variations in content of OC, biogenic silica, phosphate and Globigerina bulloides have been used to infer the intensity of coastal upwelling during glacial and interglacial periods (Labracherie et al., 1983; Prell, 1984; Fontugne and Duplessy, 1986; Socci, 1986). All of these studies indicate more intense coastal upwelling during interglacial periods when SW monsoon winds are more intense, than during glacial periods in which NE monsoon winds are intense. During glacial periods, the Somali and Arabian coasts are to the right of the strongest monsoon wind direction. Glacial periods are probably associated with downwelling along these coasts.

1.3. NORTHWESTERN INDIAN OCEAN CIRCULATION

Primary productivity and the preservation of OM in the oceans are controlled by the prevailing physio-chemical conditions. The physio-chemical factors includes variations in nutrients, light, temperature, salinity, water density, dissolved carbon dioxide and oxygen, current strength and pH

(B'e and Hutson, 1977). Variations in these physio-chemical factors in the oceans are mainly governed by circulation. Oceanic circulation in the northwestern Indian ocean can be broadly divided into three groups, namely: a monsoon gyre, North Indian Intermediate Waters and Indian Deep Water (Wyrtki, 1973; Prell et al., 1989).

1.3.1. Monsoon Gyre

The monsoon gyre extends from the north in the Arabian Sea to about 10°S where it is separated from the Southern Subtropical Gyre by a hydrochemical front (Wyrtki, 1973; Wright and Thunell, 1988). As the term indicates, monsoonal surface circulation is in response to seasonally changing monsoon winds (see section 1.4.0). The surface current circulation north of the equator is composed of the NE and SW monsoon currents, respectively.

The present NE monsoon current is formed during the NE monsoon and flows from east to west (Fig. 2a; Wyrtki, 1973; Wright and Thunell, 1988). North of the equator, the NE monsoon current is composed of a southward flow off the coast of Somalia and a westward flowing North Equatorial Current (NEC) (Fig. 2a; Wyrtki, 1973; Wright and Thunell, 1988). The NEC is not strong enough to disintegrate the thermocline and, therefore, it is not associated with any marked coastal upwelling.

The NE monsoon current collapses in April and is succeeded by a strong SW monsoon current. During this period, water along the Somalia coast starts to flow northward. The net direction of the SW monsoon current is west to east and reaches its maximum strength in July (Fig. 2b; Wyrtki, 1973; Wright and Thunell, 1988). Although the net current direction is west to east, the strong Somali Boundary Current and expanded SW monsoon current, flow northward prior to July (Wyrtki, 1973; Wright and Thunell, 1988).

Another type of surface circulation in the Indian Ocean north of the equator related to the monsoons is coastal upwelling (see section 1.2.2).

1.3.2. North Indian Intermediate Waters

North Indian intermediate waters (NIIW) in the Arabian Sea originates from the Red Sea, Persian Gulf and within the Arabian Sea. It extends from 200 to more than 1200m water depth and is characterized by very low oxygen content ($< 1.0ml\ L^{-1}$) north of 3^oN. These inland seas have a Mediterranean type of circulation where emission of saline water through the bottom of the strait takes place and is compensated by introduction of fresher Indian Ocean waters at the surface (Bowen and Jux 1987).

Annual evaporation in the central and northern Arabian Sea is higher than precipitation, leading to the formation of highly saline intermediate waters. During the NE monsoon this water spreads southwest, while it spreads to the SE during the SW monsoon (Wyrtki, 1973).

Oxygen-poor and highly saline waters from the Red Sea, which are rich in nitrate, enter the Arabian Sea via the Gulf of Aden over a sill of about 140m depth at the Strait of Bab-el Mandeb. Large amounts of subsurface Red Sea waters are transported to the Indian Ocean during the winter (Bethoux, 1988; Souvermezoglou et al., 1989). Maximum salinities in the Gulf of Aden induced by influx of highly saline Red Sea waters, occur at a depth of about 800m (Wyrtki, 1973).

The waters from the Persian Gulf, which are characterized by high salinities, high temperatures and very low oxygen contents, enter the Gulf of Oman across a sill at about 100m in the Strait of Hormuz. According to Qasim (1982), the Persian Gulf waters spread as a tongue southward in a region west of 60°E. A strong salinity maximum, in the Gulf of Oman caused by the inflow of Persian Gulf waters occurs at a depth of about 300m (Wyrtki, 1973).

1.3.3. The Indian Deep Water

Indian deep water (IDW) in the Arabian basin is a mixture of Antarctic Bottom Water (AABW), North Atlantic Bottom Water (NABW) and Circumpolar Deep Water (CDW) (Shipboard Scientific Party, 1989). This water mass enters the Arabian Sea basin over a sill at a water depth of about 3800 to 4000m on the Calsberg Ridge in the Owen Fracture Zone region (Kolla et al., 1976). However, IDW has no effect on the preservation and deposition of sediments at Sites 724 and 725.

1.4. MONSOON WIND REGIME

A large part of the Indian Ocean and the surrounding continents are under the influence of monsoon winds which seasonally reverse directions. The monsoon wind regime alternates between the weaker NE monsoon and stronger SW monsoon. The NE monsoon winds blow from a northeasterly direction from November through April. The SW monsoon winds blow from a southwesterly direction from May through October.

Based on the first appearance of Globigerina bulloides, a key upwelling species, monsoon wind regime is inferred to have been established in the middle Miocene (Leclaire, 1974; Shipboard Scientific Party, 1989). Monsoon wind patterns and its intensity are the result of the altitude of the Tibetan plateau and the Himalayan mountains, variations in solar radiation and surface albedo (Hahn and Manabe, 1975; Prell, 1984; Webster, 1987; Barnett et al., 1989). It is thought that changes in the intensity of monsoon winds are in phase with Milankovitch cycles (Rossignol-Strick, 1983; Prell, 1984).

Differential heating of the land and sea, caused by differences in heat capacities, cause landmasses (having low heat capacity) to be heated faster than the ocean, by solar radiation, during the summer. Less dense air masses over the land rise, forming a low pressure zone over the Tibetan plateau and the Himalayan mountains (Fig. 3) and a high pressure zone in the southern Indian Ocean. Pressure differences cause the movement of air masses from high to low pressure zones. Because of the rotation of the Earth, the air mass is deflected by the Coriolis force to the right/left in the northern/southern hemispheres causing summer monsoon winds to blow from a southwestly direction. During the winter, land loses heat faster than the ocean and air over the ocean rises forming a low pressure zone in the ocean and a high pressure zone in the Tibetan plateau and the Himalayan mountains (Fig. 3). As a result, dry winds blow from the continent toward the ocean in the northeastly direction.

The intensity of monsoon winds and monsoonal coastal upwelling have changed with time since its establishment in the middle Miocene. Monsoonal variations in the Indian Ocean and the surrounding continents have been inferred from a variety of geological data and general circulation simulation models. Geological data include stable oxygen and carbon isotopes of planktonic foraminifera and OM (Duplessy, 1982;

Van Campo et al., 1982; Fontugne and Duplessy, 1986), distribution of planktonic fauna (Prell and Curry, 1981; Cullen and Prell, 1984; Prell, 1984; Prell and Van Campo, 1986), pollen spectral (Van Campo et al, 1982; Rossignol-Strick, 1983; Prell and Van Campo, 1986), and wind transported materials (Kolla and Biscaye, 1977). Circulation models include simulation of the climate for 9,000 yr B.P. (Kutzbach, 1981; Kutzbach and Guetter, 1986), for the Last Glacial Maximum at 18,000 yr B.P. (Gates, 1976; Manabe and Hahn, 1977; Hansen et al., 1984), and for 150,000 yr B.P. (Prell and Kutzbach, 1987).

The above mentioned studies indicate that the SW monsoon winds were weaker during glacial periods and stronger during interglacial periods. During the glacial times, NE monsoon winds were very strong and probably produced upwelling and/or greater surface mixing in the northeastern Arabian Sea.

1.5. PRODUCTIVITY

As mentioned above (section 1.3.), primary productivity in an area is governed by both physio-chemical and biological factors. Sunlight and the nutrient supply of ammonia, nitrate and silica, are considered to be the major factors controlling productivity. Water in the euphotic zone is generally depleted in nutrients owing to biological consumption during primary production. Nutrients transported to the intermediate water depths, in the form of remineralized carbon and nitrogen from dead organisms or fecal pellets, are not utilized by phytoplankton because of lack of sunlight and are brought to the surface during upwelling. The replenishment of nutrients in the euphotic zone can be sourced in water of intermediate depths, river runoff and regeneration of primary productivity.

1.5.1. Productivity in the Arabian Sea

Like the other four areas in the world oceans, which are influenced by coastal upwelling, the Arabian Sea is a center of high primary productivity. According to Romankevich (1984), productivity in this area (Arabian Sea) may be as much as five times that commonly observed in the other world oceans. On average, more than 500 mg $C/m^2/day$ are produced by primary producer in the euphotic zone from May to October (Krey and Babenerd, 1976, in Shipboard Scientific Party, 1989). Productivity drops to values as low as 150 mg $C/m^2/day$ from November through April (Krey and Babenerd, 1976, in Shipboard Scientific Party, 1989). Replenishment of nutrients in this area comes primarily from intermediate water depths below the thermocline. In comparison to upwelling, riverine contributions of nutrients are insignificant in this area owing to the absence of major rivers in the area.

Enhanced production of OM in this area leads to a high demand for oxygen during microbial remineralization. High productivity coupled with a lack of horizontal advection, due to the semi-enclosed nature of the region, and strong density gradients, which hinder exchange of oxygen between the photic zone and water below the thermocline, results in the development of an OMZ in water depths between 200 and 1200m (Wyrtki, 1973; Qasim, 1982; Sen Gupta and Naqvi, 1984; Slater and Kroopnick, 1984).

1.5.2. Paleoproductivity in the Northwestern Indian Ocean

Although only a small portion of OM settling to the sea floor becomes fossilized, appreciable amounts of OM can be found in areas of high primary production. In regions of high primary productivity, OC content in the underlying sediments is distinctly higher than in surrounding regions. Thus, paleoproductivity in these regions may be inferred from variations in OC content. However, care must be taken in interpreting OC content, because there are many factors which may affect the OC content in sediments. High OC content can be found in areas of high or low primary productivity. High OC content in sediments can result from the influx of large amounts of terrestrial OM which are more resistant to microbial degradation than planktonic debris (Toth and Lerman, 1977; Tissot and Welte, 1978; Billen, 1982; Glenn and Arthur, 1985); enhanced production in the photic zone;

enhanced preservation due to high sedimentation rate or anoxic conditions (Stein, 1986a, b); and intrusion of highly saline waters (Klinkhammer and Lambert 1989).

Various workers have used total OC, biogenic silica and other productivity indicators preserved in both oxic and anoxic marine environment to infer paleoproductivity in the world's oceans (Muller and Suess, 1979; Muller et al., 1983; Labracherie et al., 1983; Pedersen, 1983; Glenn and Arthur, 1985; Fontugne and Duplessy, 1986; Stein, 1986a; Sarnthein et al., 1988; Macko, 1989). In the northwestern Indian Ocean, OC and silica content, stable isotope composition of OM and total phosphate have been used to infer productivity during the last glacial maximum (LGM) and interglacial periods (Labracherie et al., 1983; Fontugne and Duplessy, 1986; Socci, 1986). During the LGM, productivity was low, owing to a weak SW monsoon and sluggish coastal upwelling and high during the interglacial period, when the SW monsoon and coastal upwelling in the region were intense (Duplessy, 1982; Labracherie et al., 1983; Fontugne and Duplessy, 1986; Prell and Van Campo, 1986; Socci, 1986; Prell and Kutzbach, 1987).

1.6. PRESERVATION OF ORGANIC MATTER

The use of OC content to estimate paleoproductivity relies on the premise that OC content in sediments reflects productivity in the euphotic zone. In order to be a "mirror image" of surface productivity, organic material must be preserved. Although high productivity in upwelling regions can create a rapid flux of OM to the sea floor, only a small portion is fossilized. A major part of OM is recycled during microbial oxidation and reduction in the water column and after deposition. Generally, the amount of OM found in the sediments will depend on the source (whether it is terrestrial or marine), the rate of primary production in the water column and intrusion of saline water in a depositional basin. Other factors include the dissolved oxygen content, the availability of other electron acceptors (NO_3, MnO_2, SO_4, $FeO(OH)$), the water current strength, the grain size and the sedimentation rate (Muller and Suess, 1979; Emerson and Hedges, 1988; Mayer et al., 1988).

The amount of dissolved oxygen in the water column is considered to be a primary factor controlling preservation of organic materials (Emerson, 1985; Emerson and Hedges, 1988). When the dissolved oxygen concentration at intermediate water depth is low, a common phenomena in highly productive areas, sulphate reduction and denitrification processes, usually considered to be secondary, become dominant (Sen Gupta and Naqvi, 1984; Emerson, 1985; Canfield, 1989). Under such conditions, nitrate ions become an immediate, thermodynamically favoured, source of free energy for the oxidation of OM (Cline and Kaplan, 1975; Deuser et al., 1978; Sapozhnikov and Sviridova, 1979; Sen Gupta and Naqvi, 1984).

For areas in the ocean where sedimentation rates are low, sulphate reduction is less important. But in continental margins where sedimentation may be greater than 100cm per thousand years, the effect of sulphate reduction and oxygen content on the degradation of OM are equally important (Canfield, 1989). Generally, the effect of sulphate reduction will depend on the availability of sulphate and reactivity (lability) of the OM.

1.7. STABLE ISOTOPES OF CARBON AND NITROGEN

The stable isotope compositions of carbon and nitrogen in natural materials are highly variable. This variability primarily results from biological fractionation of isotopes during metabolism/catabolism and variations in the starting isotopic compositions of carbon and nitrogen compounds used during the synthesis of OM and diagenetic processes.

The fractionation, or partial separation of isotopes, during physical or chemical processes is caused by kinetic effects, thermodynamic (chemical equilibrium) effects and/or nuclear spin (Galimov, 1985; Faure, 1986; Hoefs, 1987). Fractionation caused by kinetic isotope effects are due to differences in the reaction rates of the individual isotopes of an element and is generally associated with biological processes. Since smaller masses have greater velocities and weaker chemical bonds, the products are enriched in light isotopes and residues are enriched in heavy isotopes. Kinetic effects occur during photosynthesis, nitrogen fixation, denitrification, diffusion and deamination (Hoef, 1987). The thermodynamic effect is associated with both biological and abiological processes. It occurs because of differences in the properties of the isotopes of an element, such as bond strength and minimum free energy. Both kinetic and thermodynamic effects are temperature dependent.

1.7.1. Nitrogen

Nitrogen has two stable isotopes, ^{14}N and ^{15}N, and their relative abundance in the atmosphere as determined by Neir (1950) are 99.64% and 0.36% respectively. Nitrogen occurs in various media (atmosphere, hydrosphere, biosphere, lithosphere) but atmospheric nitrogen is the principal reservoir of nitrogen. In aqueous environment, nitrogen occurs as dissolved molecular nitrogen, NO_3^{-2} , NH_4^+, N_2O, NO_2^-, NO_2 and organic nitrogen.

During the synthesis of OM, terrestrial plants utilize fixed atmospheric nitrogen, which has a $\delta^{15}N$ value similar to the atmospheric value of $0^{o}/_{oo}$. In aquatic environments, phytoplankton mainly utilize dissolved inorganic nitrogen in the form of nitrate during the synthesis of OM, but other forms of nitrogen; such as dissolved molecular nitrogen, ammonium and nitrite; are also used (Saino and Hattori, 1985). The uptake of the nitrogen species by aquatic plants depends on the isotopic composition of the available nitrogen species and the intensity of light (Saino and Hattori, 1985). Because of differences in the stable isotope composition of nitrogen used during OM synthesis, the isotope composition of terrestrial organic materials are usually different from those of a marine origin.

The stable isotope composition of nitrogen in natural materials is highly variable and δ^{15}N values ranges from $-20^{\circ}/_{oo}$ to $+30^{\circ}/_{oo}$ (Kaplan, 1983). This high variability results from the isotopic fractionation during nitrogen fixation and assimilation (synthesis of OM) and biogeochemical transformations of nitrogen in the water column, at the sediment-water interface and within the sediments (Altabet and McCarthy, 1985; Saino and Hattori, 1985; Macko et al., 1987; Libes and Deuser, 1988). Because nitrogen is more rapidly lost than carbon during degradation of particulate organic matter (POM), greater variability of ^{15}N in suspended POM is expected than in ^{13}C.

An increase in δ^{15}N values of POM with depth has been observed (Altabet and McCarthy, 1985; Saino and Hattori, 1980,1985; Sigleo and Macko, 1985; Altabet, 1988). The observed variations have been attributed to bacterial degradation of small OM particles and fractionation associated with disaggregation of sinking particles (Altabet and McCarthy, 1985; Saino and Hattori, 1980, 1985; Sigleo and Macko, 1985; Altabet, 1988). The stable nitrogen isotope composition of sediments and POM do not co-vary (Libes and Deuser, 1988). This may be in part due to the fact that rapidly sinking particles, which make up most of the mass flux to the sea floor, are less enriched in ^{15}N than small suspended particles. Because of this, one expects to find low $\delta^{15}N$ values in areas of high sedimentation related to high water column productivity. This simple relationship is complicated by the diagenetic effects at the sediment-water interface, the concentration and recycling of nutrients, and other environmental factors.

In regions where there is a high supply of nutrients and high productivity, oceanic nitrate may be fully utilized. When nitrate is fully utilized, the nitrogen stable isotope is less fractionated (Macko et al.,1987). This may cause underlying sediments to be more enriched in ^{15}N. When only a part of the dissolved nitrate is used by organisms, OM formed may be more depleted in ^{15}N owing to fractionation (Macko et al., 1987). Areas where there are high $\delta^{15}N$ values may indicate high productivity. However, very high $\delta^{15}N$ values may be caused by recycling, denitrification, oxidative degradation of OM and diagenesis of disaggregated particles

(Cline and Kaplan, 1975; Saino and Hattori, 1980; Altabet and McCarthy, 1985; Altabet, 1988; Liu and Kaplan, 1989). The denitrification process is associated with a large nitrogen isotope fractionation. The fractionation factor associated with denitrification as determined by Delwiche and Steyn (1970) is at least 1.02. According to Seitzinger (1988), factors controlling denitrification include temperature, the supply of nitrate and OM, and oxygen concentration. The denitrification process occurs in sub-oxic to anoxic environments where oxygen levels are low. Molecular nitrogen formed as a result of denitrification is depleted in ^{15}N, leaving the residue nitrate enriched in ^{15}N. During upwelling, this, isotopically heavy nitrate is brought to the euphotic zone, where it may be fully assimilated by phytoplankton, resulting in the formation of isotopically heavy POM (Macko, 1989).

1.7.2. Carbon

Carbon has two stable isotopes, ^{12}C and ^{13}C and their relative abundance is approximately 98.89% and 1.11% respectively (Hoefs, 1987). In the hydrosphere, carbon occurs as dissolved carbon dioxide and HCO_3^-. The isotopic composition of plant organic carbon is variable and is controlled by metabolic pathways followed during photosynthesis, concentration and isotopic composition of

CO_2 or other carbon sources used in food synthesis and other environmental factors. Values of $\delta^{13}C$ in recent sediments vary from -10 to -30$^\circ/_{oo}$ with a maximum between -20 and -27$^\circ/_{oo}$ (Degens, 1969; Deines, 1980). Variability of the $\delta^{13}C$ of terrestrial organic material in the sediments is greater than that of marine origin and covers the range of values for marine sediments. Apart from this overlap, there is a difference in isotopic compositions between terrestrial higher plants and marine plankton in lower latitudes. This is due to differences in the type of CO_2 used during photosynthesis. Terrestrial plants utilize atmospheric carbon dioxide while aquatic plants use dissolved CO_2 and $HCO3^-$. The equilibrium isotope fractionation associated with dissolution of CO_2 in water is -0.91$^\circ/_{oo}$ with ^{13}C concentrating in the dissolved phase (O'Leary, 1981). The atmospheric CO_2 has a $\delta^{13}C$ value between -6.4 and -7.0$^\circ/_{oo}$ in areas where industrial input is low (Keeling, 1961). Oceanic HCO_3^- has a $\delta^{13}C$ value of approximately 0$^\circ/_{oo}$.

The degree of carbon isotope fractionation by green plants varies from one plant type to another, depending on the metabolic pathway followed by the plant during photosynthesis. Three groups of plants can be distinguished based on the isotopic composition of plants, namely: Calvin-Benson or non-Kranz (C_3), Hatch-Slack or Kranz (C_4) and Crassulacean Acid Metabolism (CAM) (Bender, 1968; Smith and

Epstein, 1971). The C_3 metabolic pathway is associated with a maximum carbon isotope fractionation, while C_4 metabolic pathway causes a small fractionation. The C_4 plants have $\delta^{13}C$ values which range from about $-9^{o}/_{oo}$ to $-16^{o}/_{oo}$ (Deines, 1980). Terrestrial higher plants following the C_3 metabolic pathway are more negative than C_3 aquatic plants. The $\delta^{13}C$ values of terrestrial higher plants range from $-23^{o}/_{oo}$ to $-33^{o}/_{oo}$, averaging about $-26^{o}/_{oo}$, while that of phytoplankton, which are the major producer of OM in the oceans, range from $-18^{o}/_{oo}$ to $-24^{o}/_{oo}$ with an average of $-21^{o}/_{oo}$ for temperate areas (Anderson and Arthur, 1983; Gearing et al., 1984). Carbon isotopic enrichments in aquatic plants have been attributed to the slow diffusion of CO_2 in water, which is associated with a lower fractionation than in the atmosphere (O'Leary, 1981; Popp et al., 1989). Higher plants use the C_3 photosynthetic pathway, whereas most tropical grasses and some marsh plants follow the C_4 metabolic pathway.

High concentrations of CO_2 cause discrimination between ^{12}C and ^{13}C by plant cells to be high, thus causing large fractionation. Conversely, when CO_2 concentrations are low, plant cells use available CO_2 for growth regardless of isotopic composition, causing less fractionation (O'Leary, 1981).

During photosynthesis, the lightest carbon isotope (^{12}C) is preferentially fixed by organisms and is released during biodegradation. The CO_2 formed during OM degradation redissolves and is brought to the surface by upwelling processes where it mixes with oceanic dissolved carbon dioxide. In upwelling regions, the rate of mixing is high and large volumes of water are exposed to atmospheric-sea water interactions, leading to more dissolved CO_2 in the water column. Organisms utilizing regenerated over oceanic dissolved CO_2 will be relatively depleted in ^{13}C. It would be predicted that OM in upwelling regions will have lower $\delta^{13}C$ values than other areas. In general however, carbon isotopic compositions of OM in an area dominated by marine productivity are determined by environmental factors such as temperature, availability of CO_2, growth rate and species distribution (Sackett et al., 1965; Degens, 1969; Gearing et al., 1977; Fontugne and Duplessy, 1978; 1981).

In areas where there are high discharges of river water, $\delta^{13}C$ values can vary as a result of mixing of freshwater, having $\delta^{13}C$ of dissolved inorganic carbon (DIC) ranging between -5 and -10‰, with the approximately 0‰ DIC of ocean water (Degens, 1969; Sherr, 1982; Tan and Strain, 1979a, b; 1983). DIC with more negative $\delta^{13}C$ values causes marine plants to have more negative $\delta^{13}C$ values. Estimation of terrestrial contribution in areas off major rivers using mixing equation may overestimate its contribution. However, fresh water dilution has little influence in the western Arabian Sea.

1.8. SOURCE INDICATORS OF ORGANIC MATTER IN THE OCEAN

The stable isotope compositions of organic carbon and nitrogen have been used as a tool to delineate sources of OM in marine sediments and to infer paleoproductivity and glacial-interglacial changes in geological time (Fontugne and Duplessy, 1986; Macko, 1989; Parker et al., 1972; Newman et al., 1973). However, its use relies on the premise that there is a relatively constant isotopic composition of organic carbon and nitrogen produced by each source, no change in isotope ratio during decomposition and degradation into detritus, and small but constant isotopic fractionation between consumer and source (Gearing et al., 1984).

The OM in sediments is a remnant of organic degradation and diagenetic changes, which commence after the death of an organism. Furthermore, it is a complex mixture of original and altered organic materials and their decomposition products (carbohydrate, protein, lipids, lignin, pigments of various types, humic material and kerogen). Isotopically, proteins and carbohydrates are more enriched in ^{13}C than lipids, lignin and cellulose (Degens, 1969) and comprise more than 60% of the total composition of marine planktonic OM. The average carbon and nitrogen isotopic compositions of OM in the sediments will therefore depend on which group of compounds have been preferentially eliminated during diagenesis and bacterial degradation, and whether decarboxylation reactions have taken place. Decarboxylation reactions remove ^{13}C enriched groups from the OM leading to ^{13}C depletion in the residue. An increase of several per mil downcore for the upper 20-30cm of sediment, as a result of degradation and diagenesis, has been observed (Behrens and Frishman, 1971; Macko, 1981). This phenomenon has been attributed to a loss of isotopically depleted lipids, leaving a residue which is enriched in ^{13}C.

The OM in marine sediments may be regarded as a mixture of various sources including terrestrial OM, anthropogenic wastes and autochthonous materials. The allochthonous materials are transported to the ocean by rivers, wind, and in high latitudes, by ice. Rivers, and to a lesser extent wind and ice, are major contributors of terrestrial materials. Most riverine materials are deposited in estuaries, lagoons, deltas and the continental shelf and are redistributed into the deep basins by turbidity currents. Although over 80% of atmospheric particulate OM over remote areas of the oceans are terrestrially derived (Chesselet et al., 1981) the rate of deposition of clastic and OM carried by winds is very small. Thus, the amount of terrestrially derived materials decreases with increasing distance from the land (Sackett and Thompson, 1963; Schultz and Calder, 1976; Sherr, 1982).

1.8.1. Carbon and Nitrogen Stable Isotopes

In determining the sources of OM in marine environments, stable isotopes of carbon and nitrogen have been used either separately or in combination. This is because of the fact that the isotopic composition of organic carbon and nitrogen of land plants are normally different from that of marine plants. (section 1.7.1 and 1.7.2).

Some workers who have used carbon isotopes as source indicators for OM in marine environment, include Sackett and Thompson (1963), Sackett (1964), Sackett and Rankin (1970), Schultz and Calder (1976), Hedges and Parker (1976), Gearing et al. (1977), Joyce et al. (1985) and Jasper and Gagosian (1989) in the Gulf of Mexico; and Hunt (1970), Burnett and Schaeffer (1980), Sherr (1982), Brinson and Matson (1983), Gearing et al. (1984) in U.S. Atlantic estuaries. Other workers who have used carbon isotopes as source indicators include Pocklington (1976), Tan and Strain (1979a,b; 1983) in the St. Lawrence estuary (Canada); Nissenbaum and Kaplan (1972) in the area off southern California; Hedges et al. (1988) in Dabob Bay and Fontugne and Duplessy (1986) in the northern Indian Ocean.

The nitrogen isotopic composition or both $\delta^{13}C$ and $\delta^{15}N$

have not been widely used. Some of the workers who have used $\delta^{15}N$ as source indicator of OM in marine sediments include Sweeney and Kaplan (1980) in the Santa Barbara Basin California; and Mariotti et al. (1984) in the Scheldt estuary. Both $\delta^{13}C$ and $\delta^{15}N$ have been used by Peters et al. (1978), Sweeney et al. (1980), Kaplan (1983) on the San Pedro shelf; Macko (1981, 1983) on the Gulf of Mexico and Atlantic coast of the U.S.; Wada et al. (1987) along the Otsuch watershed; Cifuentes et al. (1988) in Delaware Estuary; and Mayer et al. (1988) in the Gulf of Maine.

1.9. CLIMATIC VARIATION

Many of the factors causing isotope fractionation are temperature dependent. Thus, it is expected that the isotopic compositions of phytoplankton, which are major producers of OM in the oceans, will be temperature dependent. In high latitudes, low $\delta^{13}C$ values, up to $-30^{o}/_{oo}$, have been observed for plankton (Sackett et al., 1965; Sackett et al.,1974; Fontugne and Duplessy, 1981; Sackett, 1986; Rau et al., 1989) and have been partially attributed to temperature effect. In other studies no correlation between temperature and isotopic composition was observed (Calder and Parker, 1973; Fontugne and Duplessy, 1978; Gearing et al., 1984). Carbon isotopic variations may also be associated with variations in phytoplankton populations (Sackett et al., 1974). Temperature dependence can be masked by other environmental factors, such as terrestrial input of OM, diagenetic effects, concentration of dissolved carbon dioxide and nitrogen, rate of consumption, and growth rate of cells (Galimov, 1985). For example, the solubility of carbon dioxide and nitrogen increases with decreasing temperature. This implies that in high latitude regions, there is more dissolved carbon dioxide and nitrogen than in the warm tropical waters (Rau et al., 1989). Phytoplankton in polar regions will use a smaller fraction of the dissolved source materials when compared to those in tropical waters, causing high fractionation. Since phytoplankton utilize mainly NO_3^- during the synthesis of OM, concentrations of dissolved molecular nitrogen will not cause major fractionation of the nitrogen isotopes.

Apart from such constraints, downcore variations in the stable isotope composition have been correlated with changes in climatic conditions, i.e. glacial and inter-glacial periods (Parker et al., 1972; Newman et al., 1973; Fontugne and Duplessy, 1986). The stable isotope composition of OM deposited during glacial periods are expected to be isotopically depleted. The opposite trend may be observed in some areas where glaciation causes a low rate of mixing of ocean waters, due to the low rate of dissolution of both carbon dioxide and nitrogen.

2. METHODOLOGY

2.1. Source of Material

The samples used in this work are from ODP Leg 117, sites 724 and 725. Three holes, designated A, B and C, were drilled at sites 724 and 725 but only samples from hole C were utilized for this study. Hole 724C was drilled to 252.4 mbsf and the sediments recovered range from the early Pliocene to Holocene in age. The sampling interval was at approximately 20cm intervals and the thickness of homogenized sample was about 2cm. A total of 804 samples collected from the upper 204m were analyzed.

Drilled depth at Site 725C is 162.8m and samples from the upper 32.2m only were used in this study. The sampling interval was similar to that of hole 724C. The samples analyzed in this study range from Pleistocene to Recent in age. A total of 124 samples were analyzed in this study.

Because northeast and east Africa is one of the probable sources of allochthonous OM in the Arabian Sea, 9 samples of tree leaves and 11 grass samples from the mainland Tanzania were analyzed for the stable isotope composition of both carbon and nitrogen.

2.2. Stable Isotope Analysis

Sediment samples collected for stable isotope analysis were stored frozen. With the exception of samples from the upper 30m, from Site 724, which were shipped separately, all samples thawed during shipment (Shipboard Scientific Party, 1989). These samples were refrozen upon arrival at the ODP office and maintained with the other samples at -10°C until processing.

Samples were dried at 40°C, ground to a fine powder (less than 80 phi) and acidified using 30% HCl to remove inorganic carbon. Because of the high carbonate content of the sediment, the carbonate free residue was washed of all salts. This was done so as to avoid absorption of water vapour, by salts and errors arising during weighing. The carbonate free residue was then dried at about 40°C for two weeks. Such a washing process may cause the loss of soluble organic materials.

A portion of the dried material was then weighed and placed in the quartz tube. On average, 100mg of the sample was used. Because the samples are combusted while in vacuum state, purified cupric oxide wire and high purity granular copper were added as oxidizing and reducing agents respectively (Macko et al., 1984). After the samples were evacuated to a high vacuum and sealed, they were combusted at 850°C for one hour. In order to avoid the formation of carbon monoxide and nitrous oxides, samples were left to cool slowly.

The N_2 and CO_2 gases were separated cryogenically from other combustion products following the technique of Macko (1981). Purified N_2 and CO_2 gases were analyzed on a V.G. Micromass PRISM stable isotope ratio mass spectrometer. On the basis of replicate analyses of samples, the reproducibility in combustion and measurement is within ±0.2°/$_{oo}$, n = 20. Isotope data are reported as δ-values defined as:

$$\delta^{N}E = \left[\frac{R_{sample}}{R_{standard}} - 1\right] * 10^{3}$$

where N is the isotope of the element E, and R is the abundance ratio of the heavy to light isotope; the standard for ^{15}N is atmospheric nitrogen and the standard for carbon is the Chicago Pee Dee Belemnite (PDB). For routine measurement, samples were analyzed against a laboratory standard of either pure nitrogen or carbon dioxide gas. The percentage of residue OC content (Table 1 and 2) was determined on the carbonate free residue as the volume of carbon dioxide gas released during the combustion. The residue OC percentages for the upper 30m of the core 724C were corrected for carbonate content (Table 3). Owing to lack of $CaCO_3$ data for the remaining part of the core 724C and core 725, the residue OC percentages (Table 1 and 2) are not corrected for carbonate content of the sediment.

2.3. Data analysis

Bivariate plots for both sites (Figs. 4 to 11) and linear regression analysis between measured variables were performed and their respective Pearson "R" correlation coefficient determined. Owing to the high variability of the data at both sites, smoothing methods using a three point moving average technique were performed to remove signal noise (Figs. 12 to 17). The average of three consecutive data points of geochemical results for core 724C, were determined and plotted against depth (Figs. 18 to 20). This was done to reduce the size of the data set without causing measurable distortion and removing noise from the signals. This was performed under the assumption that changes in the isotopic compositions and organic carbon contents of the sediment are different during glacial and interglacial periods, and durations of these periods are of the order of Kyrs. Three data points were selected for averaging, based on sedimentation rates as determined by the Shipboard Scientific Party (1989). For sedimentation rates which range from 35 m/m.y. to 142 m/m.y., a sampling interval of 20cm, represents a time interval between 5700 to 1400 years. An average of three data points would represent a time interval of about 17100 to 4200 years. Since the upper 180 mbsf of the core

724C has higher sedimentation rates, a glacial or interglacial period with a duration of about 20 kyrs, would be represented by at least four data points. If there is any marked change or trend while changing from one stage to another, it would be revealed by this average since noisy features would be smoothened out.

In order to determine if samples from the two sites (724 and 725) are statistically from the same population, t-tests (Table 4) were performed. Furthermore, stable isotope data from the upper 30m for Site 724 were correlated with oxygen isotope stages. The oxygen isotope stages for the upper 30m of the core 724C have been determined by Zahn and Pedersen (in press). Extrapolation of the oxygen isotope stages for core 724C from the determined isotopic stage 10 (Zahn and Pedersen, in press), to the determined Brunhes/Matuyama boundary (Prell et al., 1989) was performed using smoothened and three point average curves (Figs. 24 and 25). Moreover, some isotope stage boundaries as determined by Williams et al., (1988) have been assigned to the established palaeomagnetic and biostratigraphic boundaries (Prell et al., 1989). These boundaries include Brunhes/Matuyama, top of the Jaramilo, bottom of the Jaramilo and top of the Olduvai.

3. RESULTS

3.1. Stable isotopes and organic carbon content

The stable isotope composition of organic carbon and nitrogen and residue OC content data for the analyzed sediment samples from both sites are presented in Tables 1 and 2. Comparison of the means of $\delta^{13}C$, $\delta^{15}N$ and residue OC content for the two sites (724 and 725) using the t-test method (Table 4) show that these two groups of samples are statistically not from the same population. Furthermore, a plot of $\delta^{13}C$ against $\delta^{15}N$ for both sites (Figs. 7 and 11) indicate the independence of each variable. An increase or decrease in $\delta^{15}N$ values has no effect on the carbon isotopic composition of organic matter.

3.1.1. Site 724

A total of 804 samples from upper 204m of core 724C were analyzed. Results of the stable isotope composition analysis of organic carbon and nitrogen and residue OC content are presented in Table 1. As pointed out previously, the sampling interval was approximately every 20cm and a sample of about 2cm thick was homogenized. Since the sedimentation rate at this site ranges from 35 m/m.y to 142 m/m.y., the homogenized sample represents an averaged time of sediment flux to the sea floor of about 570 to 140 years.

The residue OC content of the acidified sediment residue for samples from core 724C, range from 0.2% to 18.2%, averaging about 4.9% with a standard deviation of ±2.9%. Although the range is high, most of the data points are higher than 2% and less than 10%. Generally, during the Pleistocene-Holocene, low levels of residue OC are noted (averaging 3.8%) compared to that of the late Pliocene (averaging 6.5%). A scatter plot of residue OC content against depth (Fig. 6) shows a trend of increasing residue OC with depth. This trend is even more evident when data points in 10m intervals are averaged and plotted against depth (Fig. 23). An increase in residue OC with depth can be divided into four incremental steps (0-60m, 60-90m, 90-120m, 140-204m). From the core top to 60m there is a gradual increase at a rate of about 0.1% per 10m (Fig. 23). A further increase of about 1% from the previous average (2.7%) is noted between 60m and 70m. The rate of increase in this segment, which lies between 60 and 90m, is approximately 0.2% per 10m (Fig. 23). An abrupt change in OC content near 90m, almost of the same magnitude as the one at 60m, reaches its apex at approximately 110m. Thereafter, a sharp decline in the residue OC content occurs towards the Pliocene-Pleistocene boundary (near 130m) where it reaches its lowest value near 135m. From 140m, the residue OC increase sharply with increasing depth.

The profiles of total OC content, total organic nitrogen

(ON), C/N ratios (data from Zahn and Pedersen, in press) and residue OC, versus depth for the upper 30m, do not show clear variations with changes in glacial-interglacial periods (Figs. 26, 27, 28 and 29). However, the average values of total OC, ON contents (data from Zahn and Pedersen) and residue OC for each isotope stage, show some cyclic changes (Table 5). These changes indicate some degree of correlation of OC with oxygen isotope stages for the upper 30m of the core. These average values indicate slight preferential preservation of OM during interglacial periods than glacial periods. Although the differences in averages of the total OC and ON contents between glacial and interglacial periods are not high, the observed cyclicity in the averages may be an indicator of changes in hydrographic parameters. Statistical tests for equal population means for the determined oxygen isotope stages using the one way ANOVA method, indicate differences in the population means ($F_{0.05,147,9} < 2.535$). There are no measurable differences in the averages of amount of OM preserved during oxygen stages 3 and 4.

A trend which might be indicative of the glacial and interglacial records below the determined isotope stage 10 (Zahn and Pedersen, in press) up to the Brunhes-Matuyama boundary is revealed on the smoothened and three point average curves (Figs. 24 and 25). It is interesting to note that from the boundary of oxygen isotope stage 9 (Zahn and

Pedersen in press) to the Brunhes/Matuyama boundary, about 5 glacial and 5 interglacial stages are recognizable. When these trends are assigned isotope stages (continuous numbering) from isotope stage 9, isotope stages up to 21 are clearly demarcated (Figs. 24 and 25). The assigned glacial isotope stages correspond to low residue organic carbon content. The determined boundary of top of the Olduvai, which is within the range of isotope stage 57 as defined by Williams et al., (1988), correlates well with the most depleted isotope values of organic carbon and high residue OC content. These isotope boundaries at Site 724C are approximate, and have been determined based on visual observation and abrupt changes in the stable OC isotope compositions. For accurate isotope stage boundaries, oxygen isotope data are required.

The stable isotope composition of organic carbon of the sediments from Site 724C range from $-17.7^{o}/_{oo}$ to $-24.7^{o}/_{oo}$ with an average value of $-20.4^{o}/_{oo}$. Generally, the variance of $\delta^{13}C$ values is low (standard deviation of $0.8^{o}/_{oo}$) and a plot of $\delta^{13}C$ versus depth (Fig. 4) shows a slight enrichment in the Pleistocene-Holocene section ($-20.0^{o}/_{oo}$) over the Pliocene ($-21.0^{o}/_{oo}$). This trend is emphasized when data points at 10m intervals are averaged and plotted against depth where, a slight depletion with increasing depth is noted (Fig. 21). Moreover, this type of plot shows three depletion lows or zones in ^{13}C of about 0.5°/$_{oo}$ from the average above and below it. On close examination of non-averaged plot, these zones, which correspond to low values of OC content, occur at depths of about 25m, between 80 and 90m, and near the Pliocene-Pleistocene boundary (130-140m).

Correlations of the stable carbon isotope data with oxygen stages (Zahn and Pedersen in press), show slight enrichments during glacial periods, especially during oxygen isotope stages 2 and 4. There is no measurable shift in isotopic compositions between oxygen isotope stages 6 and 7. Similar trends of slight enrichment in ^{13}C are observable for the remaining parts of the core when oxygen isotope stage boundaries (Williams et al., 1988), are fitted to biostratigraphic and palaeomagnetic age boundaries (Prell et al, 1989). Clear cyclic changes from glacial to interglacial periods are observable on the smoothened curve (Fig. 24) and when three data points are averaged and plotted against depth (Fig. 25). From isotope stage 10 (as determined by Zahn and Pedersen in press) up to Brunhes/Matuyama boundary, ten (10) isotope stages are recognized.

The stable nitrogen isotopic composition of sedimentary OM from Site 724 is highly variable (standard deviation of ±2.5°/$_{oo}$; Fig. 5). However, the Pliocene core section has the most uniform and most depleted $\delta^{15}N$ values, which are less than the overall average. The isotope values ranges from 1°/$_{oo}$

to $18.9^{\circ}/_{\infty}$ and averages $8.9^{\circ}/_{\infty}$. The samples from the very surface of the core are depleted in ^{15}N. In general, there is a gradual increase in $\delta^{15}N$ values with increasing depth. The increase in $\delta^{15}N$ reaches an average value of about $10.4^{\circ}/_{\infty}$ at a depth interval of 30 to 40m. Below this depth there is a depletion in $\delta^{15}N$ with increasing depth. This trend is evident when data points are averaged for every 3 consecutive points and 10m interval, and plotted against depth (Figs. 19 and 22). Stable nitrogen isotopic compositions below 100m are less than the overall average of $8.9^{\circ}/_{\infty}$ by about $1.2^{\circ}/_{\infty}$. This implies that the Pleistocene-Holocene samples are more enriched in ^{15}N than those of the late Pliocene.

No significant correlations at 95% confidence level (Pearson R, Table 6) were observed between either isotopic compositions or between isotopic compositions and residue OC content.

3.1.2. Site 725.

A total of 124 samples from the upper 32.2m were analyzed. The sedimentary OC contents at Site 725 are relatively low compared to that of Site 724 (Table 2). The residue OC content ranges from 0.2% to 2.1% averaging about 0.9%. A plot of residue OC content against depth (Fig. 10) shows no trend with depth. On close examination there are cyclic fluctuations. Generally, about four minima (low in OC content) are recognizable and roughly correspond to $\delta^{13}C$ minima (Fig. 17). The lowest OC content values, which correlate well with the most depleted $\delta^{13}C$ values, occur at a depth of about 28m and extends for about 2m up to 30m (Fig. 17).

The stable carbon isotope composition of sedimentary OM for samples from Site 725 range from -17.6°/₀₀ to -25.6°/₀₀ with an average of -20.1°/₀₀. There is an increase in $\delta^{13}C$ with increasing depth from the surface downcore to about 4m followed by a systematic depletion in ^{13}C with increasing depth to approximately 8m (Figs. 8 and 15). The stable isotope composition of OC for the remaining part of the core is relatively constant. The trend of constant $\delta^{13}C$ values is interrupted by depletion of about 4°/₀₀ at a depth of about 28m. Apart from the above described trends, there is a slight overall (0.5°/₀₀ ±0.2°/₀₀) depletion toward the bottom of the analyzed core (Fig. 15).

The stable nitrogen isotope composition is highly variable (Fig. 9) and range from 4.6°/₀₀ to 17°/₀₀ averaging 11.3°/₀₀. The $\delta^{15}N$ values show little variation with climatic change (glacial and interglacial cycles).

There is a statistically significant relationship or correlation between residue OC content and stable carbon isotope composition (Table 6). About 10% of the variance of $\delta^{13}C$ can be attributed to residue OC content and vice versa.

There is no significant correlation between stable nitrogen isotopic compositions and either residue OC content or stable carbon isotopic compositions.

3.2. Stable Isotope Compositions of Terrestrial Plants

Tree and grass samples collected from Tanzania are representative of both wet and semi-desert environments. According to Lind and Morrison (1974) some of grass species collected are also found in the areas surrounding Gulf of Aden.

3.2.1. Tree Leaves Samples.

The isotopic compositions of carbon and nitrogen results for tree leaves samples are presented in Table 7. The $\delta^{13}C$ for the tree leaves analyzed range from $-26.7^{o}/_{oo}$ to $-31.8^{o}/_{oo}$ with an average of $-29.3 \pm 1.4^{o}/_{oo}$. The $\delta^{15}N$ values range from $1.0^{o}/_{oo}$ to $8.6^{o}/_{oo}$ and average $5.2 \pm 2.4^{o}/_{oo}$. The $\delta^{13}C$ values indicate that most of the sampled trees follow a C_3 metabolic pathway. The average of $\delta^{13}C$ value for these higher terrestrial plants can be considered to be representative of the terrestrial end member for the allochthonous input to the Arabian Sea.

3.2.2. Grass Samples.

The results of the isotopic compositions of carbon and nitrogen for 11 grass samples are presented in Table 7. The $\delta^{13}C$ values range from $-11.2^{\circ}/_{\circ\circ}$ to $-23.8^{\circ}/_{\circ\circ}$ and average $-13.2 \pm 3.4^{\circ}/_{\circ\circ}$. The $\delta^{15}N$ values range from $2.0^{\circ}/_{\circ\circ}$ to $10.3^{\circ}/_{\circ\circ}$ and average $7.7 \pm 2.7^{\circ}/_{\circ\circ}$. Based on the stable isotopic compositions of carbon, most of the grass samples analyzed are of the C_4 type. Only the Phragmites mauritianus appears to follow C_3 metabolic pathway.

Two observations can be assessed or pointed out from the analytical result of terrestrial samples from Africa (Tanzania). First, the nitrogen isotopic compositions of both tree leaves and grasses are more enriched in ^{15}N than the reported $\delta^{15}N$ values for terrestrial materials. For example Sweeney and Kaplan (1980) reported $2.5^{\circ}/_{\circ\circ}$ as the end member value for terrestrial materials. Second, the $\delta^{13}C$ value of the terrestrial end member may be higher than those commonly used if the amount of grass material contributed to the total terrestrial OM is large. In terrestrial environment where both C_3 and C_4 plants can contribute similar amount of OM, observed $\delta^{13}C$ values represent a mixture of two metabolic pathways. Although the $\delta^{13}C$ of the dominant plant type will be reflected to a greater degree, the $\delta^{13}C$ value of the mixture will be greater than $-26^{\circ}/_{\circ\circ}$.

4. DISCUSSION

4.1. Source of Sedimentary Organic Matter

Stable isotopic compositions of organic carbon and nitrogen of sediment samples from sites 724 and 725 do not show any measurable input of allochthonous OM into the area. Since rivers are the major avenue of terrestrial materials to the ocean, the observed phenomenon is likely due to the lack of major rivers. The Indus river is the only major river which contributes considerable amounts of terrestrial materials from the Himalayan mountains to the Arabian Sea. Smaller rivers draining from western India, which may contribute a significant amount of sediments, are the Narmada and Tapti.

With the exception of the eastern side (India), which receives fairly high rainfall, land masses surrounding the Arabian Sea are arid. Transport of terrigenous materials from arid regions to the Arabian Sea is thought to be mainly by wind (Kolla et al., 1981a, b). Owing to the absence of major rivers, terrestrial materials were not transported far offshore from the present shoreline even during periods of low sea level. Correlation between depleted $\delta^{13}C$ values and low OC content at Site 725 may be indicative of a slight terrestrial influence on sites located at shallow depths.

Another possible reason for low input of allochthonous organic materials into the Arabian Sea is the low availability of terrestrial OM in the source regions. Input of OM from the surrounding continents into the Arabian Sea by wind may have been low due to low rate of OM production in Asia and Africa. Low productivity in the source regions can be attributed to limited precipitation. Based on pollen records, monsoon pollen index and quartz distribution in the Indian Ocean, the land masses surrounding the Indian Ocean were arid and the Arabian Desert dunes were more extensive during glacial periods (Kolla and Biscaye 1977; Kolla et al., 1981b; Van Campo et al., 1982; Prell and Kutzbach 1987). Because of increased aridity, low amounts of organic materials were available for wind transport whereas more clastic materials were plentiful. Thus, causing the effect of terrestrial organic material transported by monsoon winds to this depositional basin to be small relative to autochthonous production associated with the upwelling in the region. Occasional excursions of more depleted carbon and nitrogen stable isotope values are more likely to be the results of changes in primary productivity or diagenetic processes.

4.2. Productivity and Preservation of Organic Matter

The residue OC content for the whole core at Site 724 averages about 4% higher than that of Site 725, but it is only about 1.4% higher than that of Site 725, for the upper

30m of the core. This difference may be an indication of higher productivity and/or preservation at offshore Sites. The $CaCO_3$ content at Site 725 (Prell et al., 1989) is not much different from that of Site 724 (Zahn and Pedersen, in press). The $CaCO_3$ content of marine sediments is inversely related to organic matter content. Therefore, differences in $CaCO_3$ content can not account for differences in OC content between the two sites. Small differences in OC content in the upper 30m between these two sites, which also correspond to high $\delta^{15}N$, are indicative of similar processes taking place but with different degrees of magnitude. Sediment samples from Site 725 have $\delta^{15}N$ values, which are on average about $3^{o}/_{oo}$ more enriched than those at Site 724. This suggests that the nearshore environment at Site 725 may be more affected by the denitrification process associated with a more stable oxygen minimum than at Site 724.

High levels of residue OC noted in the late Pliocene relative to that of the Pleistocene-Holocene are indicative of different hydrological conditions in the region. Large amounts of organic materials preserved during late Pliocene and early Pleistocene are likely a result of high primary productivity and/or enhanced preservation. Enhanced preservation could be due to the invasion of highly saline waters from inland marginal seas of Red Sea and Persian Gulf and/or increased sedimentation rate.

Communication between these (Red Sea and the Persian Gulf) inland marginal seas and the Indian Ocean is inferred to have commenced since the late Pliocene (Kaz'min, 1986; Bowen and Jux, 1987). Salinity of outflow waters is likely to have been high at the beginning of the communication and as exchange continued, salinity was slightly reduced as a result of dilution by fresher Indian Ocean waters. Input of the Red Sea and Persian Gulf saline waters would have increased the density contrast between surface and intermediate waters, thus preventing bottom water ventilation. The lack of bottom water ventilation would have facilitated development of anoxic conditions and the formation or deposition of varves. Deposition of varves can be caused by other factors including high primary productivity. High fluxes of OM to the sea floor, owing to high primary productivity, can cause the development of anaerobic conditions due to high demand for oxygen during mineralization of OM. Anoxic condition hinders the occurrence of benthic organisms in the area below the OMZ. This allows deposition of varves or alternating organic rich lamina, with organic lean lamina deposited during periods of lower productivity and no bioturbation. During Leg 117 drilling work, laminae rich in biogenic silica were only observed for sites (723, 724, 725 and 726) located at a water depth less than 1000m on the Oman Margin sites (Prell et al., 1989). The laminae are confined to those sites (723, 724, 725 and 726) which are located within the zones influenced by highly saline waters from both the Red Sea and the Persian Gulf, and in the late Pliocene and early Pleistocene section of the cores.

The establishment of anoxic conditions alone does not diminish net microbial metabolism, and depending on the availability of labile OM, microbial activity may increase (LaRock et al., 1979). Most labile organic materials in near-shore environments where suboxic and anoxic conditions prevail, are consumed by anaerobic bacteria (Berner 1980). Apart from the enhancement of anoxic conditions, highly saline waters could have caused reduction of microbial habitats at intermediate water depths, thus reducing the rate of decay of OM. The elimination or reduction in the population of micro-organisms owing to saline water intrusion in the bottom waters will cause preservation of large amount of OC and formation of lamination even if productivity in the photic zone is low. The formation of sapropel in the Mediterranean Sea have contributed to the curtailment of microbial activity owing to salinity excursions (Klinkhammer and Lambert, 1989). One of the factors which may have contributed to enhanced preservation of OM in the late Pliocene section is likely to be anoxic conditions and reduction in microbial activity due to the intrusion of saline waters from the inland marginal Seas.

Other possible causes of the observed high OM could be high primary productivity and increased sedimentation rates. Although laminae are not always indicative of increased productivity (Howell et al., 1988), the observed laminae in association with high OC and biogenic silica contents can be partially attributed to high primary productivity. Abundances of siliceous biogenic sediments reflect the biological fertility of the surface waters. The dissolution of biogenic (skeletal) opal takes place in the water column (Berger, 1968; Lisitzin, 1972), at the sediment-water interface and in the sediments (Nigrini, 1968; Berger, 1970). The solubility of silica decreases with increasing pressure and decreasing temperature. Under open ocean conditions, all, or nearly all, of the opal is dissolved before being buried in the sediments. An increase in nutrient and silica supply in the photic zone during upwelling and mixing causes blooms of opaline skeletal building organisms such as diatoms, radiolaria and silicoflagellates. High productivity of biogenic opal in the surface waters and concurrent flux to the sea floor causes saturation of interstitial waters in terms of silica, thus reducing the dissolution and recycling of siliceous organisms immediately after death. Reduction in the rate of dissolution of biogenic silica, results in the preservation of appreciable amounts of cpaline skeletons.

Therefore, the observed varves rich in biogenic silica and OM might have been caused by high primary productivity. This inference is further supported by the observed inverse relationship between biogenic silica and carbonate contents (Prell et al., 1989) and also by the isotopic compositions of both organic carbon and nitrogen which are of marine nature.

The total amount of OM preserved in the sediments depends, among other factors, on OM flux to the sea floor (Betzer et al., 1984; Emerson et al., 1985). In areas where terrestrial OM input is negligible, there is a positive correlation between the sedimentation rate and OC content deposited in oxic environments (Muller and Suess, 1979; Stein, 1986b). This is probably due to the fact that high sedimentation rates reduce the residence time of organic materials in the water column and bioturbation zone. At lower sedimentation rates distinctions between sediments deposited in oxic and anoxic conditions can be made (Stein, 1986b). Such distinctions cannot be made in this study, owing to the high sedimentation rates at this site (724).

High amounts of OM in late Pliocene early Pleistocene sections have been reported on a worldwide scale (Emerson, 1985). Thus the cause of enhanced preservation may not be a localized phenomena.

The observed decrease in the amount of OM preserved in the sediments since the early Pleistocene is likely a result of an increase in regeneration in the water column (poor preservation) and/or a decrease in productivity. A high degree of OM degradation may result from active denitrification and/or sulphate reduction. High oxygen content is unlikely at these sites since, during drilling, high chlorinity was observed at both sites in the upper 80m. High chlorinity is associated with high salinity which causes stratification and inhibits the exchange of oxygenated surface waters with those of intermediate depths. Therefore, the most likely cause of low OC content is denitrification and persistent sulphate reduction processes. These processes could result from increased microbial activity owing to increased microbial populations, which are tolerant to elevated salinity. Evidence of sulphate reduction at Sites 723, 724 and 725 extended beyond the depth commonly observed in the world oceans (Prell et al., 1989). The operation of a denitrification process, is supported by heaviness of the stable isotopes of nitrogen in these sediments.

A decline in the OC content towards the Pliocene-Pleistocene boundary coincides with lower $\delta^{13}C$ and $\delta^{15}N$ values and may be attributed to low primary productivity. Poor preservation is unlikely, because it would have caused heavier $\delta^{15}N$ values, since nitrogenous materials are susceptible to biodegradation. These low isotope ratios of marine nature are probably a result of fractionation caused by low primary productivity coupled by incomplete utilization of available nutrients.

Slight preservation of organic matter during interglacial periods may indicate increased primary productivity or an improvement in preservation conditions. The inference of increased productivity is in agreement with previous studies (Labracherie et al., 1983; Prell, 1984; Fontugne and Duplessy, 1986). An increase in primary productivity, is likely a response to increased coastal upwelling caused by intensification of the SW monsoon winds over the weak NE monsoon winds. The difference between a weak SW monsoon and strong NE monsoon during the glacial periods may be related to the environmental conditions prevailing over the Himalayan mountains and Tibetan plateau. The thickness and spatial distribution of snow cover have a direct effect on the amount of solar radiation absorbed by the ground (Barnett et al., 1989). High snow cover associated with high albedo keeps the ground cold for longer time intervals. In modern conditions, large parts of the central and southern Tibetan plateau and Himalayan mountains remain free from snow throughout most of the year. This condition seems to accelerate the onset of the Asian monsoon and increases its ultimate intensity because of rapid heating during the spring in the northern hemisphere. During an ice age, these mountains are covered with glaciers (CLIMAP, 1976). The permanent ice coverage prevents any marked warming of the Asian continent at high altitude during the summer. Such decreases in the summer temperature difference between the land and ocean would have resulted in low pressure differences between the Asian continent and the Indian Ocean and thus, weak SW monsoon winds. During the winter, the presence of permanent ice would have helped to maintain a much cooler continent than today in the vicinity of a warm ocean. These conditions would have enhanced the dry NE monsoon winds.

The inferred local increase in productivity during interglacial periods contradicts the worldwide observation that high primary productivity in the oceans is higher during glacial periods (Muller and Suess, 1979; Pedersen, 1983; Emerson and Hedges, 1988; Sarnthein et al., 1988). This contradiction may be due to the differences in the onset of the driving forces involved in causing upwelling. Oceanic and atmospheric circulations are generally driven by the thermal contrast (temperature gradient) between low latitudes (tropical) and high latitudes (polar) regions. Steep temperature gradients are associated with intense oceanic and atmospheric circulation. During glacial periods, intense upwelling along the equator and other major upwelling centers took place in response to intense winds caused by high temperature contrast between low and high latitudes. However, in the Indian Ocean, the intensity of the monsoon winds, which determine the magnitude of coastal upwelling in the area, does not depend on temperature difference between low and high latitudes. The intensity of the monsoon winds, thus coastal upwelling, depends on variation in solar radiation, surface albedo and altitude (Hahn and Manabe, 1975; Prell, 1984; Webster, 1987; Barnett et al., 1989). As pointed out previously, glacial periods, which are associated permanent ice coverage on the Himalaya mountains and the Tibetan Plateau, cause weak SW monsoon winds, hence, weak coastal upwelling and low primary productivity.

4.3. Stable Isotope Compositions of Carbon and Nitrogen

An increase in both $\delta^{13}C$ ($1.0^{\circ}/_{oo}$) and $\delta^{15}N$ (more than $2^{\circ}/_{oo}$) associated with a decrease in OC content in the Pleistocene over that of the Pliocene, is an indication of different oceanographic conditions. A shift (depletion or enrichment) in the isotopic composition of organic carbon and nitrogen can be caused by a number of factors including; changes in species, primary productivity, recycling of nutrients, oxidative degradation and diagenetic processes (see section 1.7.1 and 1.7.2). A change in species from siliceous dominated sediments in late Pliocene to calcareous dominated type of Pleistocene sediments has been observed during Leg 117 drilling program (Prell et al., 1989). Diatoms which are major contributors of siliceous materials to the sea floor, are slightly more enriched in ^{13}C than nannoplankton (Gearing et al., 1984). Since carbon is fractionated less by diatoms relative to nannoplankton, one would expect a slight enrichment in ^{13}C in the Pliocene with respect to the Pleistocene. Although a change in species population might have contributed to the observed differences, another mechanism is required to explain enrichment trend in the opposite direction.

The possible explanation of enrichment in ^{13}C and ^{15}N may be high primary productivity and complete utilization of available nutrients. Slight enrichment in ^{13}C could be a result of a food chain effect if the observed $CaCO_3$ was produced by foraminifers and not nannoplankton. A decrease in OC content with increasing $\delta^{15}N$ values of more than $12^{\circ}/_{\circ\circ}$, which are confined to the upper 90m of the core, points to additional causes including possibly denitrification and sulphate reduction processes. Denitrification and sulphate reduction could have been caused by the presence of sufficient labile undegraded OM and anoxic conditions. Labile organic materials and anoxic conditions are likely to result from a rapid flux of OM to the sea floor than the rate of oxidation in the water column and in the surface layer of oxygenated sediments and complete utilization of oxygen during mineralization of OM.

A similar enrichment of ^{15}N of the same magnitude as the one observed in this study has been reported for particulate OM undergoing diagenetic reactions in the water column (Saino and Hattori 1980, 1987; Altabet and McCarthy, 1985; Cifuentes et al., 1988). Similarly, enrichments of ^{15}N residual nitrate of more than $12^{o}/_{oo}$, which is under the influence of enhanced oxygen minimum and extensive denitrification have been observed (Cline and Kaplan, 1975; Libes and Deuser, 1988; Liu and Kaplan, 1989). During denitrification ^{14}N enriched nitrate is preferentially removed. This leaves the residual nitrate enriched in ^{15}N, which may be brought to the surface as a result of upwelling. When this NO_3 is assimilated by plankton, the resulting biomass could be relatively enriched in ^{15}N. An increase in $\delta^{15}N$ values in the Pleistocene section may be the result of increased denitrification.

Bacterial sulphate reduction results in degradation of OM. During the sulphate reduction process, isotopically light carbon resulting from degradation of organic materials is evolved, leaving residue enriched in ^{13}C (Anderson and Arthur, 1983). The observed slight enrichment in ^{13}C and ^{15}N in the Pleistocene may be a result of these diagenetic processes. This inference is further supported by the observation of extensive sulphate reduction up to 50m at both sites 724 and 725 (Prell et al., 1989).

Based on the enrichment trend of ^{13}C and ^{15}N, the intensity of denitrification and sulphate reduction process appears to have increased since its onset, probably in early Pleistocene. The intensity of denitrification reached a maximum in the middle Pleistocene.

As noted previously, various workers have indicated that the areas surrounding the Arabian Sea were drier during glacial periods than during interglacial periods (Van Campo et al., 1982; Kolla and Biscaye, 1977; Sirocko, 1989). The SST during glacial periods were higher than interglacial periods (Van Campo et al., 1982). Moreover, the SST difference between summer and winter are inferred to have been low. This may be the result of increased aridity and lower sea-levels where continental shelves were shallower and had diminished coastal upwelling. Because of higher evaporation than precipitation coupled by weak SW monsoon winds, the Arabian Sea probably experienced heightened stratification. This contributed to the intensification of the OMZ, the recycling of nutrients, and lower rates of transport of aerated water to the intermediate depths. These factors in conjunction with the influence of ^{13}C from bicarbonate, would result in dissolved carbon dioxide and observed organic materials enriched in ^{13}C during glacial periods (Muzuka et al., in press).

4.4. Alteration of Sedimentary Records

The alteration of the intensity of the monsoon winds, upwelling and primary productivity records preserved in the sediments can be caused by a multiple of factors. These factors include water current strength, topography and annual variations in upwelling in time and space.

Strong bottom currents on the shelf and upper slope may erode unconsolidated bottom sediments, leading to the disappearance of a typical upwelling and productivity record in the sediments (Futterer, 1983). Similarly, strong surface currents increase the residence time of organic materials in oxygenated waters, and cause freshly formed organic matter in the water column to be highly remineralized, and remaining refractory material, be deposited far from its source. This would lead to poor preservation of OM and possibly the alteration of its isotopic composition. Moreover, increases in the strength of currents or downwelling increases the residence time of planktonic diatoms, which represent a main constituent of upwelling influenced sediments, not allowing biogenic silica to reach bottom quickly enough for deposition. Thus, sediments may not show an enrichment of opal tests due to dissolution of the skeletons in the water column. Phosphate lag deposits observed at Site 726, may be an indication of strong bottom currents in the area. The samples analyzed from Sites 724 and 725 might not be a true representation of the actual processes taking place in the overlying water column.

Bottom dwelling organisms homogenize newly deposited and old sediments. This may remove any upwelling and paleoproductivity signals. The observed burrow mottles between beds and other bioturbation structures on both sites (724 and 725) may be an indication of homogenization by benthic organisms. Furthermore, erosion and redeposition of sediments during repeated transgressive-regressive cycles causes periodical removal and overprint of sedimentary record.

Steep continental margins and platform morphologies hinder deposition of sediments, causing poor preservation of signals of productivity taking place in the euphotic zone, coastal upwelling and monsoon winds. In the western Arabian Sea, the continental margin is narrow and steep. Owing to the steepness of the margin, deposition of sediments might have been hindered.

The interpretation of the isotope data of organic carbon and nitrogen is complicated by seasonal variations in the intensity of coastal upwelling. Upwelling centers have been shown to vary in space and time off Peru and north Africa as well as in the Arabian Sea (Currie et al., 1973; Diester-Haass, 1983; Labracherie et al., 1983; Reimers and Suess, 1983). This phenomenon leads to differences in the amount of organic matter produced in the water column and deposited in the underlying sediments from one place to another, although intensity of the winds might have been on average constant.

Because of microbial action, diagenesis and recycling of organic materials, the original isotopic compositions of phytoplankton may have been altered. The isotopic signature of high primary productivity in the photic zone may not always be reflected in the underlying sediments.

5. CONCLUSION

The isotopic compositions of stable organic carbon and nitrogen indicate that, the organic matter in the northwestern Indian Ocean (western Arabian Sea) sediments is mainly of marine origin. The effect of terrestrial material transported by the monsoon winds since late Pliocene is minimal. Based on the differences in relative contribution of terrestrial and marine sources of organic matter, no changes in the intensity of the SW and NE monsoon winds have been recorded. The amount of allochthonous organic matter may be higher than indicated by the stable isotopes of organic carbon and nitrogen, if biomarkers like lignin are used as a supplementary tool to delineate sources of organic matter in this marine environment. Low inputs of the terrestrial organic matter into the Oman Margin depositional basin can be ascribed to the absence of major rivers and low rates of organic matter production in the surrounding continents at times when NE monsoon winds were intense.

The degree of preservation and/or primary productivity in the western Arabian Sea, was higher during the late Pliocene than during the Pleistocene-Holocene epoch as indicated by OC content. The onset of the Pleistocene, marks the beginning of greater alteration of productivity and/or preservation of organic matter as a result of changes in hydrographic conditions and/or intensity of coastal upwelling. Near the Pliocene-Pleistocene boundary, productivity was low owing probably to weak SW monsoon and coastal upwelling.

The intensity of denitrification and oxygen minimum processes have increased since its onset, probably in the early Pleistocene. The intensity of the denitrification and oxygen minimum reached its climax in middle Pleistocene as a result of increased stratification of NIIW. Based on sedimentation rate [illegible] Site 724 and $\delta^{15}N$ data, denitrification reached its optimum intensity at around 400 to 500 Kyrs.

From the organic carbon isotopes and organic carbon content data, SW monsoon winds were more intense during interglacial periods. The intense SW monsoon winds were associated with intense coastal upwelling and high primary productivity and/or preservation.

The nearshore environment experienced lower primary productivity and/or poorer preservation. Moreover, the nearshore Site (725) is associated with more enriched ^{15}N values relative to the offshore areas (Site 724).

The effect of glaciation on the carbon isotopic composition of organic matter appears to be one of the factors causing slightly more enriched values, especially during ice stages 2 and 4.

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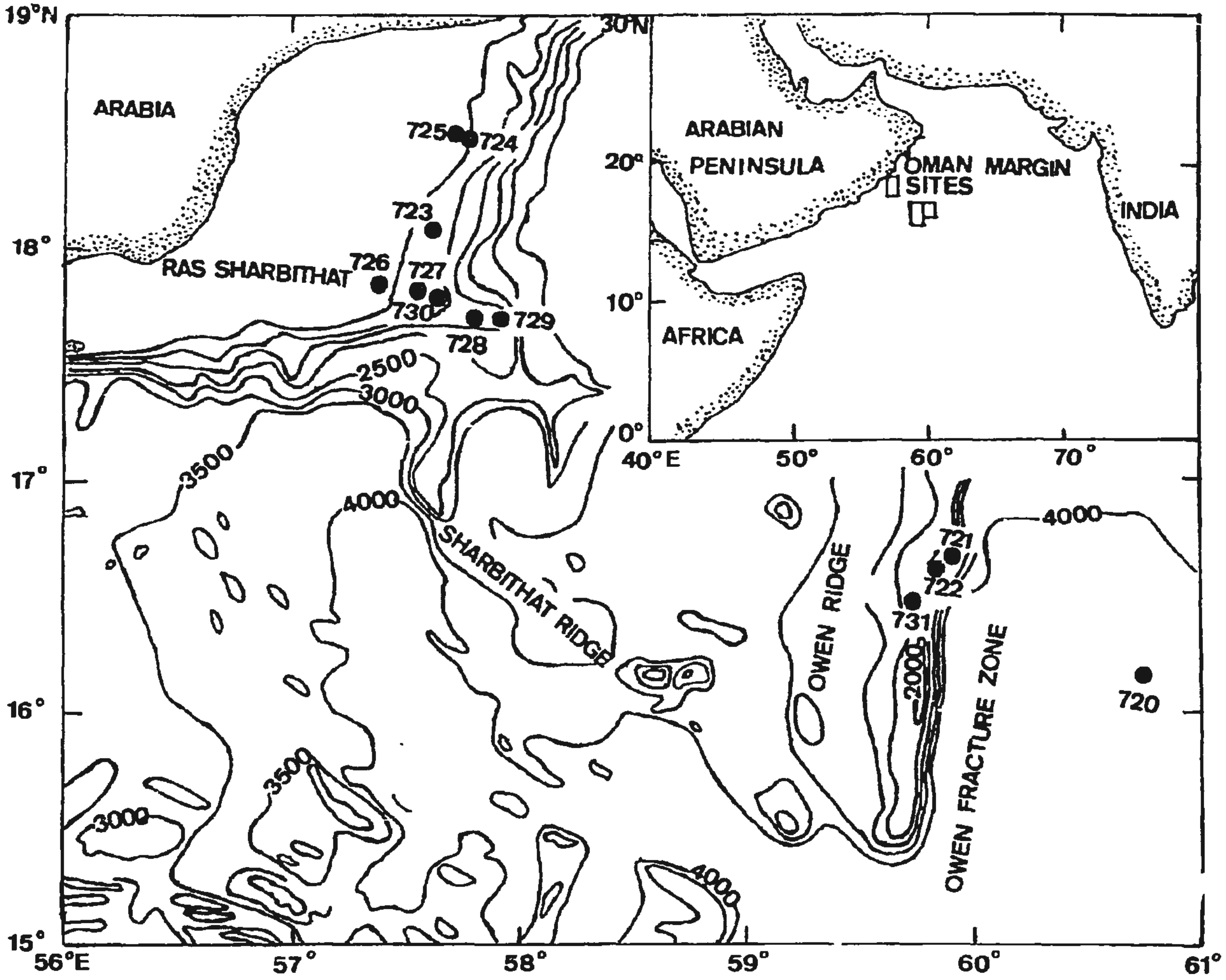
Figure 1: Location of the ODP Leg 117 Sites and bathymetry (contours in meters) of the area. (redrawn after Prell, Niitsuma et al., 1989)


19°N
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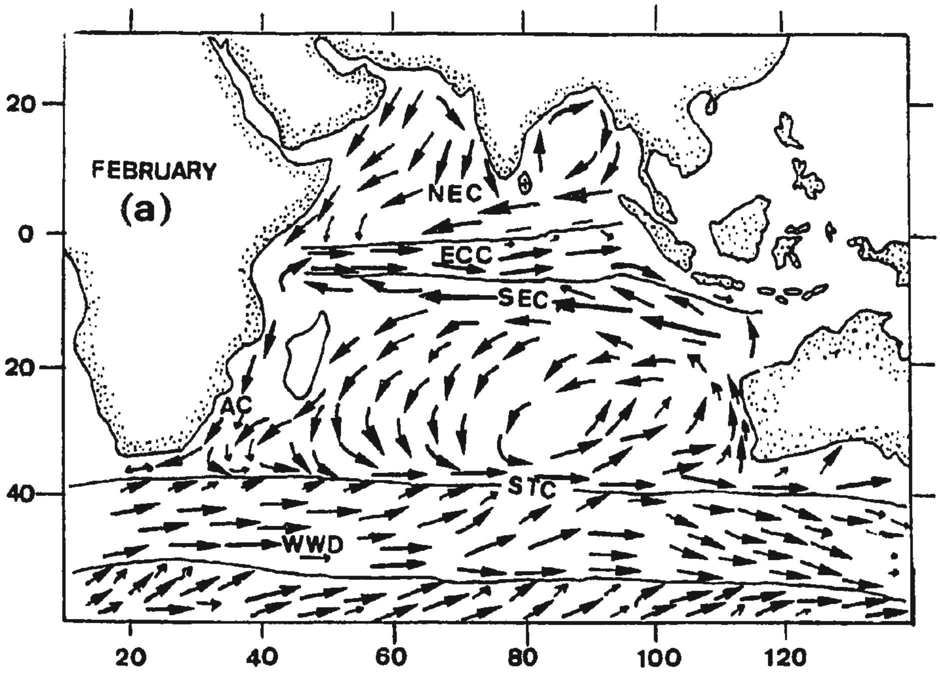
721
722
731
720
30°N
20°
10°
0°
40°E
50°
60°
70°
ARABIAN PENINSULA
OMAN MARGIN SITES
INDIA
AFRICA

Figure 2: Modern surface circulation patterns in the Indian Ocean during February (a) and August (b). WWD, West Wind Drift; STC, Subtropical Convergence; AC Agulhas Current; SEC, South Equatorial Current, ECC, Equatorial Current; NEC, North Equatorial Current; SWMC, Southwest Monsoon Current. (Redrawn from Wright and Thunell, 1988).


FEBRUARY
(a)
NEC
ECC
SEC
AC
STC
WWD
20
0
20
40
20
40
60
80
100
120

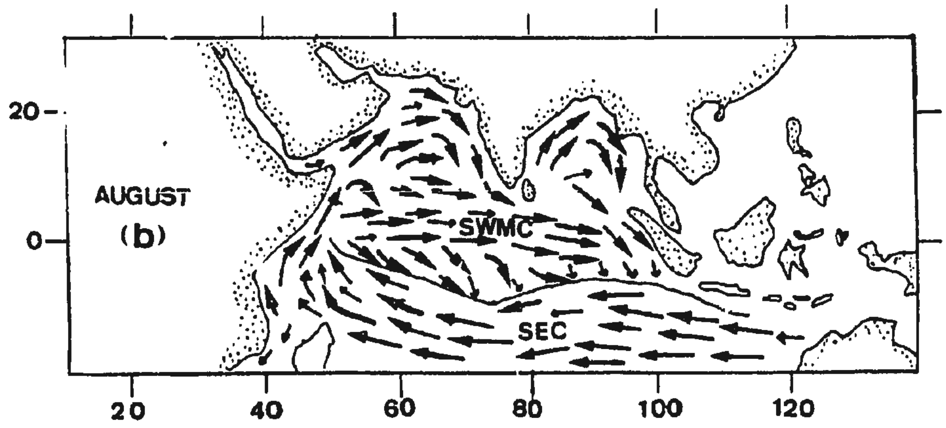

AUGUST
(b)
SWMC
SEC
20
0
20
40
60
80
100
120

Figure 3: Boundary conditions of monsoonal circulation for the modern winter and summer monsoons.

Upper: During the Northeast monsoon in modern January.

Lower: During the Southwest monsoon in modern August.

Shaded areas indicate elevation greater than 3000m (the Tibetan plateau and the Himalaya). Arrows denote wind direction,; sea surface temperatures are given in °C. During the Southwest monsoon, upwelling offshore Somalia and Oman (shaded area) lowers the sea surface temperatures to between 22° and 24°C. (Redrawn after Prell, Niitsuma et al., 1989).

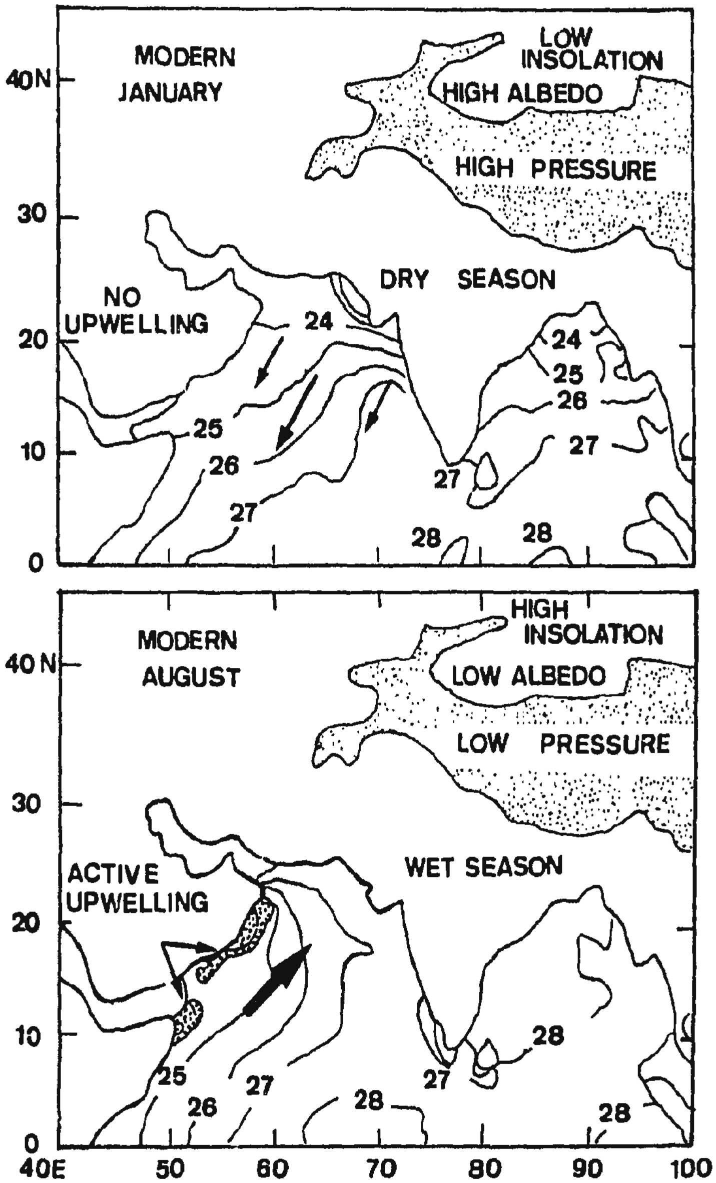

MODERN JANUARY
LOW INSOLATION
HIGH ALBEDO
HIGH PRESSURE
DRY SEASON
NO UPWELLING
24
25
26
27
28
MODERN AUGUST
HIGH INSOLATION
LOW ALBEDO
LOW PRESSURE
WET SEASON
ACTIVE UPWELLING
40N
30
20
10
0
40E
50
60
70
80
90
100

Figure 4: Variation of organic carbon isotope Values ($^{o}/_{oo}$) with depth at Site 724.

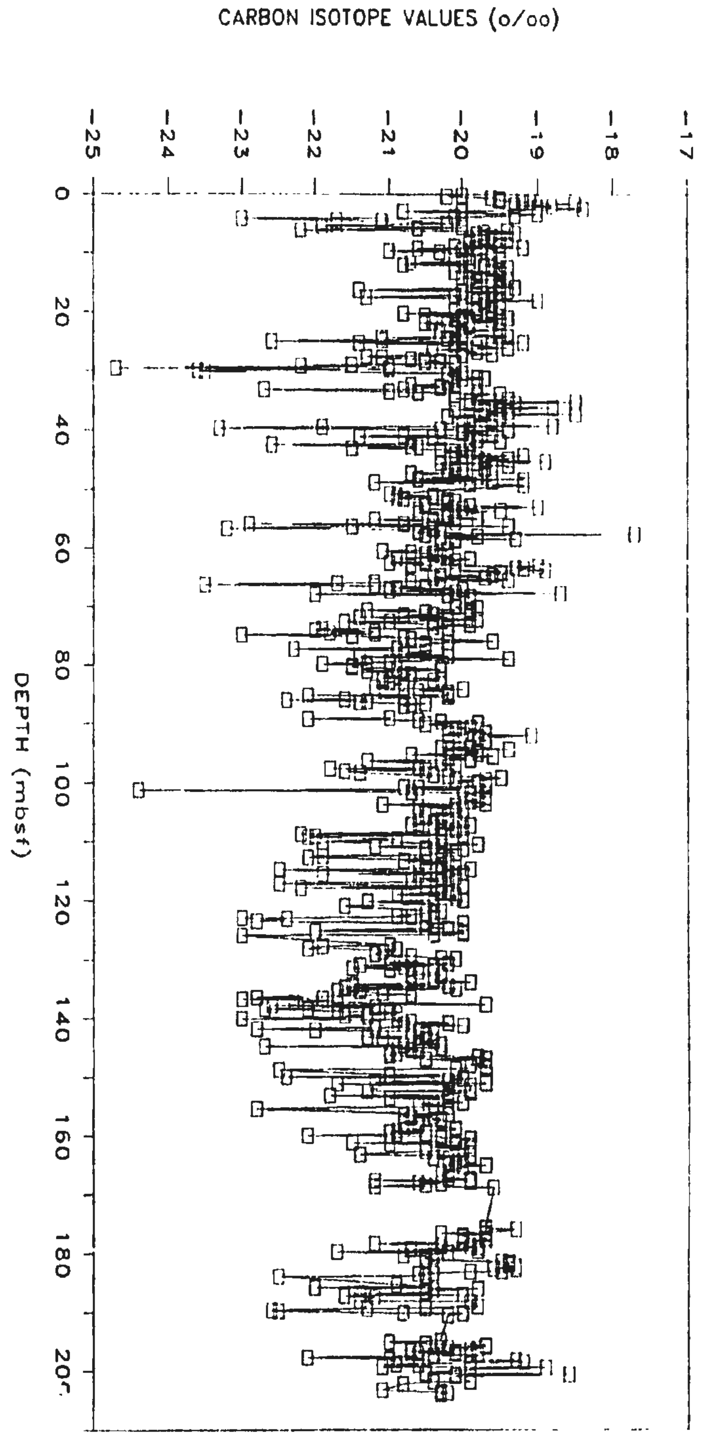

CARBON ISOTOPE VALUES (o/oo)
-25
-24
-23
-22
-21
-20
-19
-18
-17
0
20
40
60
80
100
120
140
160
180
200
DEPTH (mbsf)

Figure 5: Variation of nitrogen isotope values ($^{o}/_{oo}$) with depth at Site 724.

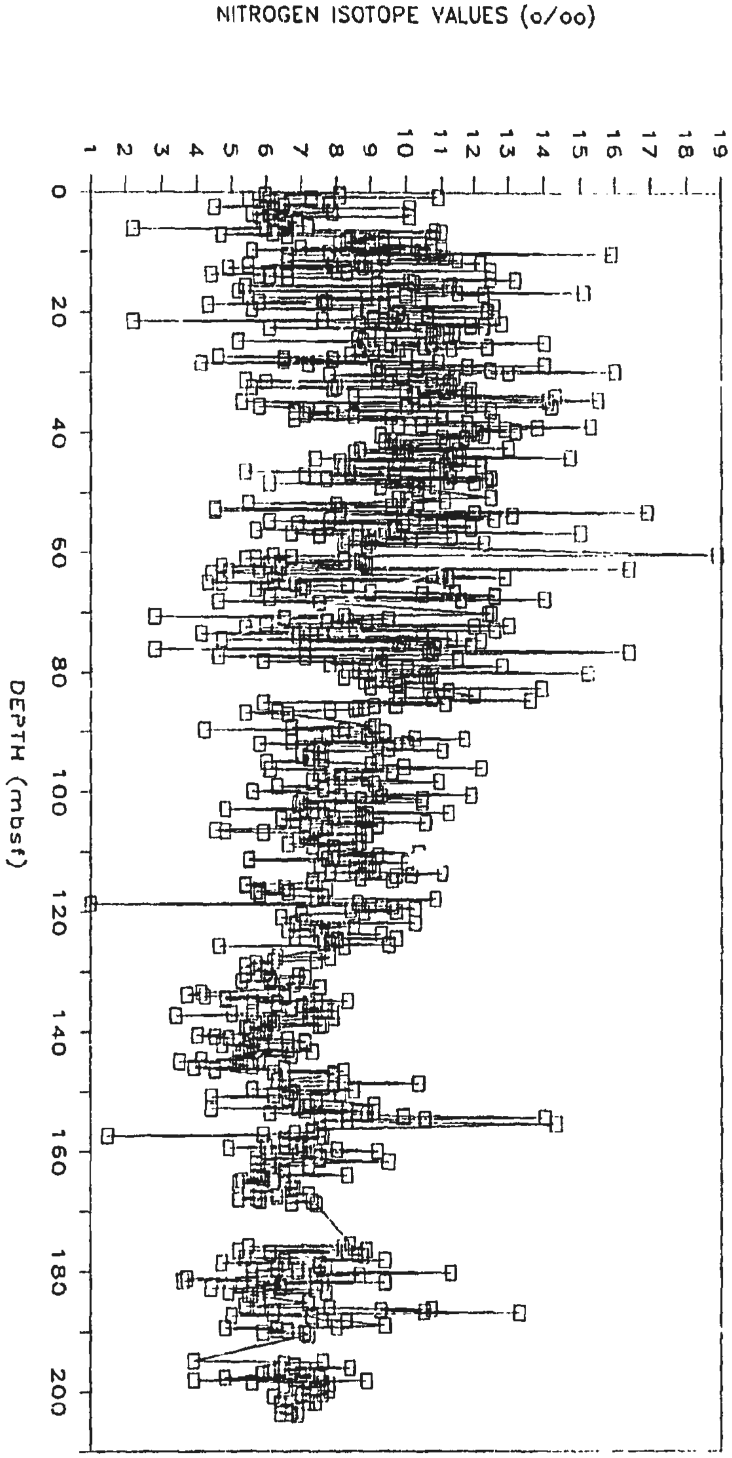

NITROGEN ISOTOPE VALUES (o/oo)
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
0
20
40
60
80
100
120
140
160
180
200
DEPTH (mbsf)

Figure 6: Variation of organic carbon content (%) with depth at Site 724.

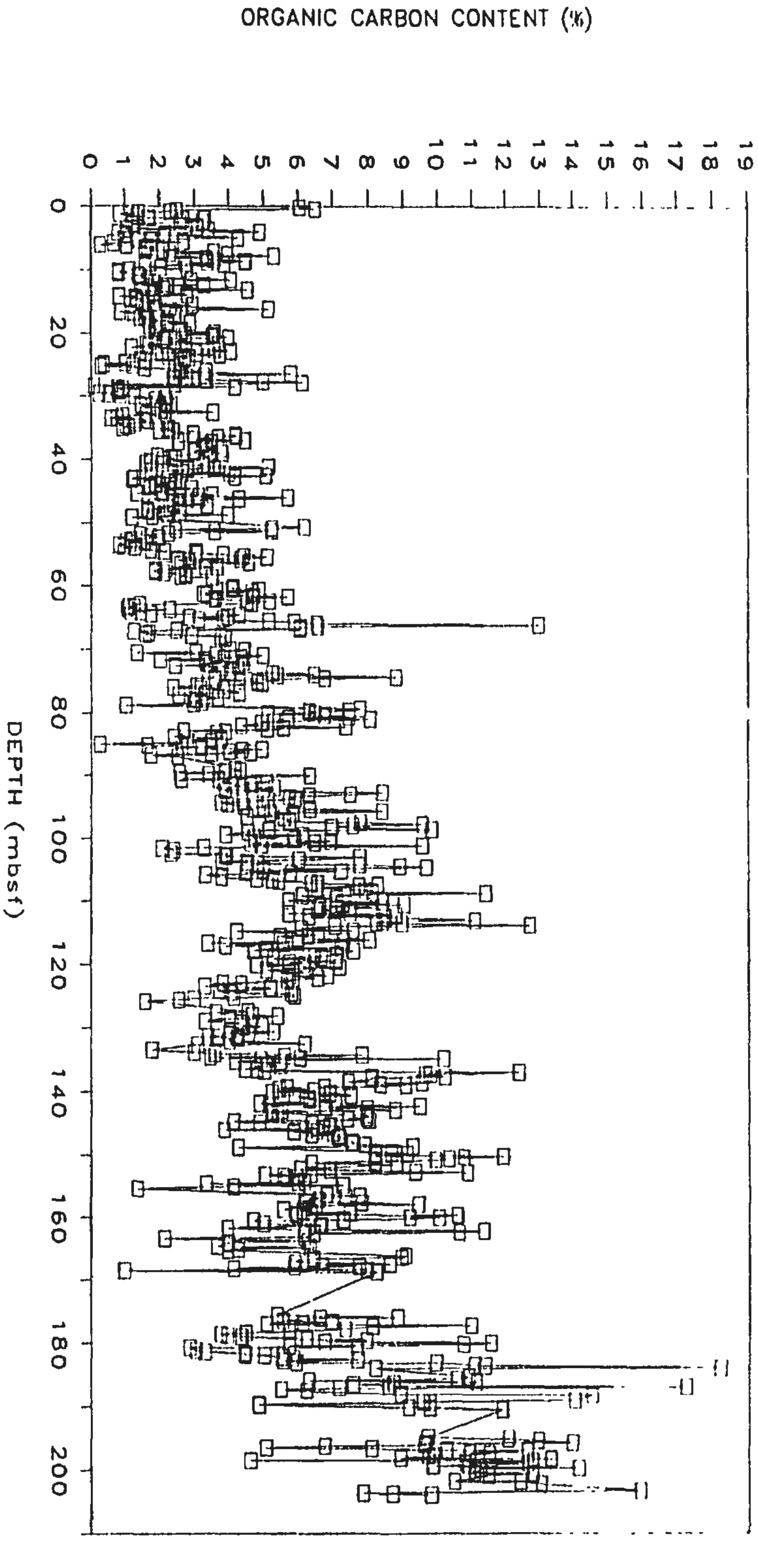

ORGANIC CARBON CONTENT (%)
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
DEPTH (mbsf)
0
20
40
60
80
100
120
140
160
180
200

Figure 7: Relationship between $\delta^{15}N$ and $\delta^{13}C$ of organic matter from Site 724.

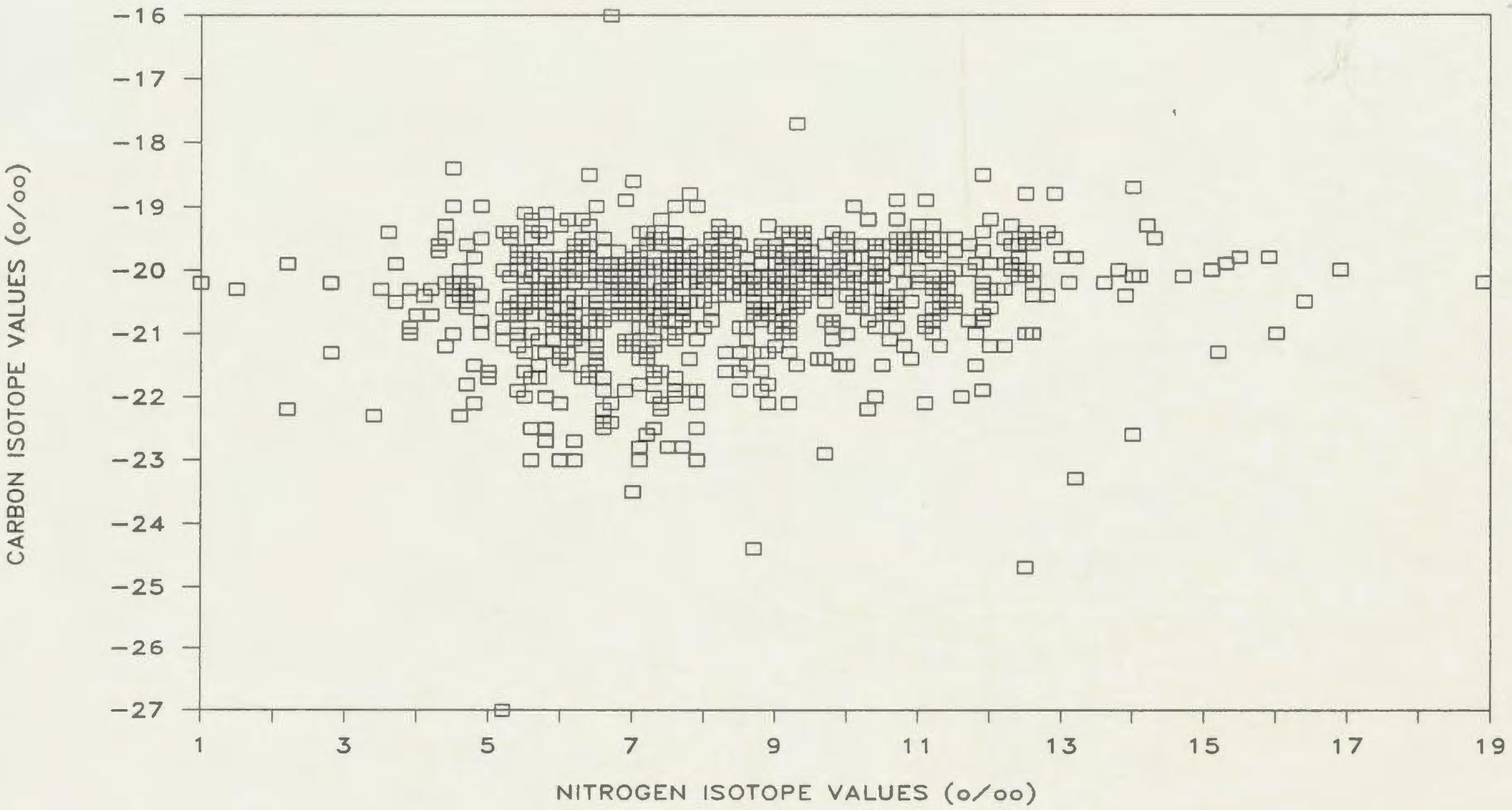

-16
-17
-18
-19
-20
-21
-22
-23
-24
-25
-26
-27
CARBON ISOTOPE VALUES (o/oo)
1
3
5
7
9
11
13
15
17
19
NITROGEN ISOTOPE VALUES (o/oo)

Figure 8: Variation of organic carbon isotope values $(^{o}/_{oo})$ with depth at Site 725.

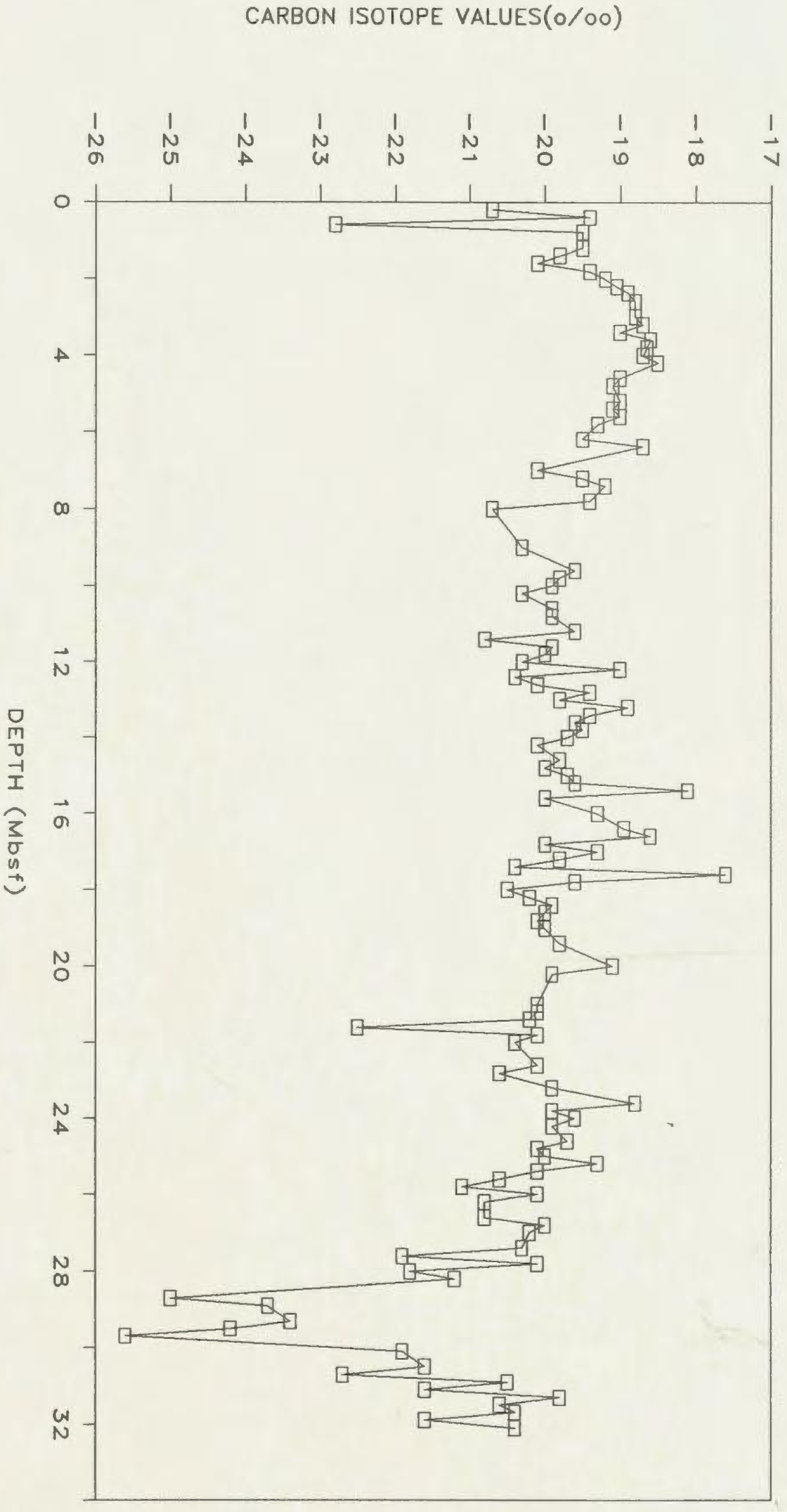

CARBON ISOTOPE VALUES(o/oo)
-17
-18
-19
-20
-21
-22
-23
-24
-25
-26
0
4
8
12
16
20
24
28
32
DEPTH (Mbsf)

Figure 9: Variation of nitrogen isotope values ($^{\circ}/_{\infty}$) with depth at Site 725.

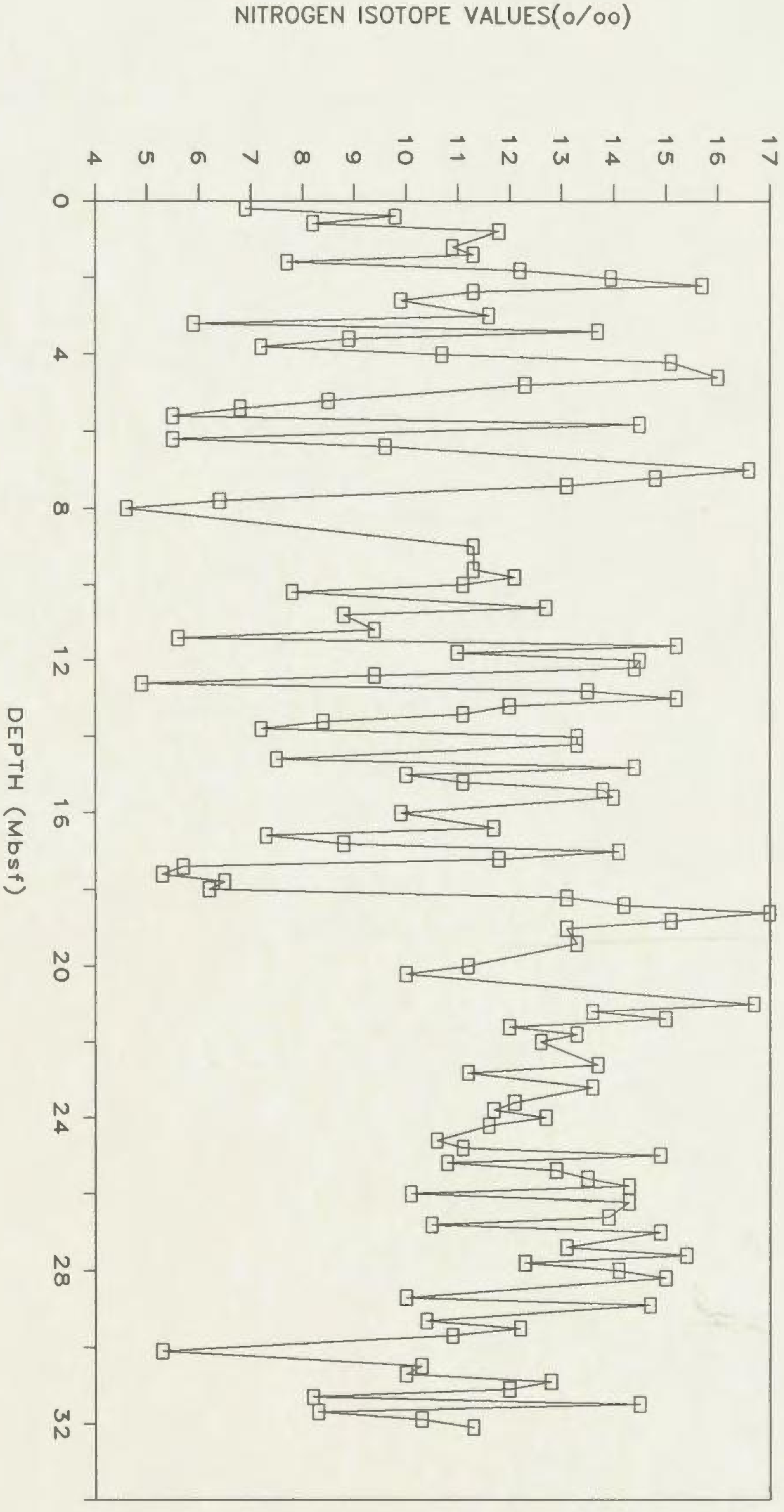

NITROGEN ISOTOPE VALUES(o/oo)
4
5
6
7
8
9
10
11
12
13
14
15
16
17
DEPTH (Mbsf)
0
4
8
12
16
20
24
28
32

Figure 10: Variation of organic carbon content (%) with depth at Site 725.

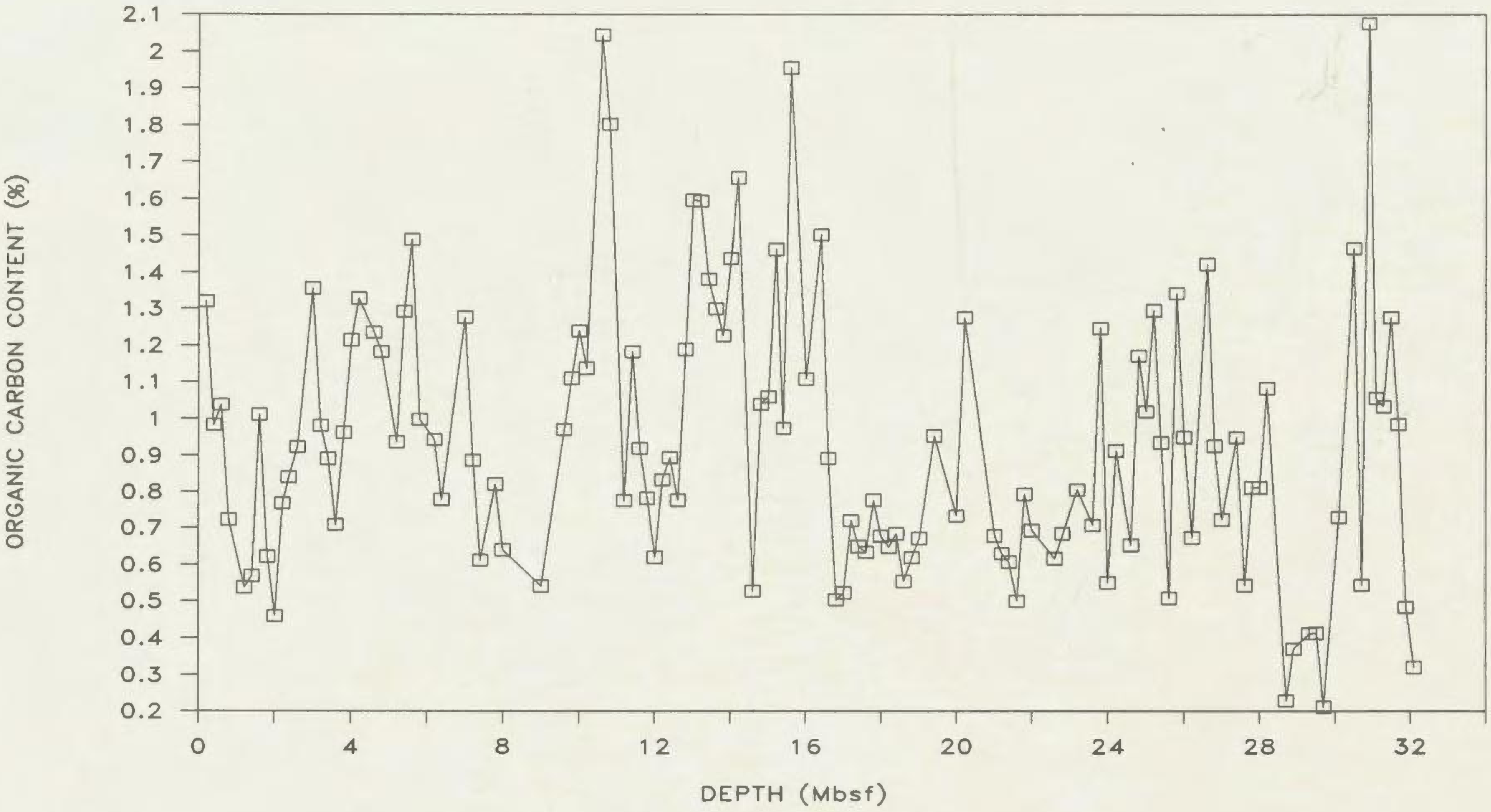

ORGANIC CARBON CONTENT (%)
2.1
2
1.9
1.8
1.7
1.6
1.5
1.4
1.3
1.2
1.1
1
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0
4
8
12
16
20
24
28
32
DEPTH (Mbsf)

Figure 11: Relationship between $\delta^{15}N$ and $\delta^{13}C$ of organic matter from Site 725.

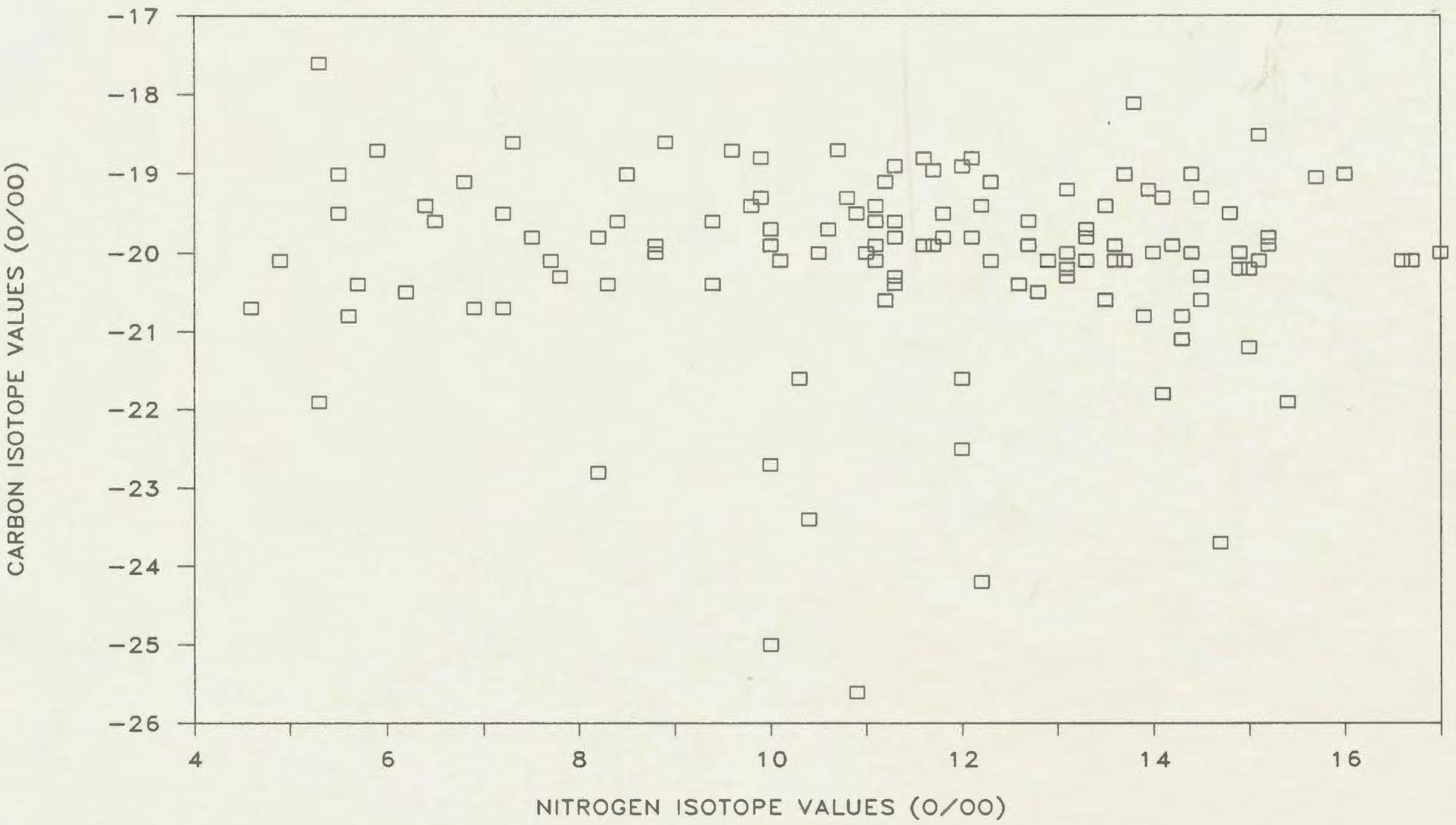

CARBON ISOTOPE VALUES (O/OO)
-17
-18
-19
-20
-21
-22
-23
-24
-25
-26
4
6
8
10
12
14
16
NITROGEN ISOTOPE VALUES (O/OO)

Figure 12: Plot of smoothened organic carbon isotope values ($^{o}/_{oo}$) versus depth at Site 724.

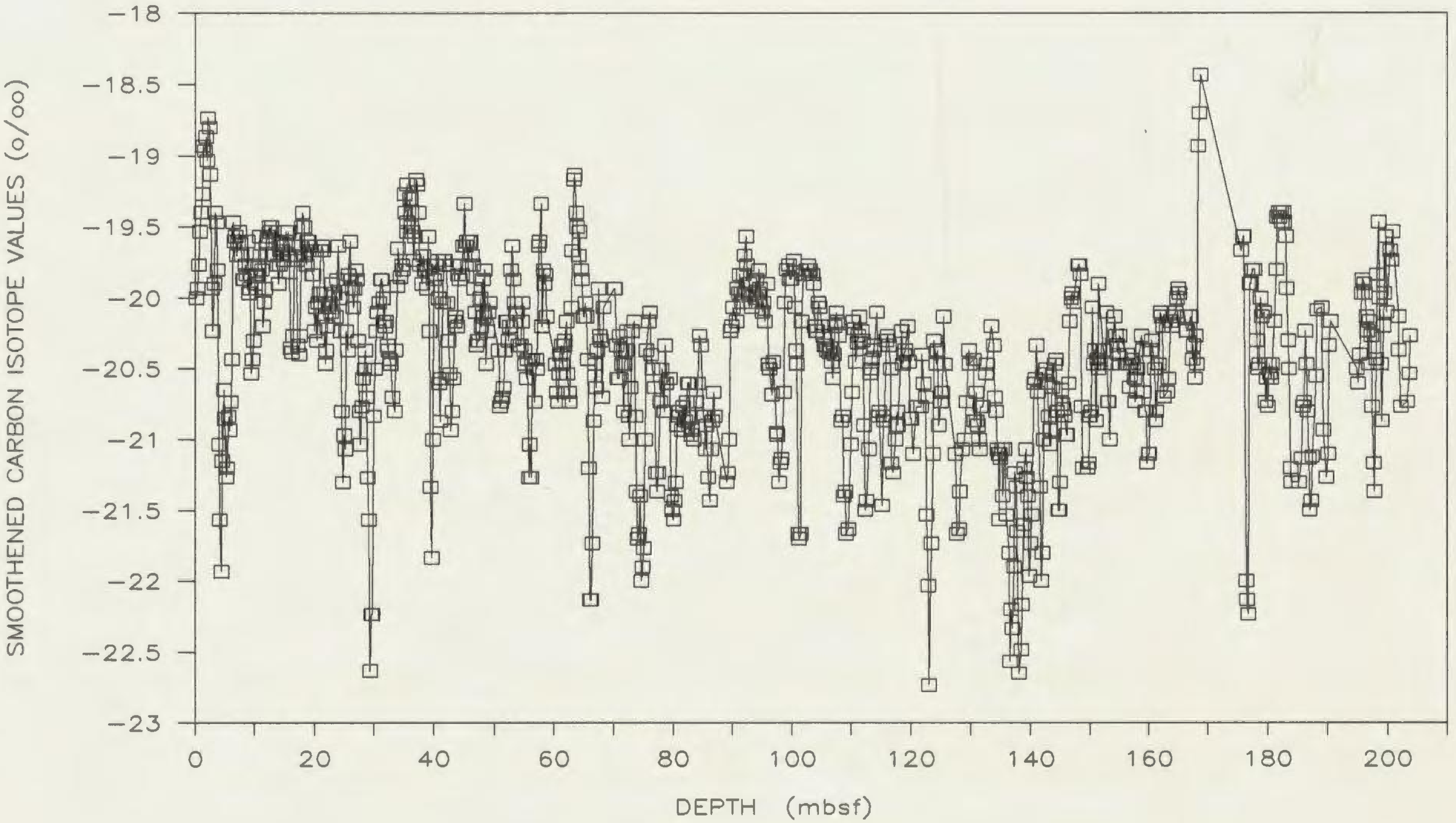

SMOOTHENED CARBON ISOTOPE VALUES (o/oo)
-18
-18.5
-19
-19.5
-20
-20.5
-21
-21.5
-22
-22.5
-23
0
20
40
60
80
100
120
140
160
180
200
DEPTH (mbsf)

Figure 13: Plot of smoothened nitrogen isotope values ($^{o}/_{oo}$) versus depth at Site 724.

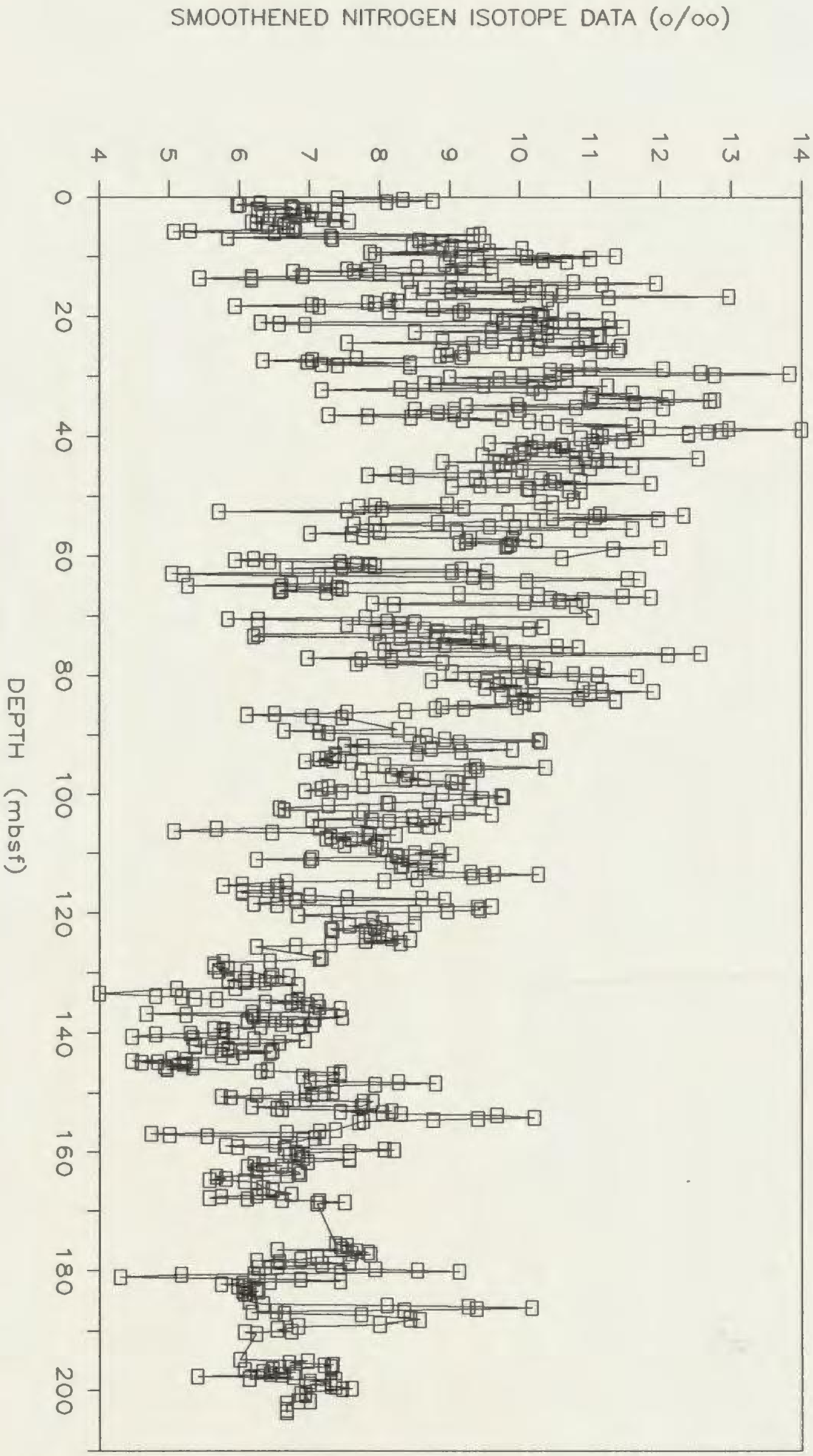
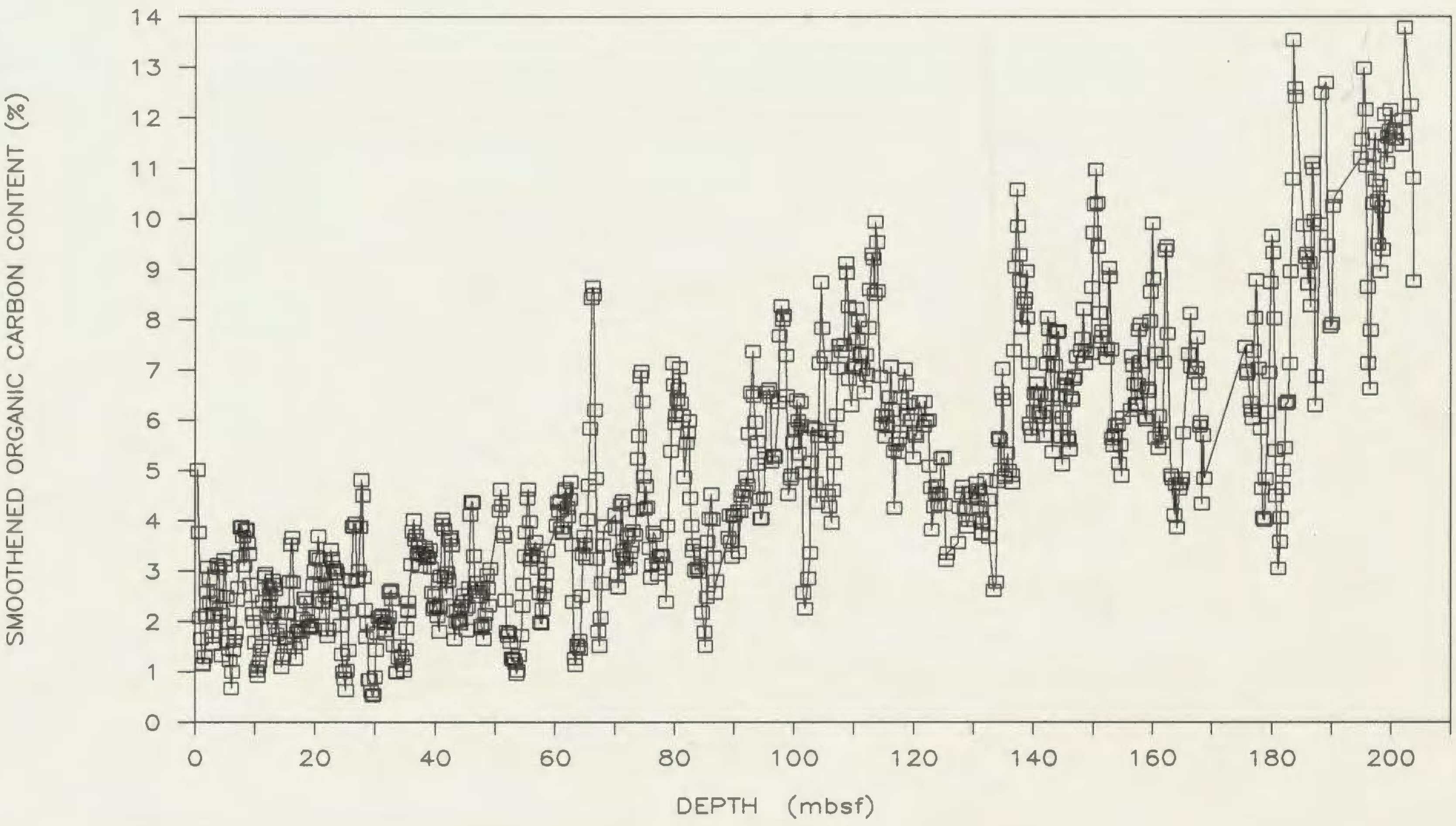

SMOOTHENED NITROGEN ISOTOPE DATA (o/oo)
4
5
6
7
8
9
10
11
12
13
14
0
20
40
60
80
100
120
140
160
180
200
DEPTH (mbsf)

Figure 14: Plot of smoothened organic carbon content (%) versus depth at Site 724.



SMOOTHENED ORGANIC CARBON CONTENT (%)
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
0
20
40
60
80
100
120
140
160
180
200
DEPTH (mbsf)

Figure 15: Plot of smoothened organic carbon isotope values ($^{\circ}/_{\circ\circ}$) versus depth at Site 725.

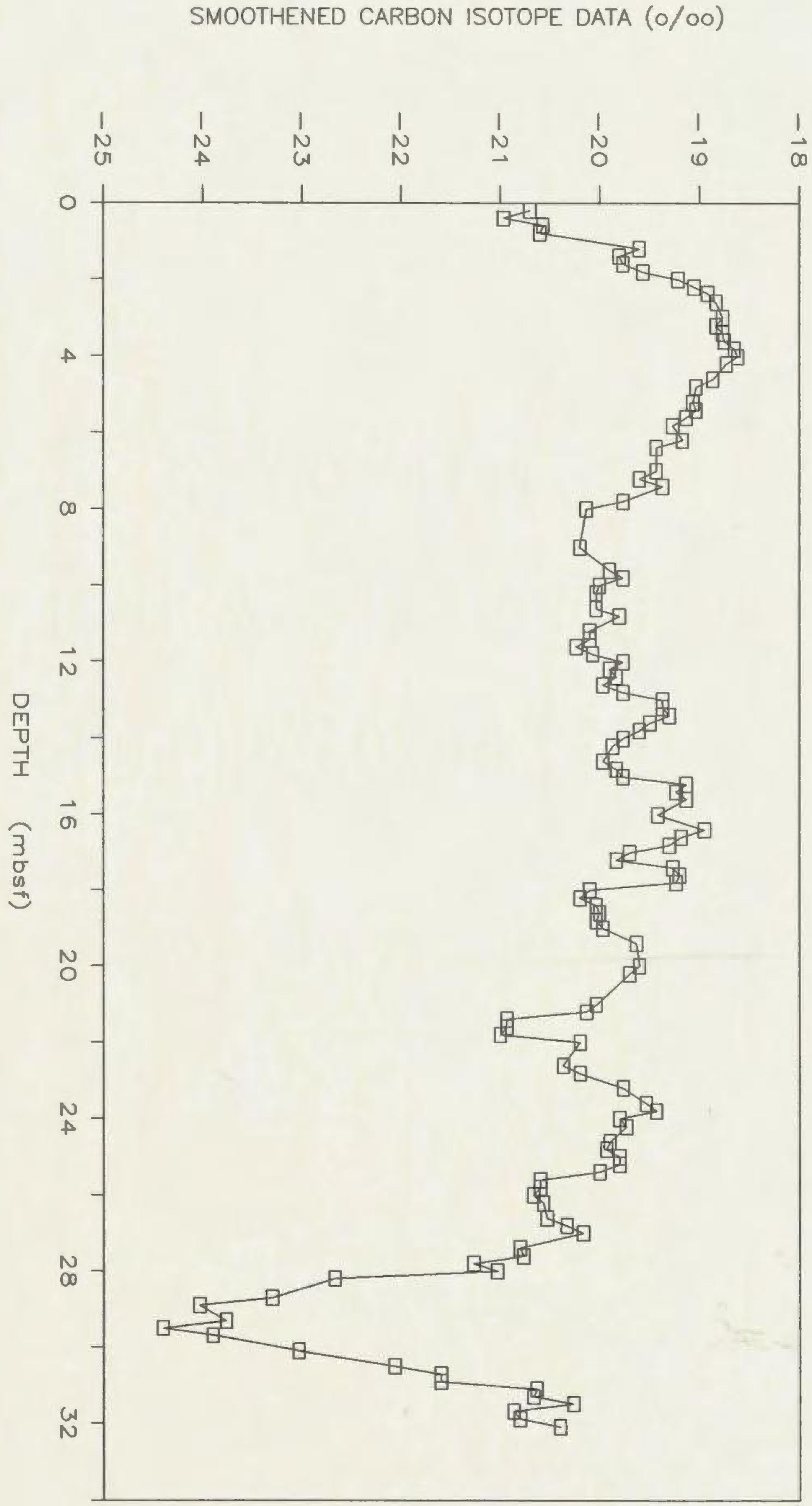

SMOOTHENED CARBON ISOTOPE DATA (o/oo)
-18
-19
-20
-21
-22
-23
-24
-25
0
4
8
12
16
20
24
28
32
DEPTH (mbsf)

Figure 16: Plot of smoothened nitrogen isotope values ($^{o}/_{oo}$) versus depth at Site 725.

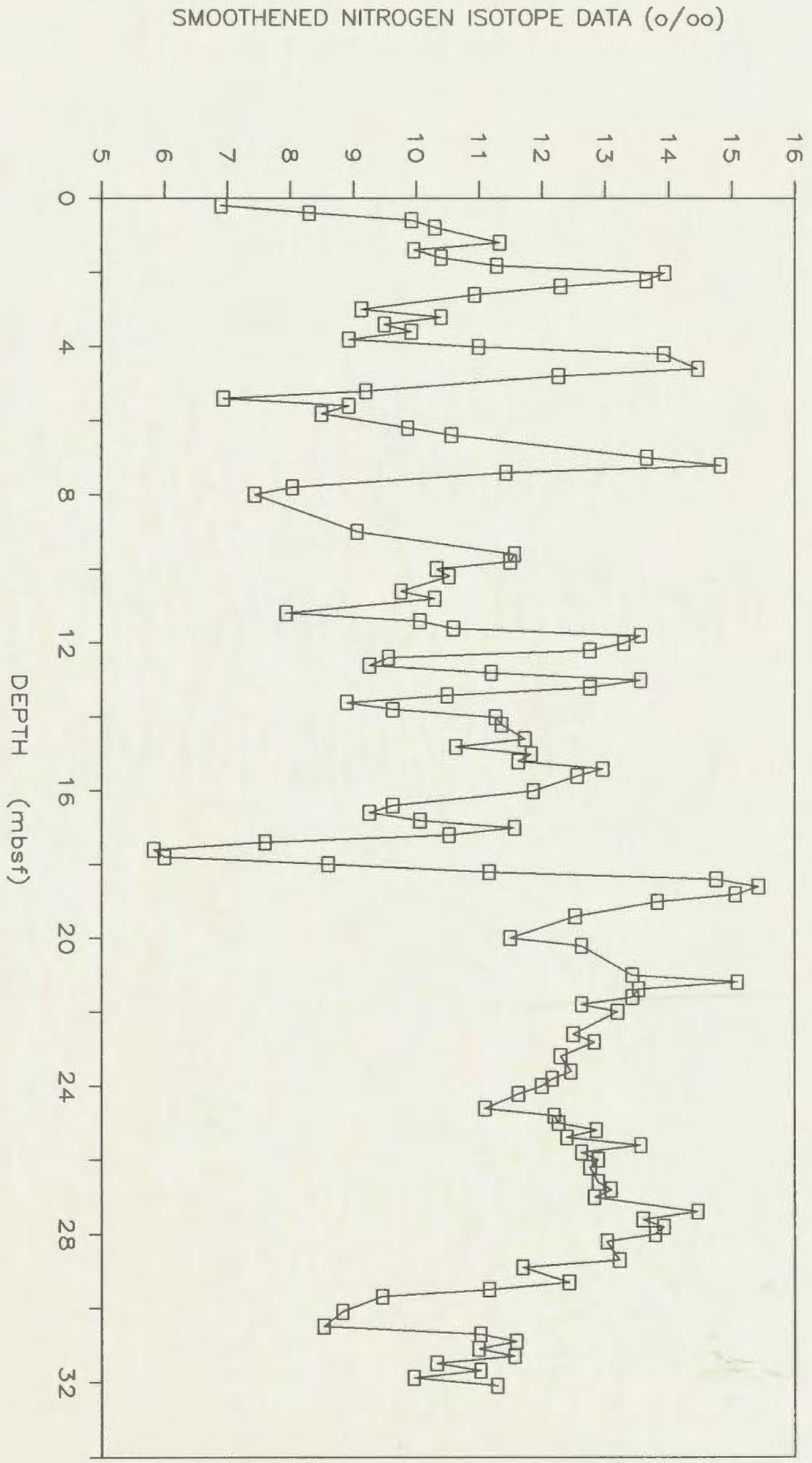

SMOOTHENED NITROGEN ISOTOPE DATA (o/oo)
5
6
7
8
9
10
11
12
13
14
15
16
0
4
8
12
16
20
24
28
32
DEPTH (mbsf)

Figure 17: Plot of smoothened organic carbon content (%) versus depth at Site 725.

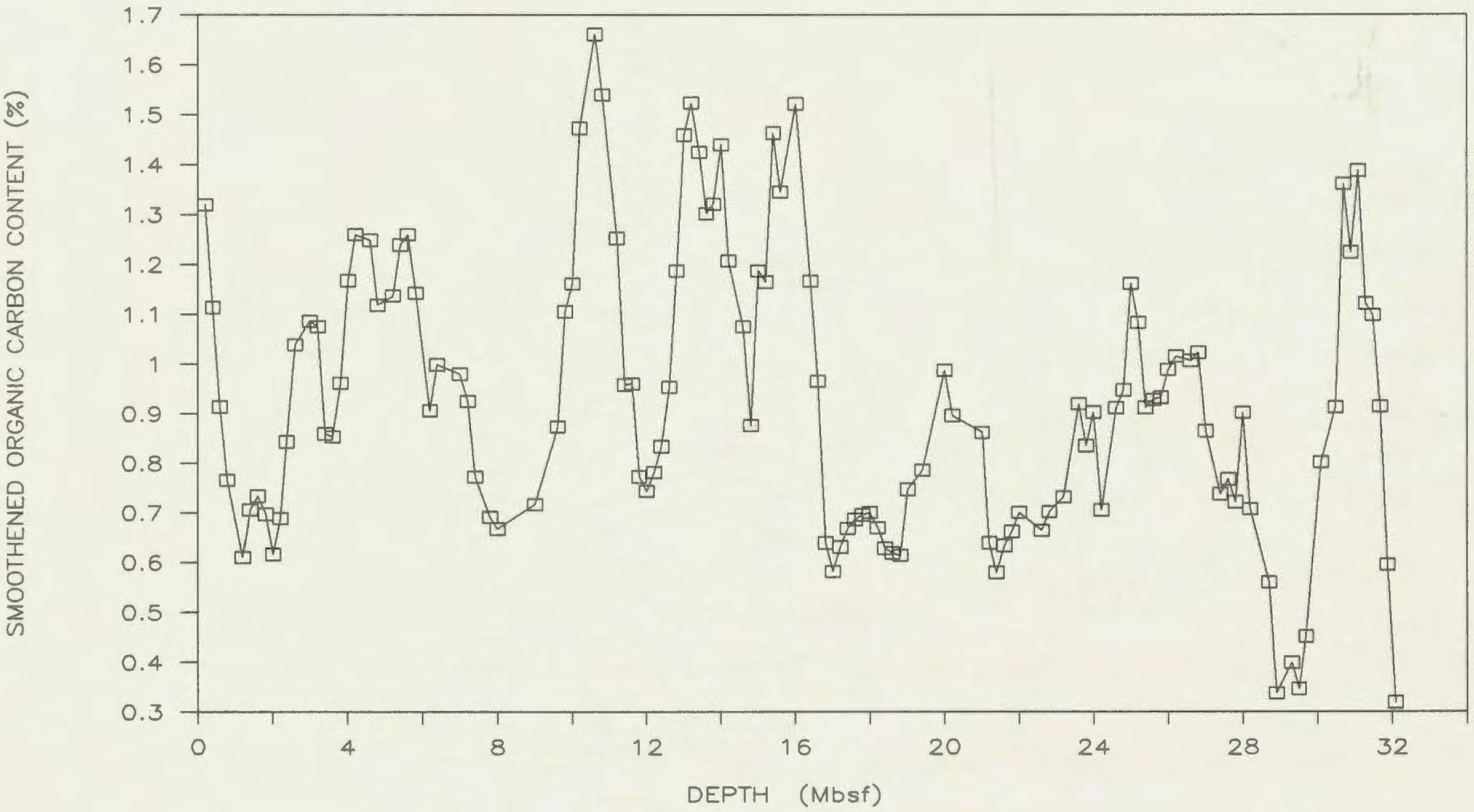

SMOOTHENED ORGANIC CARBON CONTENT (%)
1.7
1.6
1.5
1.4
1.3
1.2
1.1
1
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0
4
8
12
16
20
24
28
32
DEPTH (Mbsf)

Figure 18: Plot of average of three data points of organic carbon isotope values ($^{o}/_{oo}$) versus depth at Site 724.

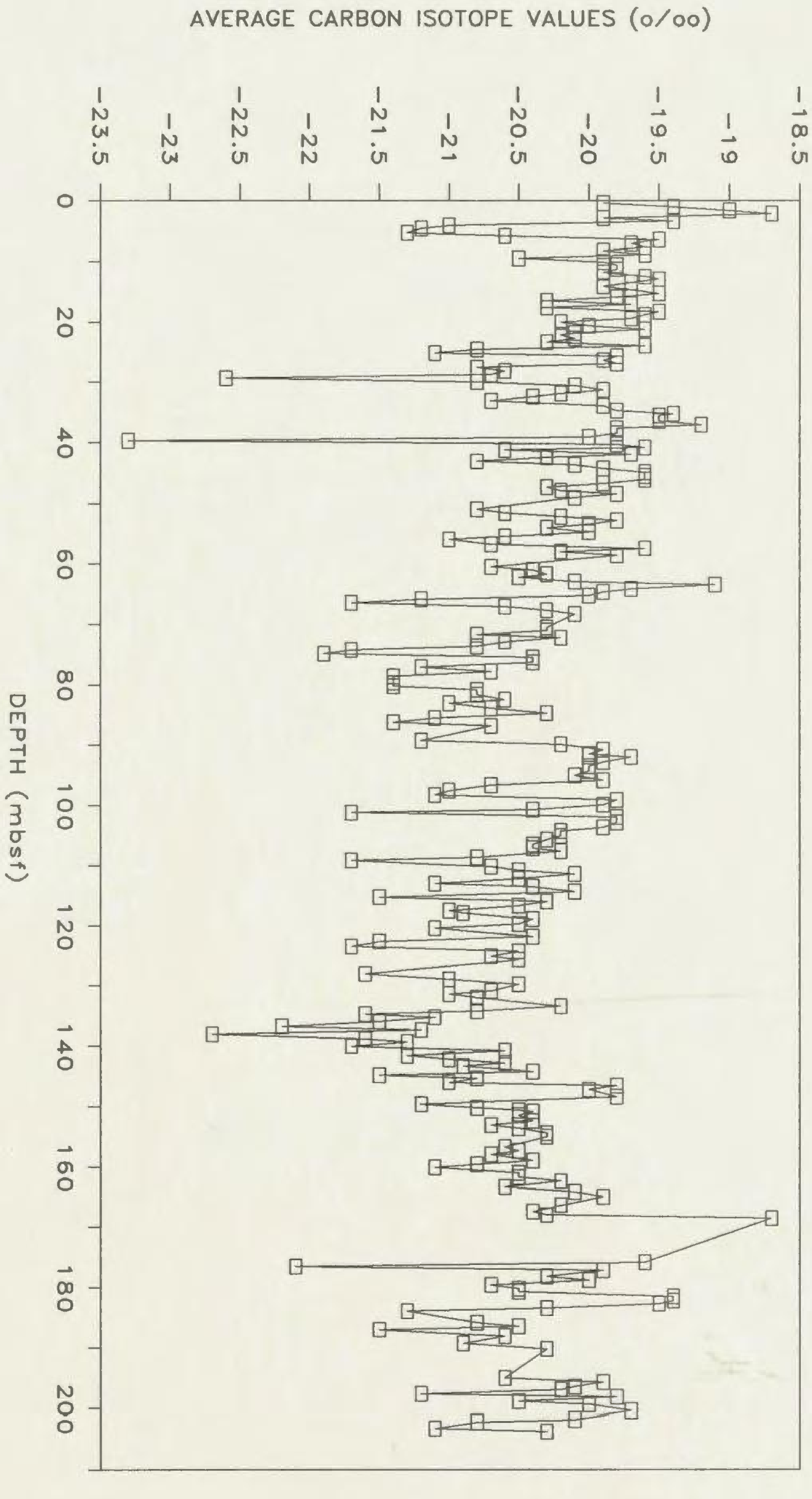

AVERAGE CARBON ISOTOPE VALUES (o/oo)
-23.5
-23
-22.5
-22
-21.5
-21
-20.5
-20
-19.5
-19
-18.5
0
20
40
60
80
100
120
140
160
180
200
DEPTH (mbsf)

Figure 19: Bivariate plot of three data points average of nitrogen isotope values ($^{o}/_{oo}$) against depth at Site 724.

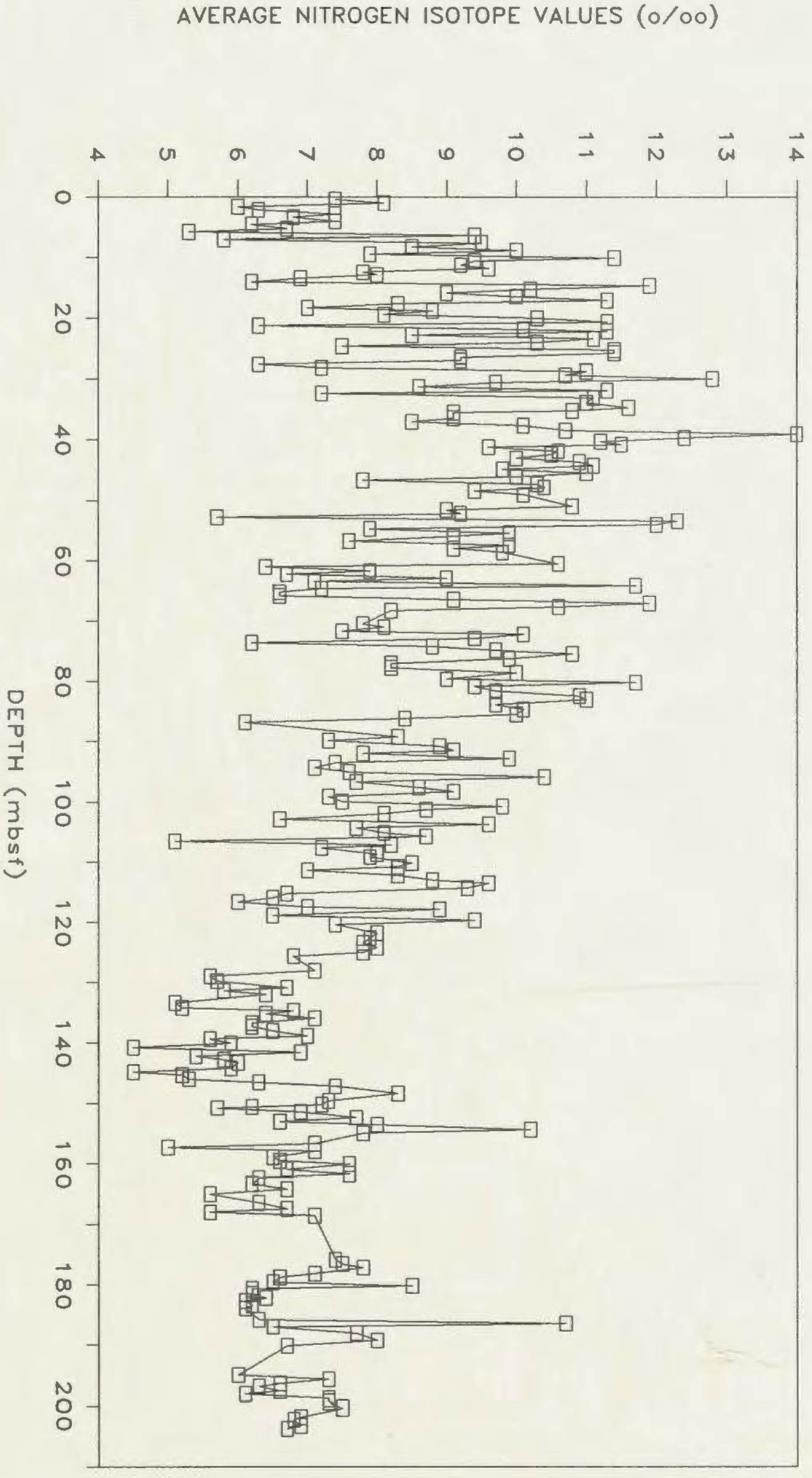

AVERAGE NITROGEN ISOTOPE VALUES (o/oo)
4
5
6
7
8
9
10
11
12
13
14
0
20
40
60
80
100
120
140
160
180
200
DEPTH (mbsf)

Figure 20: Plot of three data points average of organic carbon content (%) against depth at Site 724.

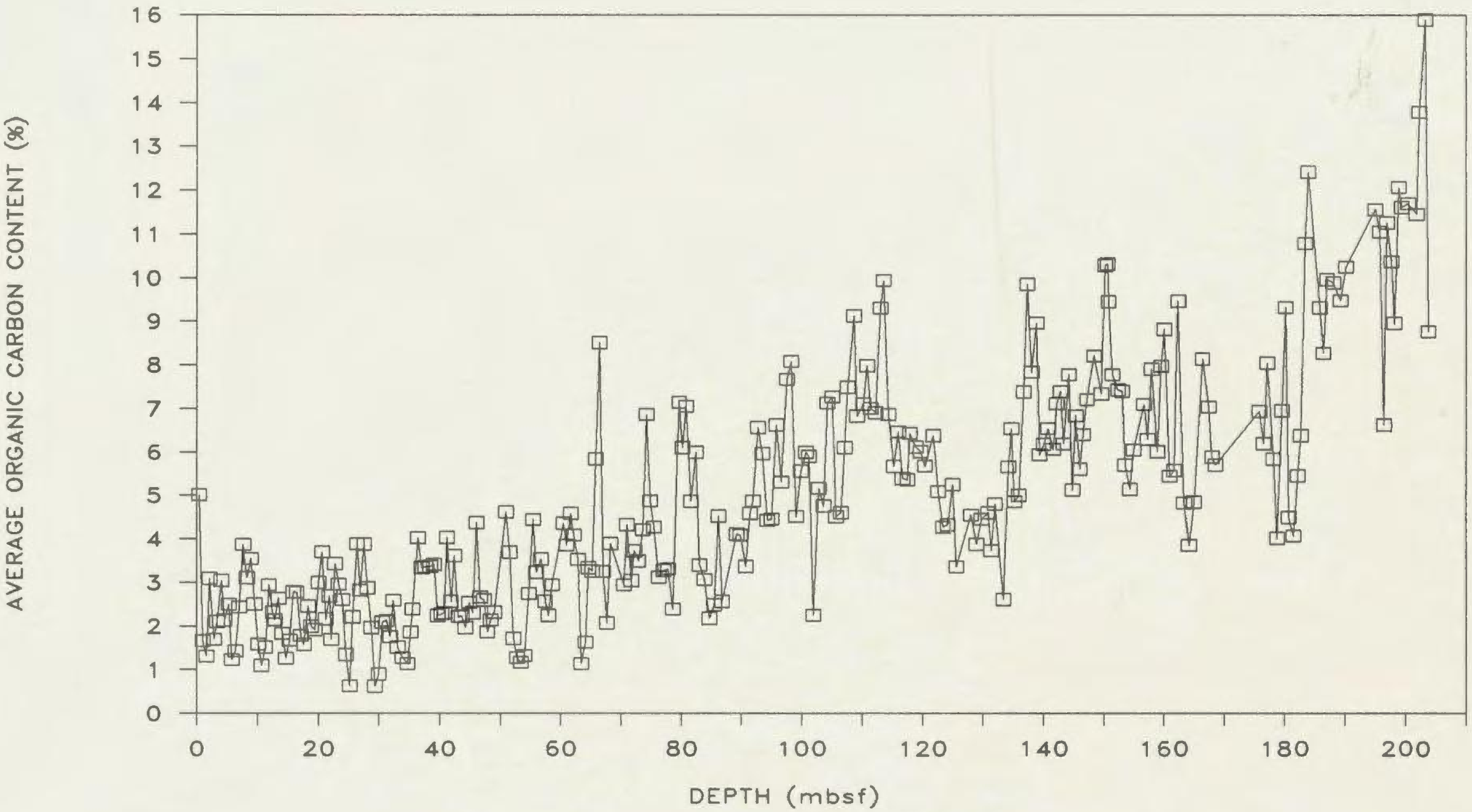

AVERAGE ORGANIC CARBON CONTENT (%)
16
15
14
13
12
11
10
9
8
7
6
5
4
3
2
1
0
0
20
40
60
80
100
120
140
160
180
200
DEPTH (mbsf)

Figure 21: Variation of averaged $\delta^{13}C$ values in 10m interval with depth at Site 724.

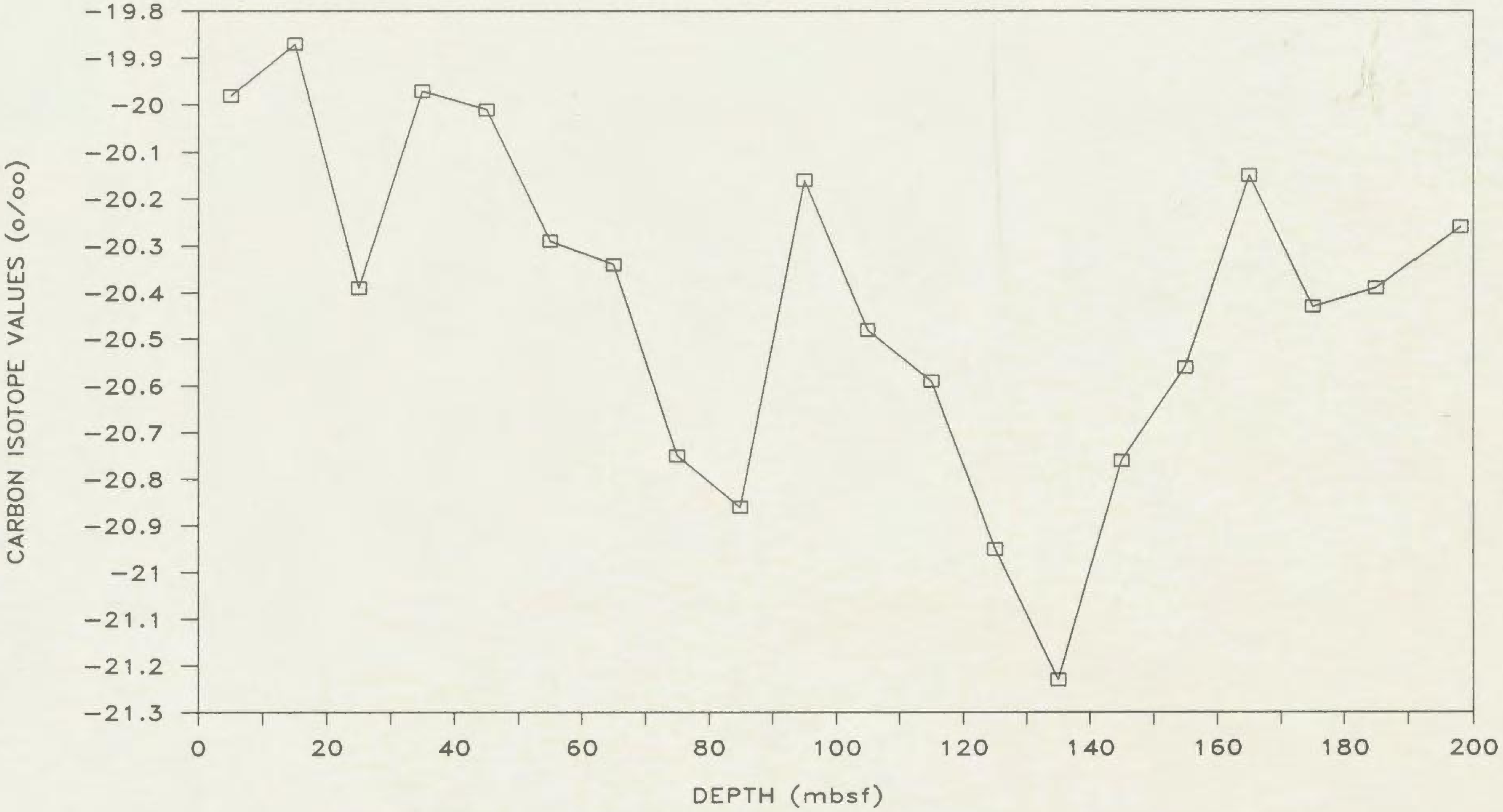

CARBON ISOTOPE VALUES (o/oo)
−19.8
−19.9
−20
−20.1
−20.2
−20.3
−20.4
−20.5
−20.6
−20.7
−20.8
−20.9
−21
−21.1
−21.2
−21.3
0
20
40
60
80
100
120
140
160
180
200
DEPTH (mbsf)

Figure 22: Variation of averaged $\delta^{15}N$ values in 10m interval with depth at Site 724.

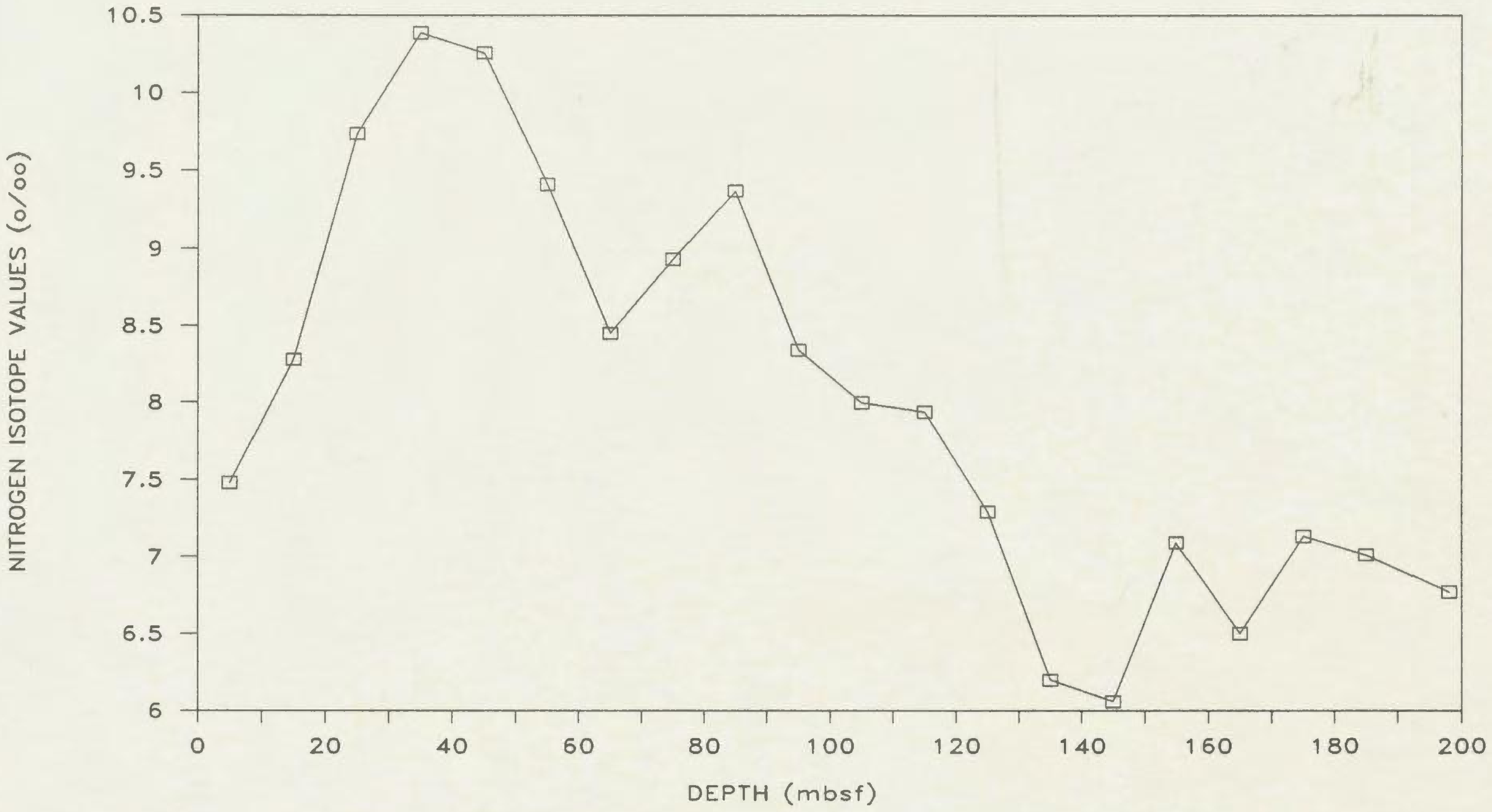

NITROGEN ISOTOPE VALUES (o/oo)
10.5
10
9.5
9
8.5
8
7.5
7
6.5
6
0
20
40
60
80
100
120
140
160
180
200
DEPTH (mbsf)

Figure 23: Variation of averaged organic carbon (%) in 10m interval with depth at Site 724.

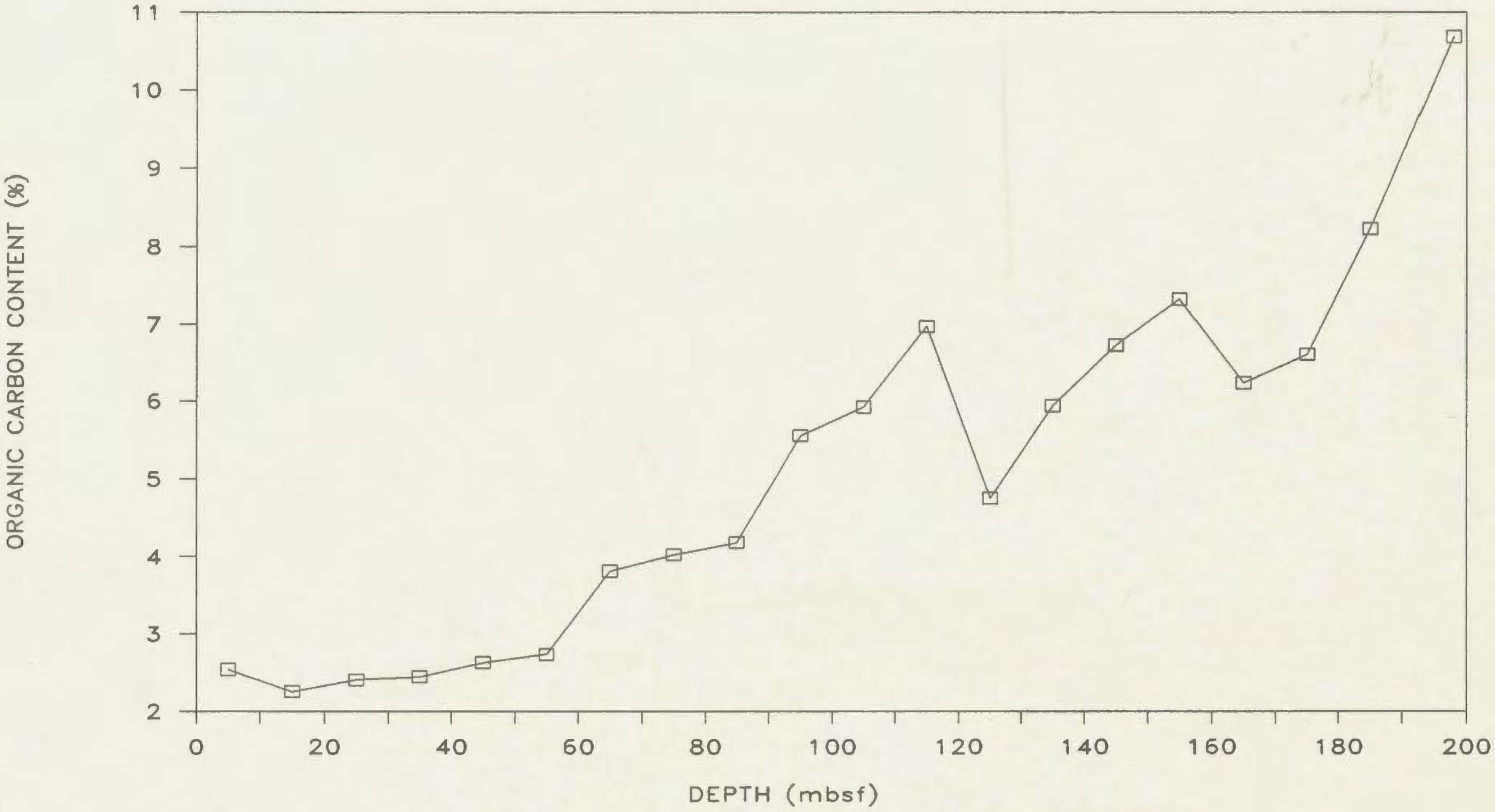

ORGANIC CARBON CONTENT (%)
11
10
9
8
7
6
5
4
3
2
0
20
40
60
80
100
120
140
160
180
200
DEPTH (mbsf)

Figure 24: Plot of smoothened organic carbon isotope values ($^{o}/_{oo}$) versus depth at Site 724. Even numbers indicate glacial periods and odd numbers indicate interglacial periods.

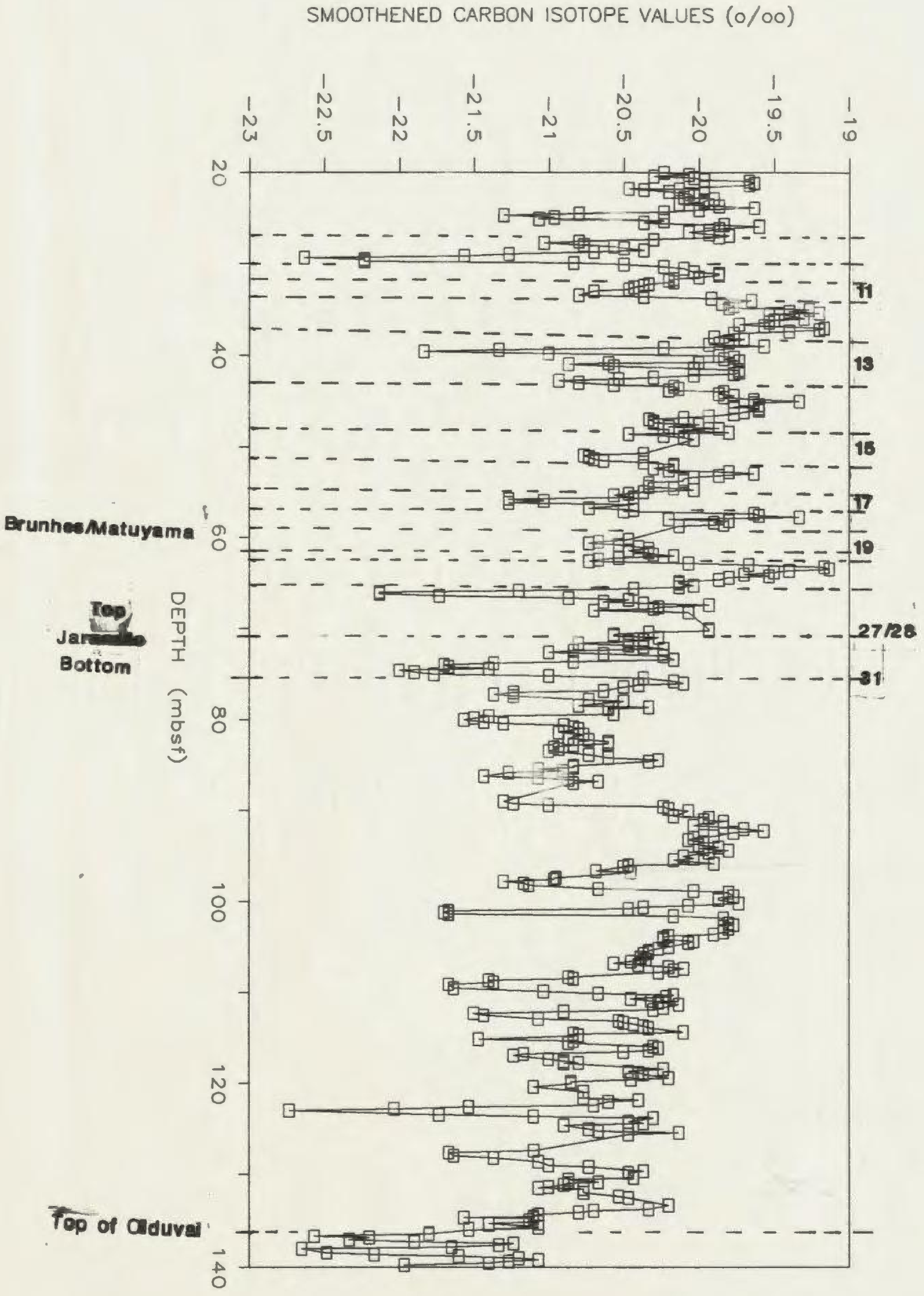

SMOOTHENED CARBON ISOTOPE VALUES (o/oo)
-23
-22.5
-22
-21.5
-21
-20.5
-20
-19.5
-19
DEPTH (mbsf)
20
40
60
80
100
120
140
11
13
15
17
19
27/28
31
Brunhes/Matuyama
Top
Jaramillo
Bottom
Top of Olduvai

Figure 25: Plot of average of three data points of organic carbon isotope values (${}^{\circ}/_{0}$) versus depth at Site 724. Even and Odd numbers indicate glacial and interglacial periods respectively.

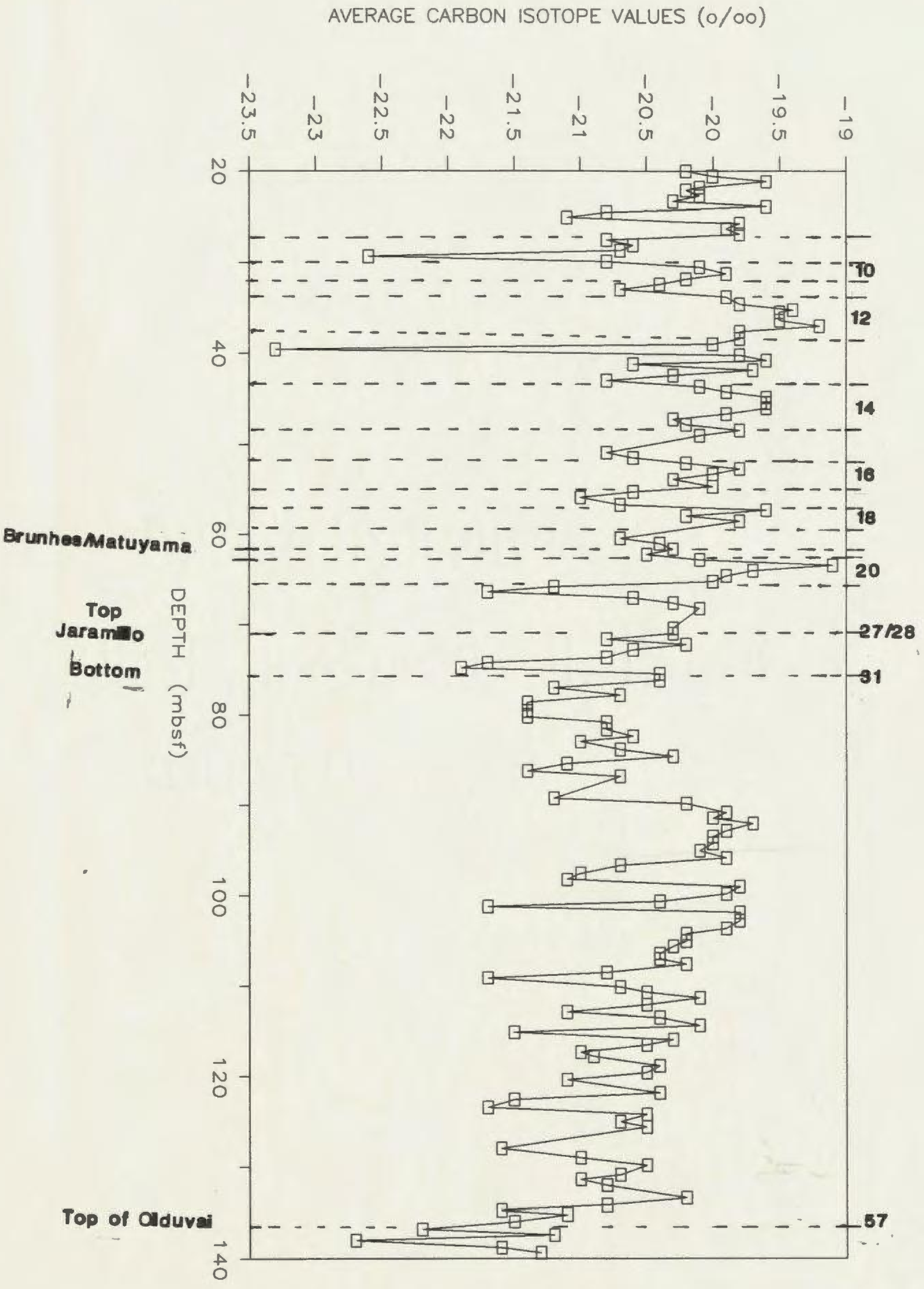

AVERAGE CARBON ISOTOPE VALUES (o/oo)
-23.5
-23
-22.5
-22
-21.5
-21
-20.5
-20
-19.5
-19
DEPTH (mbsf)
20
40
60
80
100
120
140
Brunhes/Matuyama
Top
Jaramillo
Bottom
Top of Olduvai
10
12
14
16
18
20
27/28
31
57

Figure 26: Plot of total organic carbon content (%) versus depth for Site 724 (data from Zahn and Pedersen, in Press). Odd and even numbers indicate interglacial and glacial periods respectively.

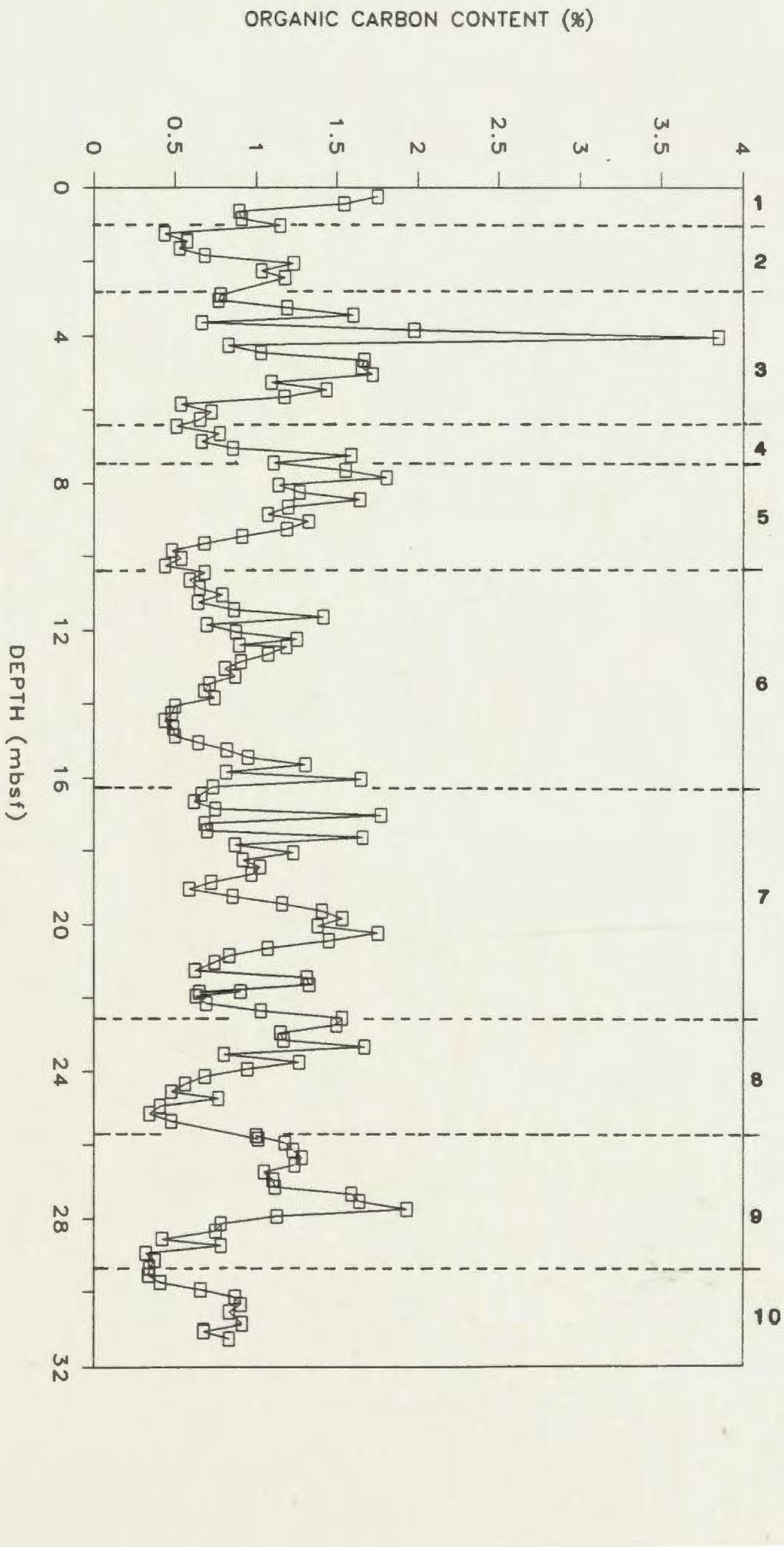

ORGANIC CARBON CONTENT (%)
0
0.5
1
1.5
2
2.5
3
3.5
4
DEPTH (mbsf)
0
4
8
12
16
20
24
28
32
1
2
3
4
5
6
7
8
9
10

Figure 27: Plot of corrected organic carbon content (%) for carbonate content at Site 724.

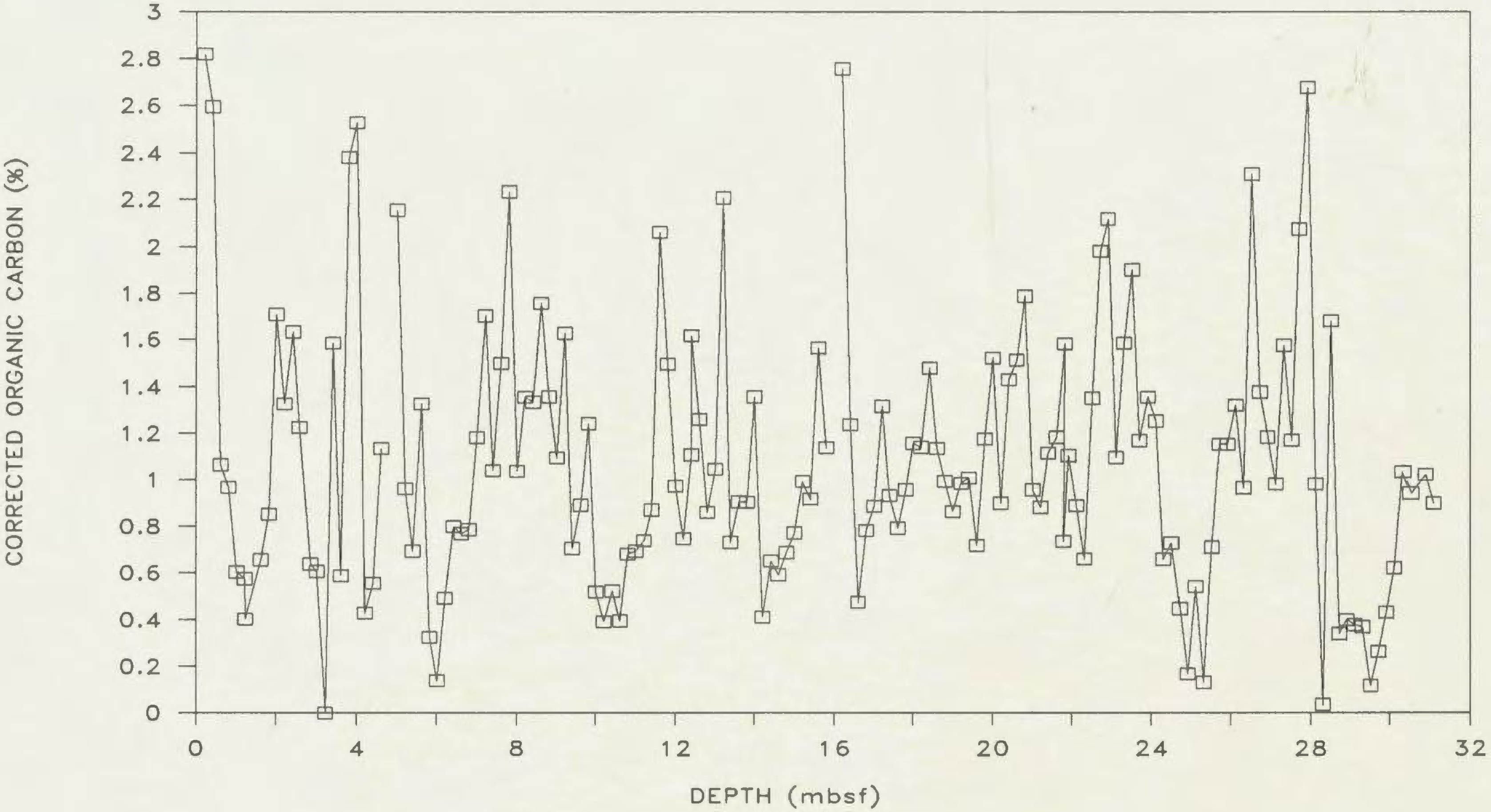

CORRECTED ORGANIC CARBON (%)
3
2.8
2.6
2.4
2.2
2
1.8
1.6
1.4
1.2
1
0.8
0.6
0.4
0.2
0
0
4
8
12
16
20
24
28
32
DEPTH (mbsf)

Figure 28: Plot of total organic nitrogen content (%) against depth for Site 724 (data from Zahn and Pedersen, in press).

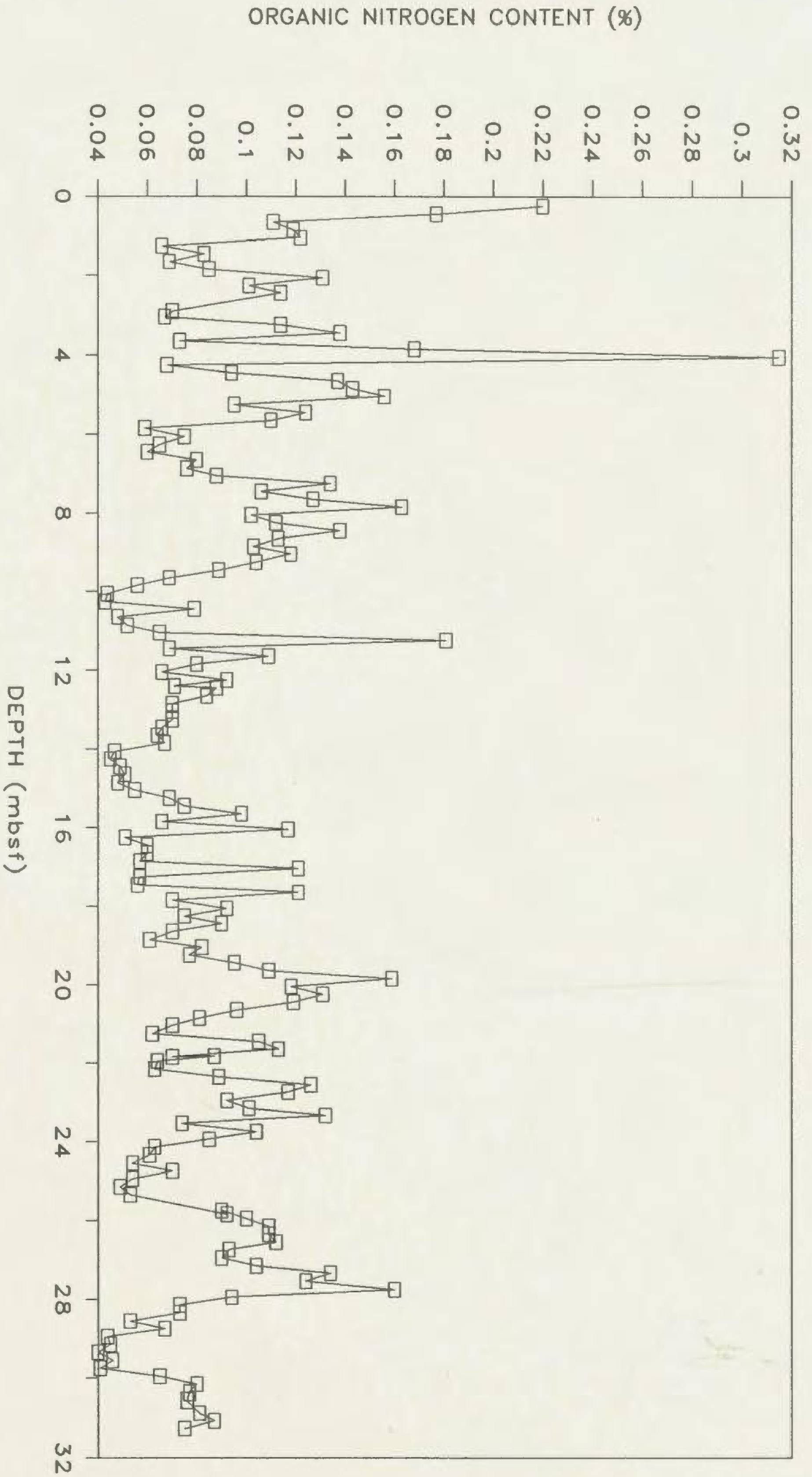

ORGANIC NITROGEN CONTENT (%)
0.04
0.06
0.08
0.1
0.12
0.14
0.16
0.18
0.2
0.22
0.24
0.26
0.28
0.3
0.32
DEPTH (mbsf)
0
4
8
12
16
20
24
28
32

Figure 29: Plot of C/N ratios versus depth for Site 724 (data from Zahn and Pedersen, in Press).

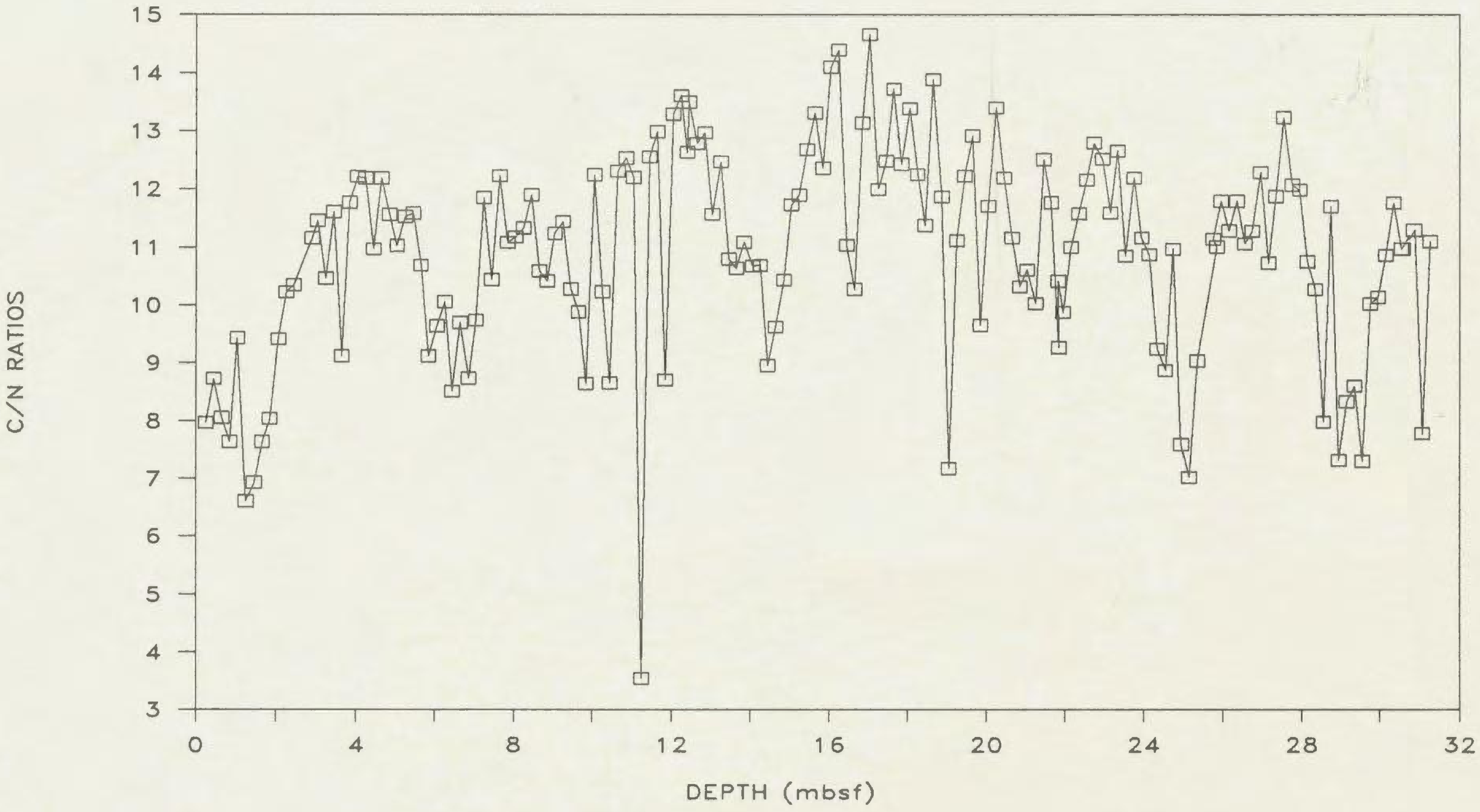

C/N RATIOS
15
14
13
12
11
10
9
8
7
6
5
4
3
0
4
8
12
16
20
24
28
32
DEPTH (mbsf)

Table 1: Stable Carbon and Nitrogen Isotope Compositions and Residue Organic Carbon Content Results for Site 724.

Leg	Hole	Core	Depth	$\delta^{15}N$	$\delta^{13}C$	%OC
117	724 C	1 H 1	0.22	8.1	-20.0	6.1
117	724 C	1 H 1	0.42	6.0	-20.2	6.5
117	724 C	1 H 1	0.62	8.1	-19.6	2.5
117	724 C	1 H 1	0.82	10.9	-19.5	2.3
117	724 C	1 H 1	1.02	7.3	-19.5	1.4
117	724 C	1 H 1	1.22	6.1	-19.2	1.2
117	724 C	1 H 1	1.24	5.5	-19.1	0.8
117	724 C	1 H 2	1.62	6.4	-18.5	1.4
117	724 C	1 H 2	1.82	6.0	-19.3	1.7
117	724 C	1 H 2	2.02	7.8	-18.8	3.3
117	724 C	1 H 2	2.22	6.5	-19.0	2.7
117	724 C	1 H 2	2.42	4.5	-18.4	3.3
117	724 C	1 H 2	2.58	10.1	-19.0	2.6
117	724 C	2 H 1	2.87	6.2	-20.0	1.3
117	724 C	2 H 1	3.02	5.9	-20.8	1.2
117	724 C	2 H 1	3.42	7.9	-19.0	3.1
117	724 C	2 H 1	3.62	5.6	-19.3	1.1
117	724 C	2 H 1	3.82	5.8	-20.1	3.4
117	724 C	2 H 1	4.02	10.1	-20.0	4.9

117	724 C	2 H 1	4.22	6.2	-23.0	0.8
117	724 C	2 H 2	4.42	6.4	-21.7	1.0
117	724 C	2 H 2	4.62	5.9	-21.1	2.2
117	724 C	2 H 2	5.02	6.9	-20.2	4.3
117	724 C	2 H 2	5.22	6.6	-21.7	1.8
117	724 C	2 H 2	5.42	6.6	-21.9	1.4
117	724 C	2 H 2	5.62	7.2	-20.0	2.7
117	724 C	2 H 3	5.82	6.5	-20.6	0.7
117	724 C	2 H 3	6.02	2.2	-22.2	0.3
117	724 C	2 H 3	6.22	6.0	-19.4	1.1
117	724 C	2 H 3	6.42	10.8	-19.7	1.7
117	724 C	2 H 3	6.62	11.0	-19.3	1.6
117	724 C	2 H 3	6.82	6.2	-19.8	1.6
117	724 C	2 H 3	7.02	4.7	-19.6	2.1
117	724 C	2 H 3	7.22	6.6	-19.7	3.6
117	724 C	2 H 4	7.42	10.7	-19.5	2.3
117	724 C	2 H 4	7.62	8.4	-19.4	4.0
117	724 C	2 H 4	7.82	9.4	-19.9	5.3
117	724 C	2 H 4	8.02	8.3	-19.8	2.4
117	724 C	2 H 4	8.22	9.0	-19.9	3.4
117	724 C	2 H 4	8.42	8.1	-19.9	3.6
117	724 C	2 H 4	8.62	10.0	-19.5	4.5
117	724 C	2 H 5	8.82	9.4	-20.1	3.4
117	724 C	2 H 5	9.02	10.7	-19.2	2.8
117	724 C	2 H 5	9.22	7.0	-20.6	3.9

117	724 C	2	H	5	9.42	11.0	-20.0	1.6
117	724 C	2	H	5	9.62	5.6	-21.0	2.0
117	724 C	2	H	5	9.82	10.4	-20.3	2.8
117	724 C	2	H	5	10.02	7.8	-19.6	1.1
117	724 C	2	H	5	10.22	15.9	-19.8	0.8
117	724 C	2	H	6	10.42	6.6	-20.1	1.1
117	724 C	2	H	6	10.62	10.5	-19.6	0.8
117	724 C	2	H	6	10.82	11.1	-19.6	1.4
117	724 C	2	H	6	11.02	9.4	-19.5	1.4
117	724 C	2	H	6	11.22	11.5	-20.0	1.5
117	724 C	2	H	6	11.42	6.6	-19.9	1.7
117	724 C	2	H	6	11.62	8.7	-20.7	4.1
117	724 C	2	H	7	11.82	12.2	-19.5	2.9
117	724 C	2	H	7	12.02	7.9	-20.8	1.9
117	724 C	2	H	7	12.22	5.5	-19.7	1.5
117	724 C	2	H	7	12.42	9.2	-19.4	2.2
117	724 C	3	H	1	12.42	8.8	-19.6	3.3
117	724 C	3	H	1	12.62	4.9	-19.5	2.6
117	724 C	3	H	1	12.82	6.6	-19.5	1.8
117	724 C	3	H	1	13.02	12.5	-19.5	2.1
117	724 C	3	H	1	13.22	8.0	-20.1	4.6
117	724 C	3	H	1	13.42	8.3	-19.5	1.5
117	724 C	3	H	1	13.62	4.4	-19.5	1.9
117	724 C	3	H	2	13.82	5.8	-19.8	1.9
117	724 C	3	H	2	14.02	6.1	-20.0	2.8

117	724	C	3	H	2	14.22	6.6	-19.9	0.8
117	724	C	3	H	2	14.42	12.5	-19.4	1.3
117	724	C	3	H	2	14.62	13.2	-19.8	1.2
117	724	C	3	H	2	14.82	10.1	-19.8	1.3
117	724	C	3	H	2	15.02	10.2	-19.6	1.5
117	724	C	3	H	3	15.22	9.2	-19.5	1.9
117	724	C	3	H	3	15.42	11.3	-19.5	1.7
117	724	C	3	H	3	15.62	5.4	-19.8	2.9
117	724	C	3	H	3	15.82	11.2	-19.3	1.9
117	724	C	3	H	3	16.22	9.7	-21.4	5.2
117	724	C	3	H	3	16.42	5.2	-19.4	2.3
117	724	C	3	H	3	16.62	15.1	-20.0	0.9
117	724	C	3	H	4	16.82	11.5	-19.5	1.3
117	724	C	3	H	4	17.02	12.3	-19.6	1.6
117	724	C	3	H	4	17.22	10.0	-20.1	2.5
117	724	C	3	H	4	17.42	8.9	-21.3	1.6
117	724	C	3	H	4	17.62	5.5	-19.8	1.5
117	724	C	3	H	4	17.82	10.4	-19.7	1.7
117	724	C	3	H	4	18.02	7.6	-19.0	2.2
117	724	C	3	H	5	18.22	5.8	-19.5	2.2
117	724	C	3	H	5	18.42	7.7	-20.0	2.9
117	724	C	3	H	5	18.62	4.3	-19.6	2.3
117	724	C	3	H	5	18.82	9.4	-19.7	2.0
117	724	C	3	H	5	19.02	12.6	-19.5	1.7
117	724	C	3	H	5	19.22	9.4	-19.6	1.9

117	724 C	3 H 5	19.42	5.6	-19.8	1.9
117	724 C	3 H 5	19.62	9.4	-19.6	1.9
117	724 C	3 H 6	19.82	12.4	-20.1	2.7
117	724 C	3 H 6	20.02	8.6	-19.8	3.6
117	724 C	3 H 6	20.22	9.8	-20.8	2.7
117	724 C	3 H 6	20.42	10.6	-19.6	3.5
117	724 C	3 H 6	20.62	10.6	-20.5	3.6
117	724 C	3 H 6	20.82	12.6	-20.0	4.0
117	724 C	3 H 6	21.02	9.1	-19.4	2.1
117	724 C	3 H 7	21.22	7.6	-19.6	1.9
117	724 C	3 H 7	21.42	2.2	-19.9	2.5
117	724 C	3 H 7	21.62	9.9	-19.5	2.8
117	724 C	3 H 7	21.82	8.7	-20.5	3.7
117	724 C	4 H 1	21.80	12.8	-20.4	1.6
117	724 C	4 H 1	21.92	9.7	-20.5	2.2
117	724 C	4 H 1	22.12	11.9	-20.2	1.7
117	724 C	4 H 1	22.32	12.3	-19.9	1.2
117	724 C	4 H 1	22.52	6.1	-20.1	2.6
117	724 C	4 H 1	22.72	10.4	-20.1	3.6
117	724 C	4 H 1	22.92	9.0	-20.1	4.1
117	724 C	4 H 1	23.12	10.8	-19.5	2.1
117	724 C	4 H 2	23.32	11.4	-20.3	3.1
117	724 C	4 H 2	23.52	11.2	-20.1	3.7
117	724 C	4 H 2	23.72	10.7	-20.0	2.3
117	724 C	4 H 2	23.92	10.9	-19.5	2.9

117	724 C	4 H 2	24.12	9.3	-19.4	2.7
117	724 C	4 H 2	24.32	8.6	-21.1	1.5
117	724 C	4 H 2	24.52	8.8	-20.2	1.6
117	724 C	4 H 3	24.72	5.2	-21.1	1.0
117	724 C	4 H 3	24.92	14.0	-22.6	0.4
117	724 C	4 H 3	25.12	10.7	-19.2	1.2
117	724 C	4 H 3	25.32	9.6	-21.4	0.3
117	724 C	4 H 3	25.52	10.5	-20.1	1.6
117	724 C	4 H 3	25.72	12.4	-19.6	2.4
117	724 C	4 H 3	25.92	11.3	-19.8	2.6
117	724 C	4 H 3	26.12	9.8	-19.4	3.3
117	724 C	4 H 4	26.32	8.7	-20.4	2.6
117	724 C	4 H 4	26.52	9.1	-20.0	5.8
117	724 C	4 H 4	26.72	9.1	-19.8	3.3
117	724 C	4 H 4	26.92	8.4	-20.0	2.8
117	724 C	4 H 4	27.12	10.0	-19.6	2.4
117	724 C	4 H 4	27.32	4.6	-20.0	3.3
117	724 C	4 H 4	27.52	6.5	-21.3	3.3
117	724 C	4 H 5	27.72	7.9	-21.1	5.0
117	724 C	4 H 5	27.92	6.5	-20.7	6.1
117	724 C	4 H 5	28.12	10.9	-20.5	2.4
117	724 C	4 H 5	28.32	4.1	-20.3	0.1
117	724 C	4 H 5	28.52	7.2	-20.5	4.2
117	724 C	4 H 5	28.72	14.0	-20.1	0.8
117	724 C	4 H 5	28.92	11.8	-21.5	0.9

117	724	C	4	H	5	29.12	10.3	-22.2	0.8
117	724	C	4	H	6	29.32	9.2	-21.0	0.8
117	724	C	4	H	6	29.52	12.5	-24.7	0.2
117	724	C	4	H	6	29.72	16.0	-21.0	0.5
117	724	C	4	H	6	29.92	13.0	-23.6	0.9
117	724	C	4	H	6	30.12	9.3	-23.5	1.3
117	724	C	4	H	6	30.32	7.8	-20.0	2.2
117	724	C	4	H	6	30.52	9.9	-20.2	1.9
117	724	C	4	H	7	30.88	11.4	-20.1	2.2
117	724	C	4	H	7	31.08	10.7	-19.8	2.0
117	724	C	4	H	7	31.28	9.2	-19.7	2.2
117	724	C	4	H	7	31.48	6.0	-20.1	2.3
117	724	C	5	H	1	31.28	5.4	-20.0	2.3
117	724	C	5	H	1	31.42	9.5	-20.1	1.5
117	724	C	5	H	1	31.62	11.2	-20.2	1.5
117	724	C	5	H	1	32.02	11.3	-20.2	2.0
117	724	C	5	H	1	32.22	8.0	-20.1	2.1
117	724	C	5	H	1	32.42	5.6	-20.7	2.2
117	724	C	5	H	1	32.62	7.9	-20.3	3.5
117	724	C	5	H	2	32.82	11.9	-20.3	2.2
117	724	C	5	H	2	33.02	11.1	-20.8	1.6
117	724	C	5	H	2	33.22	10.0	-22.7	1.0
117	724	C	5	H	2	33.42	11.8	-21.0	0.8
117	724	C	5	H	2	33.62	10.2	-20.6	0.6
117	724	C	5	H	2	33.82	14.3	-19.5	1.6

117	724	C	5	H	2	34.02	8.5	-19.8	1.7
117	724	C	5	H	3	34.42	15.5	-19.8	1.3
117	724	C	5	H	3	34.62	14.1	-20.1	1.0
117	724	C	5	H	3	34.82	5.3	-19.4	1.1
117	724	C	5	H	3	35.02	10.5	-19.9	0.9
117	724	C	5	H	3	35.22	11.9	-18.5	2.3
117	724	C	5	H	3	35.42	10.0	-19.8	2.4
117	724	C	5	H	3	35.46	14.2	-19.3	2.0
117	724	C	5	H	4	35.48	5.8	-19.5	2.2
117	724	C	5	H	4	36.02	7.2	-19.6	3.0
117	724	C	5	H	4	36.22	12.5	-18.8	4.2
117	724	C	5	H	4	36.42	6.8	-20.1	4.2
117	724	C	5	H	4	36.62	7.9	-19.7	3.7
117	724	C	5	H	4	36.82	7.1	-19.4	3.0
117	724	C	5	H	4	37.02	8.5	-19.6	4.5
117	724	C	5	H	5	37.22	9.6	-18.5	2.6
117	724	C	5	H	5	37.42	11.0	-19.5	3.5
117	724	C	5	H	5	37.62	6.8	-20.2	3.2
117	724	C	5	H	5	38.02	12.6	-19.6	3.3
117	724	C	5	H	5	38.22	11.8	-19.9	3.4
117	724	C	5	H	5	38.42	10.4	-19.6	3.8
117	724	C	5	H	5	38.62	9.8	-19.9	3.0
117	724	C	5	H	6	38.82	15.3	-19.9	3.0
117	724	C	5	H	6	39.02	13.8	-20.0	3.8
117	724	C	5	H	6	39.22	12.9	-18.8	3.4

117	724	C	5	H	6	39.42	11.9	-21.9	1.9
117	724	C	5	H	6	39.62	13.2	-23.3	2.4
117	724	C	5	H	6	39.82	12.1	-20.3	2.4
117	724	C	5	H	6	40.02	11.9	-19.4	2.1
117	724	C	5	H	7	40.22	9.3	-19.9	3.0
117	724	C	5	H	7	40.42	12.3	-20.0	1.7
117	724	C	5	H	7	40.62	11.0	-19.6	1.6
117	724	C	5	H	7	40.82	11.7	-19.6	2.1
117	724	C	6	X	1	40.84	11.7	-20.8	3.2
117	724	C	6	X	1	41.02	9.7	-21.4	3.4
117	724	C	6	X	1	41.22	9.4	-20.4	5.1
117	724	C	6	X	1	41.42	9.6	-19.9	3.6
117	724	C	6	X	1	41.62	11.1	-19.8	2.8
117	724	C	6	X	1	41.82	11.1	-19.5	2.2
117	724	C	6	X	1	42.02	9.6	-19.9	2.7
117	724	C	6	X	1	42.22	9.8	-19.9	4.0
117	724	C	6	X	2	42.42	13.0	-22.6	1.8
117	724	C	6	X	2	42.62	8.7	-20.7	5.1
117	724	C	6	X	2	42.82	11.5	-20.6	4.2
117	724	C	6	X	2	43.02	10.0	-21.5	1.3
117	724	C	6	X	2	43.22	8.5	-20.3	1.2
117	724	C	6	X	2	43.42	9.9	-19.9	2.5
117	724	C	6	X	2	43.62	11.3	-20.3	2.3
117	724	C	6	X	3	43.82	11.6	-20.0	1.9
117	724	C	6	X	3	44.02	14.7	-20.1	1.9

117	724	C	6 X 3	44.22	7.4	-19.2	2.3
117	724	C	6 X 3	44.42	11.2	-20.3	1.7
117	724	C	6 X 3	44.62	8.1	-19.8	2.9
117	724	C	6 X 3	44.82	9.8	-19.4	2.6
117	724	C	6 X 3	45.02	11.5	-19.7	2.1
117	724	C	6 X 3	45.22	11.1	-18.9	2.0
117	724	C	6 X 4	45.42	12.2	-20.3	1.3
117	724	C	6 X 4	45.62	9.7	-19.6	3.5
117	724	C	6 X 4	45.82	8.2	-19.4	3.1
117	724	C	6 X 4	46.02	10.8	-19.8	5.7
117	724	C	6 X 4	46.22	10.9	-19.6	4.3
117	724	C	6 X 4	46.42	5.4	-19.7	3.1
117	724	C	6 X 4	46.62	8.4	-20.0	2.5
117	724	C	6 X 5	46.82	9.7	-20.1	2.4
117	724	C	6 X 5	47.02	7.1	-20.2	2.6
117	724	C	5 X 5	47.22	11.3	-20.7	3.4
117	724	C	6 X 5	47.42	12.5	-20.0	1.8
117	724	C	6 X 5	47.62	7.7	-20.2	2.3
117	724	C	6 X 5	47.82	12.4	-19.8	1.6
117	724	C	6 X 5	48.02	11.2	-20.6	1.7
117	724	C	6 X 5	48.22	12.0	-19.2	1.6
117	724	C	6 X 6	48.42	6.1	-20.6	2.6
117	724	C	6 X 6	48.62	10.2	-19.6	2.2
117	724	C	6 X 6	48.82	10.8	-21.2	4.0
117	724	C	6 X 6	49.02	9.3	-19.9	1.8

117	724	C	6 X 6	49.22	10.3	-19.2	1.2
117	724	C	7 X 1	50.72	12.5	-21.0	6.2
117	724	C	7 X 1	50.92	9.8	-20.9	5.2
117	724	C	7 X 1	51.12	10.0	-20.4	2.5
117	724	C	7 X 1	51.32	11.1	-20.9	5.2
117	724	C	7 X 1	51.52	10.3	-20.8	3.6
117	724	C	7 X 1	51.72	5.5	-20.2	2.2
117	724	C	7 X 1	51.92	8.0	-20.1	1.4
117	724	C	7 X 2	52.12	9.6	-20.2	1.7
117	724	C	7 X 2	52.32	10.0	-20.2	2.0
117	724	C	7 X 2	52.52	4.5	-20.5	1.6
117	724	C	7 X 2	52.72	8.1	-19.9	1.2
117	724	C	7 X 2	52.92	4.5	-19.0	1.0
117	724	C	7 X 2	53.12	16.9	-20.0	1.6
117	724	C	7 X 2	53.32	12.0	-20.6	1.1
117	724	C	7 X 3	53.52	8.1	-19.5	0.8
117	724	C	7 X 3	53.72	13.1	-20.2	1.0
117	724	C	7 X 3	53.92	10.2	-20.6	1.3
117	724	C	7 X 3	54.32	12.6	-20.2	1.8
117	724	C	7 X 3	54.52	7.8	-20.2	2.1
117	724	C	7 X 3	54.72	6.1	-20.1	3.0
117	724	C	7 X 3	54.92	9.9	-19.8	3.1
117	724	C	7 X 4	55.12	6.9	-21.2	3.8
117	724	C	7 X 4	55.32	11.9	-20.4	4.4
117	724	C	7 X 4	55.52	11.0	-20.1	5.1

117	724	C	7	X	4	55.72	11.9	-20.8	4.3
117	724	C	7	X	4	55.92	9.7	-22.9	2.5
117	724	C	7	X	4	56.12	5.7	-19.4	2.8
117	724	C	7	X	4	56.32	8.6	-21.5	4.6
117	724	C	7	X	5	56.52	15.0	-23.2	2.6
117	724	C	7	X	5	56.72	6.7	-20.4	2.9
117	724	C	7	X	5	56.92	7.5	-20.3	3.3
117	724	C	7	X	5	57.12	9.1	-20.6	3.7
117	724	C	7	X	5	57.32	11.3	-20.6	2.2
117	724	C	7	X	5	57.52	9.3	-17.7	1.9
117	724	C	7	X	5	57.72	10.1	-20.5	1.9
117	724	C	7	X	5	57.92	8.3	-19.8	2.1
117	724	C	7	X	6	58.12	9.0	-20.3	2.7
117	724	C	7	X	6	58.32	12.3	-19.3	2.8
117	724	C	7	X	6	58.52	8.2	-20.1	2.6
117	724	C	7	X	6	58.72	8.9	-20.1	3.5
117	724	C	8	X	1	60.22	18.9	-20.2	4.1
117	724	C	8	X	1	60.42	6.2	-21.1	4.1
117	724	C	8	X	1	60.62	6.7	-20.7	4.9
117	724	C	8	X	1	60.82	5.7	-20.4	3.6
117	724	C	8	X	1	61.02	5.4	-20.5	4.7
117	724	C	8	X	1	61.22	8.2	-20.3	3.3
117	724	C	8	X	1	61.42	8.7	-20.2	3.3
117	724	C	8	X	1	61.62	6.1	-20.9	4.7
117	724	C	8	X	2	61.82	8.8	-19.9	5.7

117	724	C	8	X	2	62.02	8.9	-20.1	3.5
117	724	C	8	X	2	62.22	4.7	-20.5	3.6
117	724	C	8	X	2	62.42	6.4	-21.0	5.2
117	724	C	8	X	2	62.62	16.4	-20.5	4.5
117	724	C	8	X	2	62.82	5.8	-20.7	4.7
117	724	C	8	X	2	63.02	4.9	-19.0	1.4
117	724	C	8	X	3	63.22	4.4	-19.3	1.1
117	724	C	8	X	3	63.42	6.3	-19.2	1.3
117	724	C	8	X	3	63.62	10.7	-18.9	1.1
117	724	C	8	X	3	63.82	11.0	-20.1	2.3
117	724	C	8	X	3	64.02	12.9	-19.5	1.2
117	724	C	8	X	3	64.22	11.2	-19.5	1.5
117	724	C	8	X	3	64.42	6.2	-19.6	1.8
117	724	C	8	X	3	64.62	11.2	-20.3	4.3
117	724	C	8	X	4	64.82	4.3	-19.7	4.0
117	724	C	8	X	4	65.02	4.7	-20.4	2.8
117	724	C	8	X	4	65.22	6.8	-20.3	3.8
117	724	C	8	X	4	65.42	8.3	-19.4	3.1
117	724	C	8	X	4	65.62	7.1	-20.7	5.1
117	724	C	8	X	4	65.82	7.0	-21.2	5.9
117	724	C	8	X	4	66.02	5.7	-21.7	6.5
117	724	C	8	X	5	66.22	7.0	-23.5	12.9
117	724	C	8	X	5	66.42	9.0	-21.2	6.5
117	724	C	8	X	5	66.62	11.4	-20.5	6.0
117	724	C	8	X	5	66.82	10.4	-20.9	6.0

117	724 C	8 X 5	67.02	12.6	-20.0	2.5
117	724 C	8 X 5	67.22	12.6	-21.0	1.2
117	724 C	8 X 5	67.42	6.1	-20.1	1.7
117	724 C	8 X 5	67.62	14.0	-18.7	1.6
117	724 C	8 X 6	67.82	11.6	-22.0	2.9
117	724 C	8 X 6	68.02	4.6	-20.2	3.8
117	724 C	8 X 6	68.22	7.5	-19.9	3.9
117	724 C	9 X 1	70.02	12.5	-20.1	4.0
117	724 C	9 X 1	70.22	12.4	-19.8	4.4
117	724 C	9 X 1	70.42	8.2	-19.9	3.0
117	724 C	9 X 1	70.62	2.8	-21.3	1.3
117	724 C	9 X 1	70.82	6.5	-20.5	3.6
117	724 C	9 X 1	71.02	9.5	-19.9	5.0
117	724 C	9 X 1	71.22	8.3	-20.4	4.4
117	724 C	9 X 2	71.42	7.7	-20.8	3.9
117	724 C	9 X 2	71.62	8.9	-20.2	2.0
117	724 C	9 X 2	71.82	6.0	-21.4	3.3
117	724 C	9 X 2	72.02	13.0	-19.8	4.4
117	724 C	9 X 2	72.22	12.0	-19.9	3.3
117	724 C	9 X 2	72.42	5.4	-21.0	3.4
117	724 C	9 X 2	72.62	8.5	-21.6	2.4
117	724 C	9 X 3	72.82	12.6	-20.4	4.3
117	724 C	9 X 3	73.02	7.1	-19.9	3.8
117	724 C	9 X 3	73.42	4.1	-20.4	3.4
117	724 C	9 X 3	73.62	7.6	-20.2	4.0

117	724	C	9	X	3	73.82	6.9	-21.9	5.2
117	724	C	9	X	3	74.02	10.4	-22.0	6.4
117	724	C	9	X	3	74.22	11.3	-21.2	5.4
117	724	C	9	X	3	74.42	4.7	-21.8	8.8
117	724	C	9	X	4	74.62	12.2	-21.2	6.8
117	724	C	9	X	4	74.82	7.1	-23.0	3.6
117	724	C	9	X	4	75.02	9.9	-21.5	4.3
117	724	C	9	X	4	75.22	9.8	-20.8	4.8
117	724	C	9	X	4	75.42	11.9	-20.7	5.0
117	724	C	9	X	5	75.82	10.8	-19.6	3.1
117	724	C	9	X	5	76.02	2.8	-20.2	2.4
117	724	C	9	X	5	76.22	10.6	-20.5	3.2
117	724	C	9	X	5	76.42	16.4	-20.5	3.8
117	724	C	9	X	5	76.82	10.7	-20.5	4.3
117	724	C	9	X	5	77.02	9.2	-20.9	3.0
117	724	C	9	X	5	77.22	4.6	-22.3	2.5
117	724	C	9	X	6	77.42	7.1	-20.9	3.3
117	724	C	9	X	6	77.82	11.5	-20.5	3.7
117	724	C	9	X	6	78.02	5.9	-20.8	3.0
117	724	C	9	X	6	78.42	9.3	-20.2	3.2
117	724	C	9	X	6	78.62	7.8	-21.4	3.0
117	724	C	9	X	7	78.82	12.8	-19.4	1.0
117	724	C	10	X	1	79.42	10.0	-21.0	7.7
117	724	C	10	X	1	79.62	8.3	-21.3	7.4
117	724	C	10	X	1	79.82	8.8	-21.9	6.3

117	724	C	10	X	1	80.02	15.2	-21.3	6.5
117	724	C	10	X	1	80.22	9.3	-21.5	5.1
117	724	C	10	X	1	80.42	10.5	-21.5	6.7
117	724	C	10	X	1	80.62	10.7	-20.9	7.4
117	724	C	10	X	1	80.82	8.2	-20.3	5.7
117	724	C	10	X	2	81.02	9.2	-21.3	8.0
117	724	C	10	X	2	81.42	8.8	-20.8	5.3
117	724	C	10	X	2	81.62	10.6	-20.7	4.9
117	724	C	10	X	2	82.02	9.7	-20.8	4.4
117	724	C	10	X	2	82.22	9.8	-21.1	7.3
117	724	C	10	X	3	82.42	9.0	-20.3	5.6
117	724	C	10	X	3	82.62	13.9	-20.4	5.1
117	724	C	10	X	3	82.82	10.6	-21.1	2.7
117	724	C	10	X	3	83.02	11.2	-21.0	3.9
117	724	C	10	X	3	83.22	11.2	-20.8	3.6
117	724	C	10	X	3	83.82	12.0	-21.2	2.4
117	724	C	10	X	4	84.20	9.8	-20.0	3.8
117	724	C	10	X	4	84.42	10.7	-20.6	2.8
117	724	C	10	X	4	84.62	13.6	-20.2	3.5
117	724	C	10	X	4	85.02	5.9	-20.2	0.3
117	724	C	10	X	4	85.22	11.1	-22.1	1.6
117	724	C	10	X	5	85.42	9.7	-20.2	2.6
117	724	C	10	X	5	85.62	9.1	-20.8	3.2
117	724	C	10	X	5	85.82	8.8	-21.6	5.0
117	724	C	10	X	5	86.02	8.6	-22.4	4.4

117	724 C	10 X 5	86.22	8.5	-21.3	4.0
117	724 C	10 X 5	86.42	7.8	-21.4	4.6
117	724 C	10 X 5	86.62	6.3	-20.5	3.5
117	724 C	10 X CC	86.77	5.4	-20.7	1.7
117	724 C	10 X CC	86.97	6.6	-20.8	2.5
117	724 C	11 X 1	89.02	9.1	-21.0	4.2
117	724 C	11 X 1	89.22	6.7	-22.1	4.3
117	724 C	11 X 1	89.42	9.0	-20.6	3.9
117	724 C	11 X 1	89.62	4.2	-20.3	2.6
117	724 C	11 X 1	89.82	8.2	-19.8	3.4
117	724 C	11 X 1	90.02	9.4	-20.5	6.3
117	724 C	11 X 2	90.62	7.7	-19.9	2.6
117	724 C	11 X 2	90.82	8.9	-20.1	3.7
117	724 C	11 X 2	91.02	10.2	-19.8	3.8
117	724 C	11 X 2	91.22	11.7	-20.0	5.1
117	724 C	11 X 2	91.42	9.0	-19.7	4.6
117	724 C	11 X 2	91.62	6.7	-20.2	4.1
117	724 C	11 X 2	91.82	10.0	-20.2	4.4
117	724 C	11 X 3	92.02	5.8	-19.1	5.3
117	724 C	11 X 3	92.22	7.5	-19.8	5.0
117	724 C	11 X 3	92.42	9.2	-19.8	3.9
117	724 C	11 X 3	92.82	9.5	-19.7	8.4
117	724 C	11 X 3	93.02	11.0	-20.2	7.5
117	724 C	11 X 3	93.22	7.0	-20.2	6.3
117	724 C	11 X 3	93.42	7.6	-19.8	5.7

117	724	C	11	X	4	93.82	7.5	-19.9	5.9
117	724	C	11	X	4	94.02	7.0	-20.3	5.1
117	724	C	11	X	4	94.22	7.2	-19.4	4.4
117	724	C	11	X	4	94.42	7.2	-20.2	3.8
117	724	C	11	X	4	94.62	7.6	-19.8	4.0
117	724	C	11	X	5	95.02	6.0	-19.8	4.4
117	724	C	11	X	5	95.22	9.2	-20.7	5.0
117	724	C	11	X	5	95.42	9.0	-19.6	6.3
117	724	C	11	X	5	95.82	9.9	-20.2	8.4
117	724	C	11	X	5	96.02	12.2	-19.9	5.2
117	724	C	11	X	5	96.22	6.1	-21.3	5.8
117	724	C	11	X	6	96.62	9.6	-20.3	4.5
117	724	C	11	X	6	96.82	7.5	-20.1	5.6
117	724	C	11	X	6	97.42	8.1	-20.6	5.7
117	724	C	11	X	6	97.60	8.9	-21.8	7.8
117	724	C	11	X	6	97.82	8.9	-20.5	9.5
117	724	C	11	X	7	98.02	7.3	-21.6	7.5
117	724	C	11	X	7	98.22	10.9	-21.4	6.9
117	724	C	12	X	1	98.62	9.1	-20.4	9.8
117	724	C	12	X	1	98.82	7.9	-20.2	5.1
117	724	C	12	X	1	99.02	6.3	-19.5	4.5
117	724	C	12	X	1	99.42	7.6	-19.7	3.9
117	724	C	12	X	1	99.60	7.6	-20.1	6.1
117	724	C	12	X	1	99.80	5.6	-19.8	4.7
117	724	C	12	X	1	100.22	9.2	-19.7	5.9

117	724	C	12	X	1	100.44	11.9	-19.7	6.9
117	724	C	12	X	2	100.62	8.1	-20.8	6.5
117	724	C	12	X	2	100.82	9.3	-20.6	4.6
117	724	C	12	X	2	101.04	10.4	-20.0	4.9
117	724	C	12	X	2	101.22	8.7	-24.4	9.5
117	724	C	12	X	2	101.42	7.0	-20.7	3.3
117	724	C	12	X	3	101.62	10.4	-19.9	2.1
117	724	C	12	X	3	101.82	6.9	-19.9	2.4
117	724	C	12	X	3	102.42	7.1	-19.7	2.3
117	724	C	12	X	3	102.62	7.8	-19.8	3.9
117	724	C	12	X	3	102.82	4.8	-19.8	3.9
117	724	C	12	X	3	103.02	7.3	-19.9	7.7
117	724	C	12	X	4	103.42	11.2	-19.7	6.0
117	724	C	12	X	4	103.62	8.9	-21.1	3.8
117	724	C	12	X	4	103.82	8.7	-20.1	4.5
117	724	C	12	X	4	104.02	8.8	-20.6	4.8
117	724	C	12	X	4	104.22	7.9	-20.0	7.7
117	724	C	12	X	4	104.42	6.4	-20.1	8.9
117	724	C	12	X	5	104.62	6.8	-20.0	9.6
117	724	C	12	X	5	105.02	10.5	-20.1	5.0
117	724	C	12	X	5	105.22	7.1	-20.5	7.2
117	724	C	12	X	5	105.42	9.2	-20.2	4.5
117	724	C	12	X	5	105.62	9.2	-20.3	5.8
117	724	C	12	X	5	105.82	7.7	-20.5	3.3
117	724	C	12	X	6	106.23	4.5	-20.3	3.8

117	724	C	12	X	6	106.42	4.8	-20.2	4.8
117	724	C	12	X	6	106.62	5.9	-20.4	5.2
117	724	C	12	X	6	106.82	8.7	-20.6	5.4
117	724	C	12	X	6	107.02	8.9	-20.7	6.3
117	724	C	12	X	6	107.22	7.1	-19.9	6.5
117	724	C	12	X	6	107.42	7.6	-20.0	8.2
117	724	C	12	X	7	107.62	7.2	-20.4	7.7
117	724	C	12	X	7	107.82	6.9	-20.1	6.5
117	724	C	13	X	1	108.32	8.7	-20.3	7.9
117	724	C	13	X	1	108.52	6.6	-22.2	8.1
117	724	C	13	X	1	108.72	8.6	-20.0	11.4
117	724	C	13	X	1	108.92	7.3	-22.0	7.3
117	724	C	13	X	1	109.10	7.9	-22.1	6.1
117	724	C	13	X	1	109.52	8.6	-20.9	7.1
117	724	C	13	X	2	109.92	7.6	-21.9	5.7
117	724	C	13	X	2	110.12	10.3	-20.3	8.4
117	724	C	13	X	2	110.32	7.6	-19.8	7.2
117	724	C	13	X	2	110.52	9.2	-20.4	9.0
117	724	C	13	X	2	110.72	7.9	-21.2	6.6
117	724	C	13	X	2	110.94	7.7	-20.5	8.4
117	724	C	13	X	2	111.12	5.5	-20.0	7.1
117	724	C	13	X	3	111.32	5.5	-20.3	6.5
117	724	C	13	X	3	111.72	10.0	-20.1	7.5
117	724	C	13	X	3	111.92	9.0	-20.3	5.7
117	724	C	13	X	3	112.12	7.5	-20.5	8.8

117	724	C	13	X	3	112.32	8.5	-21.9	6.3
117	724	C	13	X	3	112.52	8.9	-22.1	8.5
117	724	C	13	X	4	112.92	8.5	-20.3	11.1
117	724	C	13	X	4	113.12	9.1	-20.8	8.4
117	724	C	13	X	4	113.32	7.8	-20.5	8.2
117	724	C	13	X	4	113.52	11.0	-20.2	8.9
117	724	C	13	X	4	113.72	10.1	-20.6	12.7
117	724	C	13	X	4	113.92	9.7	-20.3	7.0
117	724	C	13	X	5	114.32	8.7	-20.1	6.1
117	724	C	13	X	5	114.52	9.6	-19.9	7.5
117	724	C	13	X	5	114.72	7.3	-22.5	4.2
117	724	C	13	X	5	115.14	7.3	-20.0	6.6
117	724	C	13	X	5	115.32	5.4	-21.9	6.3
117	724	C	13	X	5	115.52	5.4	-20.6	5.5
117	724	C	13	X	6	115.92	6.5	-20.1	6.0
117	724	C	13	X	6	116.10	7.7	-20.2	8.0
117	724	C	13	X	6	116.32	5.8	-20.5	7.3
117	724	C	13	X	6	116.52	5.7	-20.3	3.4
117	724	C	13	X	6	116.72	6.6	-20.7	5.5
117	724	C	13	X	6	116.92	5.8	-22.5	3.9
117	724	C	13	X	7	117.32	7.6	-20.5	7.2
117	724	C	13	X	7	117.52	7.6	-20.0	5.1
117	724	C	13	X	7	117.72	7.4	-22.2	4.7
117	724	C	14	X	1	117.82	10.8	-20.2	7.5
117	724	C	14	X	1	118.40	8.6	-20.3	7.1

117	724 C	14 X 1	118.67	1.0	-20.2	6.4
117	724 C	14 X 1	118.81	9.0	-20.9	6.6
117	724 C	14 X 1	119.02	9.6	-20.1	5.3
117	724 C	14 X 2	119.42	10.2	-20.1	7.1
117	724 C	14 X 2	119.60	8.4	-20.4	5.7
117	724 C	14 X 2	119.85	9.7	-20.0	6.1
117	724 C	14 X 2	120.02	8.8	-21.3	4.8
117	724 C	14 X 2	120.38	7.0	-20.4	7.2
117	724 C	14 X 3	120.87	6.4	-21.6	5.1
117	724 C	14 X 3	121.63	7.1	-20.3	6.8
117	724 C	14 X 3	121.81	10.2	-20.4	5.7
117	724 C	14 X 3	122.02	6.8	-20.5	6.5
117	724 C	14 X 4	122.42	8.5	-20.9	5.7
117	724 C	14 X 4	122.62	7.4	-20.7	5.8
117	724 C	14 X 4	122.82	7.9	-23.0	3.8
117	724 C	14 X 4	123.02	6.6	-22.4	4.4
117	724 C	14 X 4	123.42	7.5	-22.8	3.3
117	724 C	14 X 4	123.62	9.3	-20.0	5.2
117	724 C	14 X 5	123.82	7.5	-20.5	5.2
117	724 C	14 X 5	124.23	6.8	-20.4	3.7
117	724 C	14 X 5	124.42	9.7	-20.5	4.1
117	724 C	14 X 5	124.62	8.0	-20.2	5.8
117	724 C	14 X 5	125.02	7.6	-22.0	5.9
117	724 C	14 X 5	125.22	7.8	-20.0	4.1
117	724 C	14 X 6	125.42	9.5	-20.0	3.0

117	724	C	14	X	6	125.62	4.6	-20.4	2.6
117	724	C	14	X	6	125.82	8.2	-23.0	1.6
117	724	C	15	X	1	127.42	6.3	-21.0	4.5
117	724	C	15	X	1	127.62	7.8	-21.9	3.6
117	724	C	15	X	1	128.02	7.4	-22.1	4.7
117	724	C	15	X	1	128.20	6.2	-20.9	5.4
117	724	C	15	X	1	128.60	5.7	-21.1	4.0
117	724	C	15	X	2	129.02	5.4	-21.2	3.3
117	724	C	15	X	2	129.22	5.8	-20.7	4.3
117	724	C	15	X	2	129.65	5.7	-20.3	4.4
117	724	C	15	X	2	129.83	6.0	-20.1	4.9
117	724	C	15	X	3	130.42	5.4	-21.0	4.0
117	724	C	15	X	3	130.60	6.9	-20.2	5.3
117	724	C	15	X	3	130.84	7.1	-21.4	4.0
117	724	C	15	X	3	131.00	6.1	-20.4	4.6
117	724	C	15	X	3	131.20	6.1	-20.8	4.2
117	724	C	15	X	3	131.43	6.1	-21.5	3.4
117	724	C	15	X	3	131.52	5.3	-20.7	3.6
117	724	C	15	X	3	131.60	6.1	-21.0	4.3
117	724	C	15	X	4	132.00	6.8	-20.6	4.0
117	724	C	15	X	4	132.45	6.2	-20.7	6.2
117	724	C	15	X	4	132.61	7.5	-20.3	3.1
117	724	C	15	X	5	133.42	4.1	-20.4	1.8
117	724	C	15	X	5	133.80	3.7	-19.9	3.0
117	724	C	15	X	5	134.00	4.2	-20.7	3.6

117	724	C	15 X 5	134.20	6.5	-21.5	7.8
117	724	C	15 X 5	134.40	4.8	-20.2	5.6
117	724	C	15 X 5	134.65	4.8	-21.5	3.4
117	724	C	15 X 5	134.74	7.4	-21.6	6.0
117	724	C	15 X 5	134.82	8.3	-21.6	10.2
117	724	C	15 X 6	135.00	5.6	-20.1	4.9
117	724	C	15 X 6	135.22	6.3	-21.7	4.2
117	724	C	15 X 6	135.40	7.2	-21.4	5.5
117	724	C	15 X 6	135.72	7.0	-21.1	5.1
117	724	C	15 X 6	135.94	6.7	-20.7	5.5
117	724	C	15 X 7	136.42	7.7	-22.8	4.5
117	724	C	15 X 7	136.62	7.9	-21.9	4.8
117	724	C	15 X 7	136.75	5.6	-23.0	5.0
117	724	C	16 X 1	137.02	5.0	-21.7	12.4
117	724	C	16 X 1	137.22	3.4	-22.3	9.8
117	724	C	16 X 1	137.42	7.3	-21.7	9.6
117	724	C	16 X 1	137.62	7.9	-19.7	10.2
117	724	C	16 X 1	137.82	7.2	-22.6	8.1
117	724	C	16 X 1	138.02	6.1	-21.2	8.1
117	724	C	16 X 1	138.42	6.2	-22.7	7.4
117	724	C	16 X 2	138.62	6.0	-22.1	9.5
117	724	C	16 X 2	138.82	7.6	-21.7	8.3
117	724	C	16 X 2	139.02	7.5	-21.0	9.1
117	724	C	16 X 2	139.22	5.4	-20.9	6.7
117	724	C	16 X 2	139.42	6.0	-21.3	5.7

117	724	C	16	X	2	139.60	5.5	-21.6	5.4
117	724	C	16	X	2	139.82	5.8	-21.3	6.4
117	724	C	16	X	3	140.02	6.0	-23.0	5.2
117	724	C	16	X	3	140.22	5.9	-20.9	6.9
117	724	C	16	X	3	140.62	4.0	-20.7	7.5
117	724	C	16	X	3	140.82	4.5	-20.2	5.9
117	724	C	16	X	3	141.02	4.9	-20.8	6.3
117	724	C	16	X	3	141.22	6.6	-20.0	6.3
117	724	C	16	X	4	141.60	7.1	-21.2	7.0
117	724	C	16	X	4	141.82	7.1	-22.8	4.9
117	724	C	16	X	4	142.02	5.5	-22.0	5.1
117	724	C	16	X	4	142.22	4.7	-20.6	9.5
117	724	C	16	X	4	142.40	5.9	-20.4	6.8
117	724	C	16	X	4	142.60	5.5	-20.6	7.9
117	724	C	16	X	4	142.82	5.4	-20.5	8.7
117	724	C	16	X	5	143.07	6.6	-20.7	5.5
117	724	C	16	X	5	143.23	5.5	-21.3	5.2
117	724	C	16	X	5	143.40	7.3	-21.1	5.4
117	724	C	16	X	5	143.62	5.3	-20.4	7.9
117	724	C	16	X	5	144.02	6.7	-20.4	7.9
117	724	C	16	X	5	144.22	5.2	-20.6	7.4
117	724	C	16	X	5	144.42	5.8	-20.3	8.0
117	724	C	16	X	6	144.62	4.1	-20.8	4.1
117	724	C	16	X	6	144.82	5.8	-22.7	4.9
117	724	C	16	X	6	144.98	3.5	-20.3	6.4

117	724	C	16	X	6	145.22	5.2	-20.9	6.9
117	724	C	16	X	6	145.42	5.1	-20.7	6.9
117	724	C	16	X	6	145.62	5.3	-20.6	6.7
117	724	C	16	X	6	145.82	5.6	-21.0	6.4
117	724	C	16	X	7	146.02	3.9	-20.9	3.9
117	724	C	16	X	7	146.22	6.5	-21.0	6.5
117	724	C	16	X	7	146.42	4.5	-21.0	5.9
117	724	C	17	X	1	146.62	8.2	-19.8	7.0
117	724	C	17	X	1	146.82	6.2	-19.7	6.4
117	724	C	17	X	1	147.02	7.9	-20.5	7.1
117	724	C	17	X	1	147.22	7.9	-19.7	7.2
117	724	C	17	X	1	148.02	6.5	-19.9	7.5
117	724	C	17	X	2	148.20	6.3	-19.7	7.5
117	724	C	17	X	2	148.42	8.2	-19.7	7.9
117	724	C	17	X	2	148.62	10.3	-20.1	9.3
117	724	C	17	X	2	148.81	7.9	-22.5	4.3
117	724	C	17	X	3	149.62	5.6	-21.0	8.6
117	724	C	17	X	3	149.82	8.5	-20.1	9.1
117	724	C	17	X	3	150.02	6.7	-22.4	8.2
117	724	C	17	X	3	150.22	6.8	-20.0	11.9
117	724	C	17	X	3	150.42	8.1	-20.0	10.8
117	724	C	17	X	3	150.62	6.2	-20.2	10.3
117	724	C	17	X	3	150.82	4.4	-21.2	9.9
117	724	C	17	X	3	151.02	6.6	-19.7	8.2
117	724	C	17	X	4	151.20	6.6	-21.7	6.4

117	724	C	17	X	4	151.42	6.8	-19.9	8.2
117	724	C	17	X	4	151.48	7.4	-19.9	8.8
117	724	C	17	X	4	152.22	9.1	-19.9	6.0
117	724	C	17	X	4	152.42	7.2	-21.3	6.9
117	724	C	17	X	5	152.62	6.9	-19.9	9.3
117	724	C	17	X	5	152.82	4.4	-20.2	10.9
117	724	C	17	X	5	153.02	8.3	-20.2	6.3
117	724	C	17	X	5	153.22	7.1	-21.8	5.0
117	724	C	17	X	5	153.42	9.1	-20.2	5.6
117	724	C	17	X	5	153.62	6.1	-21.0	5.9
117	724	C	17	X	5	153.82	8.9	-20.2	5.6
117	724	C	17	X	6	154.22	9.9	-20.2	6.1
117	724	C	17	X	6	154.42	14.0	-20.0	6.0
117	724	C	17	X	6	154.62	10.5	-20.6	3.3
117	724	C	17	X	6	154.82	7.5	-20.4	7.3
117	724	C	17	X	6	155.02	8.3	-20.2	4.1
117	724	C	17	X	6	155.42	14.3	-22.8	1.4
117	724	C	18	X	1	156.32	7.5	-20.2	6.8
117	724	C	18	X	1	156.52	7.3	-20.8	7.7
117	724	C	18	X	1	156.72	7.3	-20.4	7.3
117	724	C	18	X	1	156.92	6.8	-20.7	6.3
117	724	C	18	X	1	157.12	5.9	-20.3	6.6
117	724	C	18	X	1	157.32	1.5	-20.3	6.1
117	724	C	18	X	1	157.52	7.6	-20.8	6.2
117	724	C	18	X	1	157.68	7.5	-20.6	7.8

117	724 C	18 X 2	157.92	6.5	-20.8	9.4
117	724 C	18 X 2	158.10	7.2	-20.6	6.5
117	724 C	18 X 2	158.72	7.0	-20.1	5.6
117	724 C	18 X 2	158.92	6.5	-20.1	6.1
117	724 C	18 X 2	159.12	6.0	-20.9	6.4
117	724 C	18 X 3	159.32	4.9	-21.0	7.5
117	724 C	18 X 3	159.52	7.0	-20.5	5.9
117	724 C	18 X 3	159.72	8.0	-20.9	10.6
117	724 C	18 X 3	159.92	9.2	-22.1	9.2
117	724 C	18 X 3	160.12	7.4	-20.3	10.0
117	724 C	18 X 3	160.32	6.1	-20.3	7.3
117	724 C	18 X 3	160.52	6.9	-19.9	4.7
117	724 C	18 X 4	160.95	7.5	-20.1	5.0
117	724 C	18 X 4	161.12	5.7	-21.5	6.7
117	724 C	18 X 4	161.32	7.5	-21.0	6.6
117	724 C	18 X 4	161.72	9.5	-19.9	4.0
117	724 C	18 X 4	161.92	5.7	-20.5	6.2
117	724 C	18 X 4	162.12	5.7	-19.9	11.4
117	724 C	18 X 5	162.32	7.2	-20.0	10.6
117	724 C	18 X 5	162.52	6.1	-20.5	6.4
117	724 C	18 X 5	163.12	6.5	-21.4	6.1
117	724 C	18 X 5	163.32	5.7	-19.9	2.1
117	724 C	18 X 5	163.72	6.5	-20.4	6.2
117	724 C	18 X 6	163.92	8.3	-20.2	4.0
117	724 C	18 X 6	164.12	5.8	-20.0	4.0

117	724	C	18	X	6	164.52	5.9	-20.2	3.6
117	724	C	18	X	6	164.72	5.3	-20.1	6.3
117	724	C	18	X	6	164.92	6.2	-19.7	4.3
117	724	C	18	X	6	165.12	5.2	-20.0	4.0
117	724	C	19	X	1	166.09	6.8	-20.2	9.1
117	724	C	19	X	1	166.31	6.7	-20.2	8.9
117	724	C	19	X	1	166.50	5.5	-20.3	6.4
117	724	C	19	X	1	167.14	7.2	-20.2	5.9
117	724	C	19	X	1	167.32	6.7	-19.9	8.6
117	724	C	19	X	2	167.52	6.3	-21.2	6.7
117	724	C	19	X	2	167.72	5.7	-19.9	7.7
117	724	C	19	X	2	167.92	5.2	-20.6	5.9
117	724	C	19	X	2	168.12	5.8	-20.3	4.1
117	724	C	19	X	2	168.25	7.3	-20.5	7.9
117	724	C	19	X	2	168.52	6.7	-21.2	1.0
117	724	C	19	X	2	168.72	7.4	-19.6	8.2
117	724	C	20	X	1	175.54	8.4	-19.7	5.4
117	724	C	20	X	1	175.72	5.5	-19.7	8.8
117	724	C	20	X	1	175.92	8.2	-19.3	6.6
117	724	C	20	X	1	176.36	8.9	-19.7	5.5
117	724	C	20	X	1	176.52	5.2	-20.3	6.9
117	724	C	20	X	1	176.74	8.3	-19.7	6.1
117	724	C	20	X	1	176.92	6.1	-20.0	5.1
117	724	C	20	X	2	177.12	8.6	-20.0	11.0
117	724	C	20	X	2	177.32	8.8	-19.7	8.1

117	724 C	20 X 2	177.72	5.4	-19.7	7.3
117	724 C	20 X 2	178.12	9.4	-20.0	5.7
117	724 C	20 X 2	178.36	6.5	-21.2	4.5
117	724 C	20 X 3	178.52	4.7	-20.3	3.8
117	724 C	20 X 3	178.72	7.5	-19.9	3.9
117	724 C	20 X 3	178.92	7.5	-19.8	4.4
117	724 C	20 X 3	179.29	7.7	-20.7	6.2
117	724 C	20 X 3	179.52	6.3	-19.8	7.9
117	724 C	20 X 3	179.72	5.6	-21.7	6.7
117	724 C	20 X 3	179.92	6.9	-20.2	11.6
117	724 C	20 X 4	180.12	11.3	-20.4	10.7
117	724 C	20 X 4	180.32	7.4	-20.8	5.7
117	724 C	20 X 4	180.52	8.7	-20.4	7.6
117	724 C	20 X 4	180.72	6.2	-20.5	2.8
117	724 C	20 X 4	181.12	3.7	-20.5	3.0
117	724 C	20 X 4	181.32	5.6	-19.5	3.3
117	724 C	20 X 5	181.52	3.6	-19.4	4.4
117	724 C	20 X 5	181.72	9.4	-19.4	4.5
117	724 C	20 X 5	181.92	7.6	-19.4	5.0
117	724 C	20 X 5	182.12	5.3	-19.5	5.5
117	724 C	20 X 5	182.32	6.4	-19.3	5.8
117	724 C	20 X 5	182.52	6.4	-19.6	7.7
117	724 C	20 X 5	182.72	4.4	-19.3	5.5
117	724 C	20 X 5	182.92	7.4	-19.5	5.9
117	724 C	20 X 6	183.12	6.1	-19.9	9.9

117	724	C	20	X	6	183.32	4.9	-20.4	11.1
117	724	C	20	X	6	183.52	7.7	-20.6	11.4
117	724	C	20	X	6	183.72	5.5	-20.5	18.2
117	724	C	20	X	6	183.92	5.6	-22.5	8.2
117	724	C	21	X	1	185.22	7.2	-20.9	10.9
117	724	C	21	X	1	185.62	5.4	-20.5	10.5
117	724	C	21	X	1	185.85	5.8	-22.0	6.3
117	724	C	21	X	1	186.02	7.8	-19.8	11.1
117	724	C	21	X	1	186.22	10.7	-20.4	8.7
117	724	C	21	X	1	186.42	9.3	-20.5	7.5
117	724	C	21	X	1	186.62	10.5	-20.5	8.6
117	724	C	21	X	2	186.82	13.3	-21.4	17.2
117	724	C	21	X	2	187.02	6.2	-20.0	7.2
117	724	C	21	X	2	187.22	5.0	-21.6	5.5
117	724	C	21	X	2	187.42	7.3	-21.2	6.2
117	724	C	21	X	2	188.02	7.6	-21.4	8.9
117	724	C	21	X	3	188.22	8.3	-19.9	14.5
117	724	C	21	X	3	189.02	9.4	-19.8	14.0
117	724	C	21	X	3	189.22	8.0	-20.5	9.6
117	724	C	21	X	3	189.42	4.8	-21.3	9.8
117	724	C	21	X	3	189.62	6.3	-22.6	4.8
117	724	C	21	X	4	189.82	6.6	-22.5	4.9
117	724	C	21	X	4	190.02	5.9	-20.8	9.2
117	724	C	21	X	4	190.22	7.1	-20.0	9.8
117	724	C	21	X	4	190.42	7.2	-20.2	11.9

117	724	C	22	X	1	194.82	3.9	-20.3	9.7
117	724	C	22	X	1	195.02	7.6	-21.0	12.1
117	724	C	22	X	1	195.42	6.5	-20.5	13.0
117	724	C	22	X	1	195.62	6.8	-19.7	14.0
117	724	C	22	X	1	195.82	6.8	-19.7	9.6
117	724	C	22	X	1	196.02	8.4	-20.2	9.6
117	724	C	22	X	1	196.22	6.4	-19.8	6.7
117	724	C	22	X	2	196.40	7.1	-19.9	5.1
117	724	C	22	X	2	196.62	6.2	-20.7	8.1
117	724	C	22	X	2	196.82	6.1	-20.1	10.2
117	724	C	22	X	2	197.02	5.9	-20.6	12.6
117	724	C	22	X	2	197.22	7.0	-19.8	10.9
117	724	C	22	X	2	197.42	7.5	-20.4	11.5
117	724	C	22	X	2	197.62	4.8	-22.1	9.9
117	724	C	22	X	3	197.82	7.5	-21.0	9.7
117	724	C	22	X	3	198.02	3.9	-21.0	8.9
117	724	C	22	X	3	198.22	8.9	-19.3	13.3
117	724	C	22	X	3	198.42	5.6	-19.2	4.6
117	724	C	22	X	3	198.62	7.6	-19.9	12.8
117	724	C	22	X	3	198.82	7.8	-20.9	10.8
117	724	C	22	X	3	199.02	6.5	-20.6	12.6
117	724	C	22	X	3	199.22	7.6	-21.1	10.9
117	724	C	22	X	4	199.42	6.9	-18.9	9.8
117	724	C	22	X	4	199.65	7.4	-19.9	14.1
117	724	C	22	X	4	199.86	7.8	-19.9	11.3

117	724 C	22 X 4	200.42	7.6	-20.5	11.1
117	724 C	22 X 4	200.62	7.0	-18.6	12.8
117	724 C	22 X 5	200.82	6.2	-20.1	11.5
117	724 C	22 X 5	201.82	7.4	-19.9	10.5
117	724 C	22 X 5	201.83	7.2	-20.4	12.4
117	724 C	22 X 5	202.28	6.4	-20.8	13.0
117	724 C	22 X 6	203.22	6.9	-21.1	15.9
117	724 C	22 X 6	203.62	6.7	-20.3	7.8
117	724 C	22 X 7	203.82	6.4	-20.2	8.7
117	724 C	22 X 7	203.97	6.9	-20.3	9.8

Table 2: Stable Carbon and Nitrogen Isotope Compositions and Residue Organic Carbon Content Results for Site 725.

Leg	Site	Core	Depth	$\delta^{15}N$	$\delta^{13}C$	% OC
117	725 C	1 H1	0.20	6.9	-20.7	1.32
117	725 C	1 H1	0.40	9.8	-19.4	0.98
117	725 C	1 H1	0.60	8.2	-22.8	1.04
117	725 C	1 H1	0.80	11.8	-19.5	0.72
117	725 C	1 H1	1.20	10.9	-19.5	0.54
117	725 C	1 H2	1.40	11.3	-19.8	0.57
117	725 C	1 H2	1.60	7.7	-20.1	1.01
117	725 C	1 H2	1.80	12.2	-19.4	0.62
117	725 C	1 H2	2.00	14.0	-19.2	0.46
117	725 C	1 H2	2.20	15.7	-19.1	0.77
117	725 C	1 H2	2.37	11.3	-18.9	0.84
117	725 C	1 H2	2.60	9.9	-18.8	0.92
117	725 C	1 H3	3.00	11.6	-18.8	1.36
117	725 C	1 H3	3.20	5.9	-18.7	0.98
117	725 C	1 H3	3.40	13.7	-19.0	0.89
117	725 C	1 H3	3.60	8.9	-18.6	0.71
117	725 C	1 H3	3.80	7.2	-20.7	0.96
117	725 C	1 H3	4.00	10.7	-18.7	1.22
117	725 C	1 H3	4.20	15.1	-18.5	1.33
117	725 C	1 H4	4.60	16.0	-19.0	1.24

117	725	C	1	H4	4.80	12.3	-19.1	1.18
117	725	C	1	H4	5.20	8.5	-19.0	0.94
117	725	C	1	H4	5.40	6.8	-19.1	1.29
117	725	C	1	H4	5.60	5.5	-19.0	1.49
117	725	C	1	H4	5.80	14.5	-19.3	1.00
117	725	C	1	H5	6.20	5.5	-19.5	0.94
117	725	C	1	H5	6.40	9.6	-18.7	0.78
117	725	C	1	H5	7.00	16.6	-20.1	1.28
117	725	C	1	H5	7.20	14.8	-19.5	0.89
117	725	C	1	H5	7.40	13.1	-19.2	0.61
117	725	C	1	H6	7.80	6.4	-19.4	0.82
117	725	C	1	H6	8.00	4.6	-20.7	0.64
117	725	C	2	H1	9.00	11.3	-20.3	0.54
117	725	C	2	H1	9.60	11.3	-19.6	0.97
117	725	C	2	H1	9.80	12.1	-19.8	1.11
117	725	C	2	H1	10.00	11.1	-19.9	1.24
117	725	C	2	H1	10.20	7.8	-20.3	1.14
117	725	C	2	H2	10.60	12.7	-19.9	2.04
117	725	C	2	H2	10.80	8.8	-19.9	1.80
117	725	C	2	H2	11.20	9.4	-19.6	0.78
117	725	C	2	H2	11.40	5.6	-20.8	1.18
117	725	C	2	H2	11.60	15.2	-19.9	0.92
117	725	C	2	H2	11.80	11.0	-20.0	0.78
117	725	C	2	H3	12.00	14.5	-20.3	0.62
117	725	C	2	H3	12.20	14.4	-19.0	0.83

117	725	C	2	H3	12.40	9.4	-20.4	0.89
117	725	C	2	H3	12.60	4.9	-20.1	0.78
117	725	C	2	H3	12.80	13.5	-19.4	1.19
117	725	C	2	H3	13.00	15.2	-19.8	1.60
117	725	C	2	H3	13.20	12.0	-18.9	1.59
117	725	C	2	H3	13.40	11.1	-19.4	1.38
117	725	C	2	H4	13.60	8.4	-19.6	1.30
117	725	C	2	H4	13.80	7.2	-19.5	1.23
117	725	C	2	H4	14.00	13.3	-19.7	1.44
117	725	C	2	H4	14.20	13.3	-20.1	1.66
117	725	C	2	H4	14.60	7.5	-19.8	0.53
117	725	C	2	H4	14.80	14.4	-20.0	1.04
117	725	C	2	H5	15.00	10.0	-19.7	1.06
117	725	C	2	H5	15.20	11.1	-19.6	1.46
117	725	C	2	H5	15.40	13.8	-18.1	0.97
117	725	C	2	H5	15.60	14.0	-20.0	1.96
117	725	C	2	H5	16.00	9.9	-19.3	1.11
117	725	C	2	H5	16.40	11.7	-19.0	1.50
117	725	C	2	H6	16.60	7.3	-18.6	0.89
117	725	C	2	H6	16.80	8.8	-20.0	0.50
117	725	C	2	H6	17.00	14.1	-19.3	0.52
117	725	C	2	H6	17.20	11.8	-19.8	0.72
117	725	C	2	H6	17.40	5.7	-20.4	0.65
117	725	C	2	H6	17.60	5.3	-17.6	0.63
117	725	C	2	H6	17.80	6.5	-19.6	0.78

117	725	C	2	H7	18.00	6.2	-20.5	0.68
117	725	C	2	H7	18.20	13.1	-20.2	0.65
117	725	C	2	H7	18.40	14.2	-19.9	0.68
117	725	C	3	H1	18.60	17.0	-20.0	0.55
117	725	C	3	H1	18.80	15.1	-20.1	0.62
117	725	C	3	H1	19.00	13.1	-20.0	0.67
117	725	C	3	H1	19.40	13.3	-19.8	0.95
117	725	C	3	H1	20.00	11.2	-19.1	0.73
117	725	C	3	H2	20.20	10.0	-19.9	1.28
117	725	C	3	H2	21.00	16.7	-20.1	0.68
117	725	C	3	H2	21.20	13.6	-20.1	0.63
117	725	C	3	H2	21.40	15.0	-20.2	0.61
117	725	C	3	H3	21.60	12.0	-22.5	0.50
117	725	C	3	H3	21.80	13.3	-20.1	0.79
117	725	C	3	H3	22.00	12.6	-20.4	0.69
117	725	C	3	H3	22.60	13.7	-20.1	0.62
117	725	C	3	H3	22.80	11.2	-20.6	0.68
117	725	C	3	H4	23.20	13.6	-19.9	0.81
117	725	C	3	H4	23.60	12.1	-18.8	0.71
117	725	C	3	H4	23.80	11.7	-19.9	1.25
117	725	C	3	H4	24.00	12.7	-19.6	0.55
117	725	C	3	H4	24.20	11.6	-19.9	0.91
117	725	C	3	H5	24.60	10.6	-19.7	0.65
117	725	C	3	H5	24.80	11.1	-20.1	0.17
117	725	C	3	H5	25.00	14.9	-20.0	1.02

117	725	C	3	H5	25.20	10.8	-19.3	1.30
117	725	C	3	H5	25.40	12.9	-20.1	0.93
117	725	C	3	H5	25.60	13.5	-20.6	0.51
117	725	C	3	H5	25.80	14.3	-21.1	1.34
117	725	C	3	H5	26.00	10.1	-20.1	0.95
117	725	C	3	H6	26.20	14.3	-20.8	0.67
117	725	C	3	H6	26.60	13.9	-20.8	1.42
117	725	C	3	H6	26.80	10.5	-20.0	0.92
117	725	C	3	H6	27.00	14.9	-20.2	0.72
117	725	C	3	H6	27.40	13.1	-20.3	0.95
117	725	C	3	H7	27.60	15.4	-21.9	0.54
117	725	C	3	H7	27.80	12.3	-20.1	0.81
117	725	C	3	H7	28.00	14.1	-21.8	0.81
117	725	C	3	H7	28.20	15.0	-21.2	1.08
117	725	C	4	X1	28.70	10.0	-25.0	0.23
117	725	C	4	X1	28.90	14.7	-23.7	0.37
117	725	C	4	X1	29.30	10.4	-23.4	0.41
117	725	C	4	X1	29.50	12.2	-24.2	0.41
117	725	C	4	X1	29.70	10.9	-25.6	0.21
117	725	C	4	X2	30.10	5.3	-21.9	0.73
117	725	C	4	X2	30.50	10.3	-21.6	1.47
117	725	C	4	X2	30.70	10.0	-22.7	0.55
117	725	C	4	X2	30.90	12.8	-20.5	2.08
117	725	C	4	X2	31.10	12.0	-21.6	1.06
117	725	C	4	X2	31.30	8.2	-19.8	1.03

117	725	C	4	X3	31.50	14.5	-20.6	1.28
117	725	C	4	X3	31.70	8.3	-20.4	0.98
117	725	C	4	X3	31.90	10.3	-21.6	0.48
117	725	C	4	X3	32.10	11.3	-20.4	0.32

Table 3: Percentages of Total Carbon (TC), Organic Carbon (OC), Total Nitrogen (TN), CaCO$_3$ Contents, Corrected OC and C/N Ratios (OC/T.N) for Site 724.

Depth	%C	CaCO3	%T.C	%OC	%T.N	OC/T.N	Cor %OC
0.25	6.41	53.41	8.16	1.75	0.22	7.97	2.82
0.45	7.20	60.00	8.75	1.54	0.18	8.73	2.60
0.65	6.90	57.50	7.79	0.89	0.11	8.05	1.06
0.85	6.99	58.25	7.90	0.91	0.12	7.64	0.97
1.05	6.87	57.25	8.02	1.15	0.12	9.43	0.60
1.25	6.43	53.58	6.87	0.44	0.07	6.61	0.57
1.45	6.25	52.08	6.83	0.58	0.08	6.93	0.40
1.65	6.21	51.75	6.74	0.53	0.07	7.64	0.66
1.85	6.07	50.58	6.75	0.68	0.09	8.04	0.85
2.05	5.81	48.41	7.04	1.23	0.13	9.42	1.71
2.25	6.09	50.75	7.12	1.03	0.10	10.23	1.33
2.45	5.97	49.75	7.15	1.18	0.11	10.35	1.63
2.90	6.38	53.16	7.16	0.78	0.07	11.16	1.22
3.05	6.04	50.33	6.81	0.77	0.07	11.46	0.64
3.25	6.03	50.25	7.22	1.19	0.11	10.46	0.61
3.45	6.02	50.16	7.62	1.60	0.14	11.61	
3.65	5.85	48.75	6.52	0.67	0.07	9.12	1.59
3.85	5.30	44.16	7.28	1.98	0.17	11.77	0.59
4.05	3.66	30.50	7.51	3.85	0.32	12.22	2.38

4.25	5.82	48.50	6.65	0.83	0.07	12.19	2.53
4.45	5.72	47.66	6.75	1.03	0.09	10.97	0.43
4.65	5.44	45.33	7.11	1.67	0.14	12.19	0.56
4.85	5.70	47.50	7.35	1.65	0.14	11.56	1.14
5.05	6.16	51.33	7.88	1.72	0.16	11.03	
5.25	5.97	49.75	7.07	1.10	0.10	11.53	2.16
5.45	5.59	46.58	7.03	1.44	0.12	11.59	0.96
5.65	6.13	51.08	7.31	1.18	0.11	10.69	0.69
5.85	6.12	51.00	6.66	0.54	0.06	9.12	1.33
6.05	6.32	52.66	7.04	0.72	0.08	9.64	0.32
6.25	6.31	52.58	6.96	0.65	0.07	10.06	0.14
6.45	6.40	53.33	6.91	0.51	0.06	8.52	0.49
6.65	6.26	52.16	7.04	0.78	0.08	9.70	0.80
6.85	6.25	52.08	6.91	0.66	0.08	8.74	0.77
7.05	6.10	50.83	6.96	0.86	0.09	9.74	0.79
7.25	5.34	44.50	6.93	1.59	0.13	11.85	1.18
7.45	6.29	52.41	7.40	1.11	0.11	10.44	1.70
7.65	6.61	55.08	8.16	1.55	0.13	12.23	1.04
7.85	7.48	62.33	9.29	1.81	0.16	11.09	1.50
8.05	6.96	58.00	8.10	1.14	0.10	11.18	2.23
8.25	6.79	56.58	8.06	1.27	0.11	11.33	1.04
8.45	7.15	59.58	8.79	1.64	0.14	11.90	1.35
8.65	7.54	62.83	8.74	1.20	0.11	10.59	1.33
8.85	7.32	61.00	8.39	1.07	0.10	10.42	1.76
9.05	7.20	60.00	8.53	1.33	0.12	11.24	1.36

9.25	7.25	60.41	8.44	1.19	0.10	11.44	1.09
9.45	6.96	58.00	7.88	0.92	0.09	10.28	1.63
9.65	6.73	56.08	7.41	0.68	0.07	9.88	0.70
9.85	6.79	56.58	7.27	0.48	0.06	8.64	0.89
10.05	6.70	55.83	7.24	0.54	0.04	12.25	1.24
10.25	6.55	54.58	6.99	0.44	0.04	10.23	0.52
10.45	6.29	52.41	6.97	0.68	0.08	8.66	0.39
10.65	6.32	52.66	6.91	0.59	0.05	12.31	0.52
10.85	6.26	52.16	6.91	0.65	0.05	12.54	0.39
11.05	6.16	51.33	6.95	0.79	0.07	12.20	0.68
11.25	6.12	51.00	6.76	0.64	0.18	3.55	0.69
11.45	5.99	49.91	6.86	0.87	0.07	12.55	0.74
11.65	5.75	47.91	7.17	1.42	0.11	12.99	0.87
11.85	5.92	49.33	6.62	0.70	0.08	8.71	2.06
12.05	5.85	48.75	6.73	0.88	0.07	13.29	1.50
12.25	5.75	47.91	7.00	1.25	0.09	13.61	0.97
12.40	5.83	48.58	6.73	0.90	0.07	12.63	0.75
12.45	6.04	50.33	7.23	1.19	0.09	13.50	1.11
12.65	6.09	50.75	7.16	1.07	0.08	12.79	1.62
12.85	6.09	50.75	7.00	0.91	0.07	12.97	1.26
13.05	6.13	51.08	6.94	0.81	0.07	11.57	0.86
13.25	6.17	51.41	7.04	0.87	0.07	12.47	1.04
13.45	6.18	51.50	6.89	0.71	0.07	10.80	2.21
13.65	6.22	51.83	6.90	0.68	0.06	10.64	0.73
13.85	6.17	51.41	6.91	0.74	0.07	11.09	0.91

14.05	6.18	51.50	6.68	0.50	0.05	10.68	0.90
14.25	6.19	51.58	6.67	0.48	0.05	10.69	1.36
14.45	6.05	50.41	6.49	0.44	0.05	8.96	0.41
14.65	5.99	49.91	6.48	0.49	0.05	9.63	0.65
14.85	5.91	49.25	6.41	0.50	0.05	10.44	0.59
15.05	5.84	48.66	6.49	0.65	0.06	11.73	0.69
15.25	5.82	48.50	6.64	0.82	0.07	11.90	0.77
15.45	5.57	46.41	6.52	0.95	0.08	12.68	0.99
15.65	5.50	45.83	6.81	1.31	0.10	13.32	0.92
15.85	5.61	46.75	6.43	0.82	0.07	12.36	1.57
16.05	4.85	40.42	6.50	1.65	0.12	14.10	1.14
16.25	5.49	45.75	6.22	0.73	0.05	14.39	
16.45	5.58	46.50	6.24	0.66	0.06	11.03	2.76
16.65	5.60	46.66	6.22	0.62	0.06	10.28	1.24
16.85	5.51	45.91	6.26	0.75	0.06	13.14	0.48
17.05	4.55	37.92	6.32	1.77	0.12	14.66	0.78
17.25	5.51	45.91	6.19	0.68	0.06	12.00	0.89
17.45	5.59	46.58	6.29	0.70	0.06	12.48	1.32
17.65	4.95	41.25	6.61	1.66	0.12	13.73	0.93
17.85	5.46	45.50	6.33	0.87	0.07	12.43	0.79
18.05	5.19	43.25	6.42	1.23	0.09	13.39	0.96
18.25	5.83	48.58	6.75	0.92	0.08	12.25	1.16
18.45	5.90	49.16	6.92	1.02	0.09	11.38	1.14
18.65	5.87	48.91	6.84	0.97	0.07	13.89	1.48
18.85	5.99	49.91	6.71	0.72	0.06	11.87	1.13

19.05	6.09	50.75	6.68	0.59	0.08	7.17	0.99
19.25	6.00	50.00	6.86	0.86	0.08	11.12	0.86
19.45	5.81	48.41	6.97	1.16	0.10	12.23	0.99
19.65	5.79	48.25	7.20	1.41	0.11	12.92	1.01
19.85	7.46	62.16	9.00	1.53	0.16	9.65	0.72
20.05	6.70	55.83	8.08	1.38	0.12	11.71	1.17
20.25	6.92	57.66	8.68	1.76	0.13	13.40	1.52
20.45	8.08	67.33	9.53	1.45	0.12	12.19	0.90
20.65	7.13	59.41	8.20	1.07	0.10	11.17	1.43
20.85	6.92	57.66	7.76	0.84	0.08	10.32	1.51
21.05	6.63	55.25	7.37	0.74	0.07	10.61	1.79
21.25	6.60	55.00	7.22	0.62	0.06	10.03	0.96
21.45	6.42	53.50	7.73	1.31	0.11	12.51	0.88
21.65	6.60	55.00	7.93	1.33	0.11	11.77	1.11
21.85	6.85	57.08	7.50	0.65	0.07	9.27	1.18
21.83	6.88	57.33	7.79	0.91	0.09	10.41	1.58
21.95	6.57	54.75	7.20	0.63	0.06	9.87	0.74
22.15	6.01	50.08	6.70	0.69	0.06	11.00	1.10
22.35	5.67	47.25	6.70	1.03	0.09	11.58	0.89
22.55	5.38	44.83	6.91	1.53	0.13	12.16	0.66
22.75	5.86	48.83	7.36	1.50	0.12	12.79	1.35
22.95	5.38	44.83	6.53	1.15	0.09	12.52	1.98
23.15	5.75	47.91	6.92	1.17	0.10	11.59	2.12
23.35	5.67	47.25	7.34	1.67	0.13	12.66	1.10
23.55	5.82	48.50	6.62	0.80	0.07	10.85	1.59

23.75	5.90	49.16	7.17	1.27	0.10	12.19	1.90
23.95	5.81	48.41	6.76	0.95	0.09	11.16	1.17
24.15	6.30	52.50	6.99	0.69	0.06	10.87	1.35
24.35	6.48	54.00	7.04	0.56	0.06	9.25	1.25
24.55	6.56	54.66	7.04	0.48	0.05	8.87	0.66
24.75	6.52	54.33	7.29	0.77	0.07	10.97	0.73
24.95	6.67	55.58	7.08	0.41	0.05	7.59	0.45
25.15	6.77	56.41	7.11	0.34	0.05	7.02	0.17
25.35	6.70	55.83	7.18	0.48	0.05	9.04	0.54
25.85	6.57	54.75	7.58	1.01	0.09	11.01	0.13
25.75	6.56	54.66	7.56	1.00	0.09	11.14	0.71
25.95	6.33	52.75	7.51	1.18	0.10	11.80	1.15
26.15	6.74	56.16	7.97	1.23	0.11	11.28	1.15
26.35	7.26	60.50	8.55	1.29	0.11	11.80	1.32
26.55	7.47	62.25	8.71	1.24	0.11	11.06	0.97
26.75	7.21	60.08	8.26	1.05	0.09	11.28	2.31
26.95	6.96	58.00	8.07	1.11	0.09	12.29	1.38
27.15	6.91	57.58	8.03	1.12	0.10	10.73	1.18
27.35	7.07	58.91	8.66	1.59	0.13	11.88	0.98
27.55	6.28	52.33	7.92	1.64	0.12	13.23	1.58
27.75	7.80	65.00	9.73	1.93	0.16	12.07	1.17
27.95	7.01	58.41	8.14	1.13	0.09	11.99	2.07
28.15	6.74	56.16	7.53	0.79	0.07	10.75	2.68
28.35	7.16	59.66	7.91	0.75	0.07	10.27	0.98
28.55	7.05	58.75	7.47	0.42	0.05	7.98	0.04

28.75	7.18	59.83	7.96	0.78	0.07	11.70	1.68
28.95	6.87	57.25	7.19	0.32	0.04	7.32	0.34
29.15	6.81	56.75	7.19	0.38	0.05	8.33	0.40
29.35	6.47	53.91	6.81	0.34	0.04	8.60	0.38
29.55	6.52	54.33	6.86	0.34	0.05	7.30	0.37
29.75	6.30	52.50	6.71	0.41	0.04	10.02	0.12
29.95	6.10	50.83	6.76	0.66	0.07	10.14	0.26
30.15	6.11	50.91	6.98	0.87	0.08	10.86	0.43
30.35	6.19	51.58	7.10	0.91	0.08	11.77	0.62
30.55	6.28	52.33	7.11	0.83	0.08	10.97	1.03
30.85	6.03	50.25	6.95	0.92	0.08	11.30	0.94
31.05	6.46	53.83	7.14	0.68	0.09	7.78	1.02
31.25	6.48	54.00	7.31	0.83	0.08	11.11	0.90

Table 4: Comparison of the means between Sites 724 and 725 for unequal variance.

	$\delta^{15}N$	$\delta^{13}C$	%OC
t	11.200	2.850	36.130
DF	152.000	137.000	700.000
$t_{0.05}$	1.976	1.977	1.963

Difference in means is statistically significant at 5% level when t-value is greater than critical value ($t_{0.05}$).

Table 5: Average Values of Total Organic Carbon (OC), Total Nitrogen and Residue OC for each Oxygen Isotope Stage (OIS) for Site 724.

OIS	Total OC		Total N		Residue OC	
	Mean	Std	Mean	Std	Mean	Std
1	1.25	0.334	0.15	0.042	3.76	2.09
2	0.806	0.287	0.09	0.022	2.04	0.89
3	1.283	0.764	0.115	0.06	1.98	1.31
4	1.091	0.364	0.102	0.022	2.54	0.93
5	1.028	0.408	0.095	0.033	2.77	1.29
6	0.834	0.287	0.073	0.027	2.12	1.04
7	1.033	0.363	0.088	0.026	2.27	0.76
8	0.891	0.382	0.081	0.024	2.15	1.12
9	1.016	0.451	0.09	0.032	2.81	1.68
10	0.716	0.202	0.07	0.015	1.40	0.73

Table 6: Pearson (R) correlation coefficients.

(a) Site 724

	$\delta^{15}N$	$\delta^{13}C$	%OC
$\delta^{15}N$	1.000	0.096	0.266
$\delta^{13}C$	0.096	1.000	0.144
%OC	0.266	0.144	1.000

Degrees of Freedom (DF) 802

Critical Value at 95% confidence level 0.069

(b) Site 725

	$\delta^{15}N$	$\delta^{13}C$	%OC
$\delta^{15}N$	1.000	0.027	0.002
$\delta^{13}C$	0.027	1.000	0.319
%OC	0.002	0.319	1.000

Degrees of Freedom (DF) 122

Critical Value at 95% confidence level 0.176

Although R-values at Site 724 are greater than critical value at 95% confidence level, there is no significant relationship. This is due to a large sample size. Less than 7% of the variance can be explained by the dependent variable.

Table 7: Stable Carbon and Nitrogen Isotope Compositions of Terrestrial Tree Leaves and Grasses.

BOTANICAL NAME.	δ^{15}N	δ^{13}C
TREE LEAVES.		
OCHNACEAE *Ochua holostii Engl*.	8.62	-27.76
APOCYNACEAE *Landolophia Kurkii*	3.69	-29.59
ANACARDIACEAE *Sorindeia Usambarensis*	3.90	-30.31
Haplocoelum Mombasence	8.37	-31.85
OCHNACEAE *Ochina Mosambisiensis*	4.82	-29.90
ANNONACEAE *Annona Senegalensis*	3.20	-29.29
TILIACEAE *Grewia Conocorpa*	1.02	-29.20
Afzelia quensensi	5.37	-26.75
Baphia sp.	7.60	-29.63
GRASSES.		
Chloris Virgata	5.08	-14.10
Dactyloctenium geminatum	9.11	-11.68
Phragmites Mauritianus	10.20	-23.85
Panicum trichocladum	9.42	-12.89
Dichanthium annolatum	10.29	-11.58
Hyparrhenia rufa	5.68	-12.09
Brachiaria brazantha	4.47	-11.39
Panicum maxum	10.13	-11.68
Urochloa pullulans	2.04	-12.18

Botlunclog (?)	10.10	-11.17
Aristidia andoensis	7.92	-12.41

APPENDIX C

I, Thomas F. Pedersen , being the copyright holder of the material described below:

> Analytical data from ODP Site 724, Arabian Sea, as follows:
> (a) Oxygen and carbon stable isotope analyses of foraminifera;
> (b) Organic carbon, carbonate and nitrogen measurements on bulk sediments.

do hereby permit the inclusion of the described material in the thesis entitled:

> Late Pliocene-Quaternary History of the Northwestern Indian Ocean: An Organic Geochemistry Perspective.

written by Alfred Muzuka and submitted in partial fulfillment of the requirements for the degree of Master of Science at Memorial University of Newfoundland.

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