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THE BENGUELA ECOSYSTEM PART I. EVOLUTION OF THE BENGUELA, PHYSICAL FEATURES AND PROCESSES

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INTRODUCTION

The Benguela is one of the four major eastern boundary current regions of the World ocean and the oceanography of the western coast of Africa south of about 15° S, like that off California, Peru and North West Africa is dominated by a coastal upwelling system. Research in the area during the nineteenth and first part of the twentieth century was directed mainly at taxonomy, at improving navigational safety, and the development of fisheries. In spite of the observations of Ross (1847) early workers such as Muhry (1862) and Petermann (1865) viewed the Benguela Current as a northward extension of the West Wind Drift and it was only during the 1920s that Meyer (1923) showed conclusively that they are separated by a well-defined convergence zone, the Subtropical Convergence. After World War II a detailed study of the oceanography of southern African west coast waters commenced. The literature prior to 1970 was largely descriptive in nature, and it is only during the last fifteen years that mesoscale processes (*i.e.* processes occurring over spatial scales of tens of kilometres to a few hundred kilometres with time scales of hours to a few days) have been given serious attention.

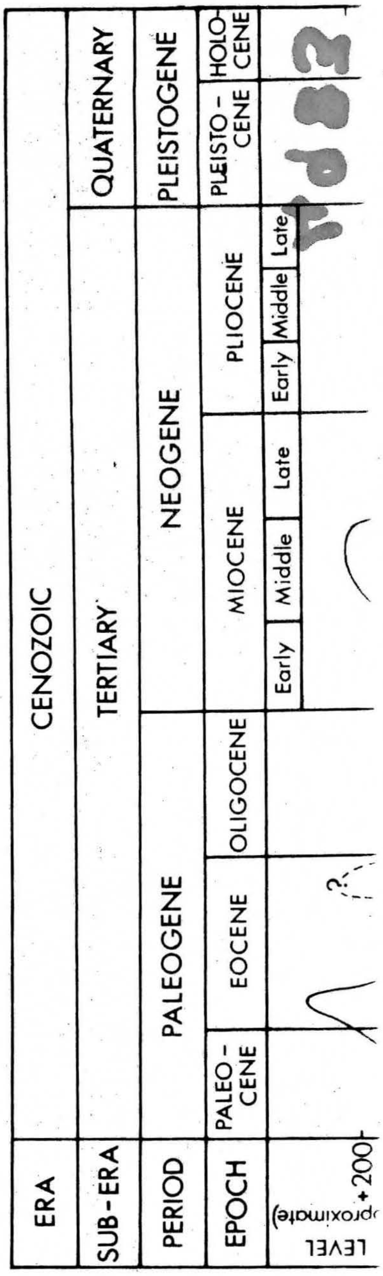
This review will take a broad view of the status of the Benguela ecosystem and the main thrust will be aimed at reviewing the present state of knowledge of the main processes governing the system. It has been divided into parts, *viz.* Part I which deals with the evolution and physics of the Benguela system, and which is the subject of this paper, Part II (Chapman & Shannon, 1985, also appearing in this volume) which deals with chemical processes, and Part III will address the various biological processes. Part III is in preparation and will appear in Volume 24 of *Oceanography and Marine Biology: An Annual Review* (1986) with L. V. Shannon, J. G. Field and W. R. Siegfried as authors. Part IV is also in preparation. Part I has been subdivided into a number of sections covering the origin of the Benguela system, large scale features, meteorology, seasonal and inter-annual variability and mesoscale physical processes. Two important papers on the physical oceanography of the region appeared during 1983 (Nelson & Hutchings, 1983; Parrish, Bakun, Husby &

Nelson, 1983) and the present article is intended to complement and build on the work of these authors. Processes in the nearshore region, *e.g.* in the intertidal zone and the surf zone, will not be considered in this review except where they are important for the broader understanding of the system as a whole.

In a review of a complex ecosystem such as that of the Benguela it is difficult, if not impossible to cite all relevant literature without making the paper cumbersome and unreadable. Accordingly the list of references, while being reasonably comprehensive, is by no means exhaustive. It should, however, be adequate to lead into the remainder of the available literature on the Benguela ecosystem and into the more specialized disciplinary literature. Readers are also referred to the comprehensive bibliography on the physical oceanography of the South East Atlantic compiled by Lutjeharms, van Ballegooyen & Valentine (1981).

EVOLUTION OF THE BENGUELA CURRENT AND UPWELLING SYSTEM

The South Atlantic Ocean began to form about 130 Myr (million year) ago (Kennett, 1982; Dingle, Siesser & Newton, 1983) following the rifting of Gondwanaland and the separation of the South American and African plates. Van Zinderen Bakker (1975) has suggested that it may have taken until mid Tertiary times before the ocean had sufficient zonal width to permit the development of its present circular wind and oceanic circulations. Kennett (1982) has indicated that the anticyclonic gyre was well established in the South Atlantic by the Eocene, which, if correct, implies that an equatorward flow (at the surface) along the west coast of southern Africa has existed at least for the past 50 Myr. This early current, the precursor of the South East Atlantic Drift, had however little in common with the eastern boundary current regime off southwestern Africa as we know it today—fossil evidence from the Eocene marine transgression implies that warm water conditions prevailed in the South East Atlantic Ocean during the early to middle Tertiary (Ward, Seely & Lancaster, 1983). A number of important geological and climatic events were necessary before the present thermohaline circulation in the World ocean could be established (see Fig. 1). During the Paleocene (65 to 55 Myr ago) Australia and Antarctica were joined. They began to drift apart 55 Myr ago (Kennett, 1978) but circum-Antarctic flow was blocked by the South Tasman Rise and Tasmania. During the Eocene the Southern Ocean was relatively warm and Antarctica largely non-glaciated, and according to Kennett (1978) a major climatic boundary was crossed at the Eocene-Oligocene boundary (38 Myr ago) when Antarctic sea ice began to form and a drop of 5 °C in the temperature of bottom waters resulted. Kennett (1978) has suggested that this threshold was crossed because of the gradual isolation of Australia and Antarctica and perhaps the opening of the Drake Passage. Furthermore, by the early Oligocene the opening of the Atlantic had progressed to a point where the South Atlantic was open to Antarctic bottom waters (Deacon, 1983). This implies that the thermohaline circulation as we know it today could not have been initiated until after the onset of the Oligocene period, *i.e.* more recently than 38 Myr ago. Other major events which must have significantly



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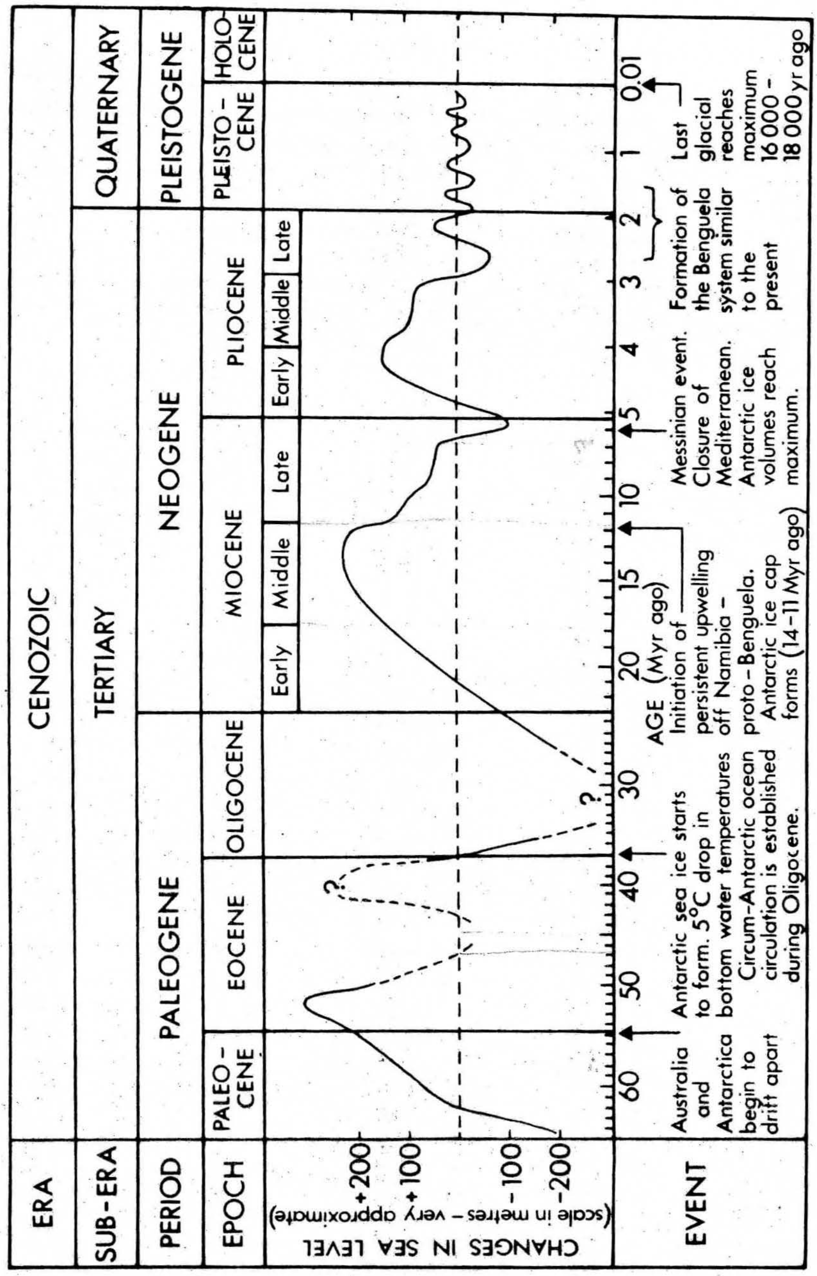

Fig. 1.—Cenozoic time scales, important events and sea level fluctuations around southern Africa from work of Hendey (1983), Deacon (1983), Kennett (1978) and others.

influenced circulation in the Atlantic Ocean were the closure of the mid-American seaway and the formation of the Antarctic ice-cap during the mid to late Miocene (Kennett, 1978) and the complete closure of the Mediterranean basin and the desiccation of this sea between 6·2 and 5·3 Myr ago (Van Zinderen Bakker, 1978)—ice volumes in the Antarctic were greater at the Miocene-Pliocene boundary (5 Myr ago) than today (Kennett, 1978). As a consequence of the substantial cooling during the mid to late Miocene, separate water masses originated in the Southern Ocean and the Subtropical and Antarctic Convergences were formed, with the result that the climate of southern Africa changed completely (Coetzee, 1978).

A dominance of arid to semi-arid conditions throughout the history of the Namib desert, which dates back to the Cretaceous, are revealed by Mesozoic-Cenozoic stratigraphic records (Ward *et al.*, 1983), with a desert sand sea in the southern and central Namib existing from early to mid-Tertiary times, and an internal sedimentary structure of the dunes reflecting a dominant southerly palaeo-wind regime, similar to the present wind patterns, which lasted for 20 to 30 Myr. Wind strengths must have been relatively low, however, as a consequence of low temperature gradients between the equator and the pole. Thus, although this suggests that a driving mechanism for upwelling might have existed along the western coast of southern Africa 50 to 20 Myr ago, the development of a proto-Benguela system could not take place until after the present thermo-haline circulation in the South Atlantic and Southern Oceans was established during the late Miocene. While upwelling probably existed during early times the water which was upwelled would have had very different temperature, salinity, and nutrient characteristics to those of late Tertiary and Quaternary upwelling waters. It is considered that the early upwelling would have appeared as tongues rather than as a continuous Ekman drift process. Consequently prior to the late Miocene, the Benguela region would not have supported flora and fauna characteristic of present-day eastern boundary current upwelling regimes.

A feature of the Cenozoic era has been the equatorwards movement of the westerly wind belt with important consequences for the contraction of the subtropical ocean gyre systems and meridional heat transfer, although throughout the era southern Africa did not at any stage lie within this belt of westerly winds (Deacon, 1983). It was the equatorwards movement of the westerlies and the growth of the Antarctic ice-cap that focused atmospheric subsidence in the present belt of subtropical high pressure cells—which was the fundamental cause of mid-latitude aridity and Mediterranean-type climates bordering the arid belt (Deacon, 1983). According to Ward *et al.* (1983) the aridification of the Namib desert since the late Tertiary, has been a progressive process, and these authors do not accept Siesser's (1978) contention that the aridification was initiated by the onset of major persistent upwelling during the late Miocene, for a number of reasons. Nevertheless the geological evidence seems to indicate that the southerly palaeo-winds were intensified over much of the Namib region from the late Tertiary and have persisted through the Quaternary sub-era.

Siesser (1978, 1980) and Diester-Haass & Schrader (1979) have postulated a late Miocene origin for strong persistent Benguela upwelling off northern Namibia, based on Deep-Sea Drilling Project cores near to the abutment of the Walvis Ridge with the African continent (site 362/362A—approximate

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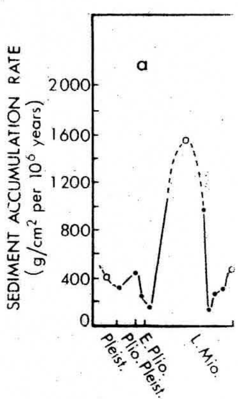

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latitude 20° S). More recent studies by Meyers *et al.* (1983) on cores taken by the GLOMAR CHALLENGER in a similar area at DSDP 530 and 532, support the earlier findings of Siesser (1980) and others that date the onset of upwelling as late Miocene. Diester-Haass & Schrader (1979) considered that the greater abundance of siliceous microfossils in the Pleistocene might have been due to an intensification of cold water current patterns and to higher wind velocities (the latter being deduced from the high abundance of displaced continental opal remains). The view of Seisser (1978) was that upwelling off northern Namibia was weak and spasmodic from the mid or late Oligocene to mid Miocene period, and intensified during the late Miocene (12 Myr ago). Siesser (1980) presented a plausible argument for the late Miocene origin of northern Benguela upwelling, and suggested that the enormous blooms of a single nanoplanktonic organism, *Braarudosphaera*, reflected in the high accumulation rates during Oligocene times were more probably a response to a regional South Atlantic event rather than to local Benguela upwelling. The sediment deposit rates and organic carbon contents from the core analysis are shown in Figure 2. Although these data related only to the northern Namibian Benguela system, Siesser (1980) argued intuitively that it was probable that upwelling would begin all along the coast with only a slight lag from place to place, once the forcing mechanisms were established. The palaeo-climatic and palaeo-ecological records of the southern part of the Benguela system (around the South West Cape) strongly suggest, however, that persistent upwelling was not established in the region until the late Pliocene, about which time, from the work of Siesser (1980) and Meyers *et al.* (1983), productivity in the northern Benguela reached a maximum.

Siesser's paper

In a recent article on fossil sea-birds from early Pliocene (5 Myr ago) deposits in the South Western Cape, Olson (1983) found that the region had a more sub-Antarctic marine environment then than at present, and that the marine avifauna of the southern Benguela area has changed drastically since the early Pliocene. Studies by Tankard & Rogers (1978), Coetzee (1978),

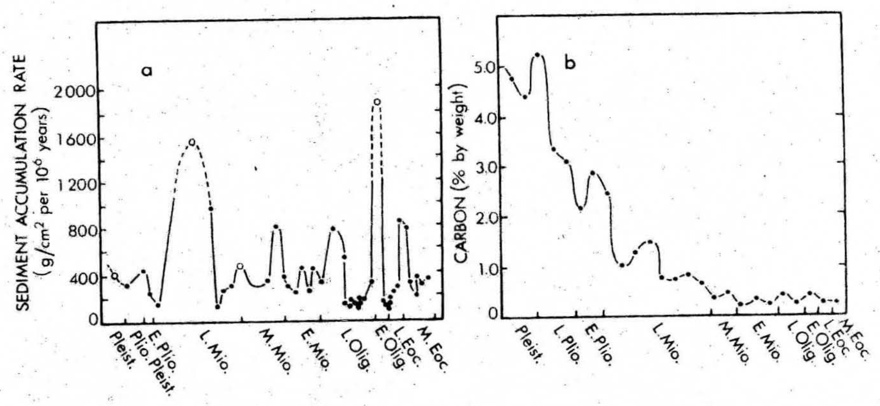

Fig. 2.—a, total corrected sediment accumulation rates for DSDP site 362/362A (after Siesser, 1980). b, organic carbon content in cores from DSDP site 362/362A (after Siesser, 1980).

Hendey (1983) and Coetzee, Scholtz & Deacon (1983) on pollens and vertebrate fauna from the late Tertiary and Quaternary fossil deposits in the South Western Cape indicate that the climate of the region during the early Pliocene was in an intermediate stage, between that of the Miocene and the present. Hendey (1983) considered that a summer wet–winter dry rainfall pattern was characteristic of the early Pliocene in the South Western Cape. The present arid summer–wet winter Mediterranean type climate in the southern Benguela region only appears to have been fully established towards the end of the Pliocene, *i.e.* about 2 Myr ago (Tankard & Rogers, 1978; Hendey, 1983). At this time there was a sharp increase in the organic carbon content of marine sediment cores from northern Namibia (Siesser, 1980, see also Fig. 2). Also by this time the Agulhas Current was similar to its present form, its quasi-modern flow patterns having been established about 5 Myr ago in the early Pliocene (Martin, 1981). Thus, although there is evidence of equatorward flow in the South East Atlantic and sporadic coastal upwelling off Namibia since the Oligocene epoch, and although a proto-Benguela system appears to have been initiated off Namibia during the late Miocene (Seisser, 1980), the Benguela system as we know it today is clearly of more recent origin, dating from the late Pliocene.

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Marine transgressions and regressions during the history of the Benguela system (*e.g.* Siesser & Dingle, 1981) would have had a significant impact on the upwelling regimes and current patterns. Sea level fluctuations since the Paleocene epoch from Hendey (1983) have been included in Figure 1. High sea levels during the mid Miocene and early to mid Pliocene times would have resulted in the present continental shelf being 100 to 200 m deeper, with the drowning of lower lying coastal areas and the formation of coastal islands (*e.g.* the Cape Peninsula) and an irregular coast having numerous embayments. Conversely, at the Miocene–Pliocene boundary and also during the mid to late Pliocene the lower sea levels would have exposed large areas of shelf, *e.g.* St Helena Bay. These bottom topographic and orographic changes would obviously have had a major effect on the positions of upwelling centres, the oceanic front and the general dynamics of the Benguela system.

glaciations

A feature of the modern global climatic regimes initiated in the Pliocene and continuing through the Pleistocene has been the characteristic rhythm with a periodicity of about 100 000 years, linked to perturbations of the orbit of the earth relative to the sun, of cooler climates, *i.e.* glacials or hypothermals, interrupted by shorter periods of warmer climates, *i.e.* interglacials or hyperthermals (Deacon, 1983). In the Benguela regime the coldest interval of the late Pleistocene, with the most severe climatic conditions, was between 16 000 and 18 000 yr ago, and temperatures approximated to those of the present from about 12 000 yr ago (Deacon, 1983). The late Cenozoic palaeoenvironments on the western coast of southern Africa have been discussed in some detail by Van Zinderen Bakker (1975), Tankard & Rogers (1978), and others. During glacial episodes the atmospheric pressure gradient between Antarctica and the equator would have been steepened with the resultant intensification of atmospheric and oceanic circulation. During the last hypothermal, the South Atlantic anticyclone was situated 5° further north (Van Zinderen Bakker, 1975) and cold polar air could penetrate the interior of southern Africa up to about 24° S (Coetzee *et al.*, 1983). The consequence of this would have been an intensification of upwelling and for the Benguela system to shift further north. Van Andel & Calvert (1971) suggested in fact that the Benguela Current was intensified during glacials in order to account for the increased erosion they observed on the Namibian shelf. In support of the concept of a northward shift, Bornhold (1973) suggested that there was a northward extension of the Benguela regime along the coast of Angola and an onset of upwelling off the Congo and Gabon, as deduced from the analysis of Angola Basin sediments. To what extent the Walvis Ridge would have acted as a barrier to this hypothesized northward extension of the system is a matter for speculation. The inferred atmospheric and oceanic systems around southern Africa during glacials and interglacials are illustrated in Figure 3, which is

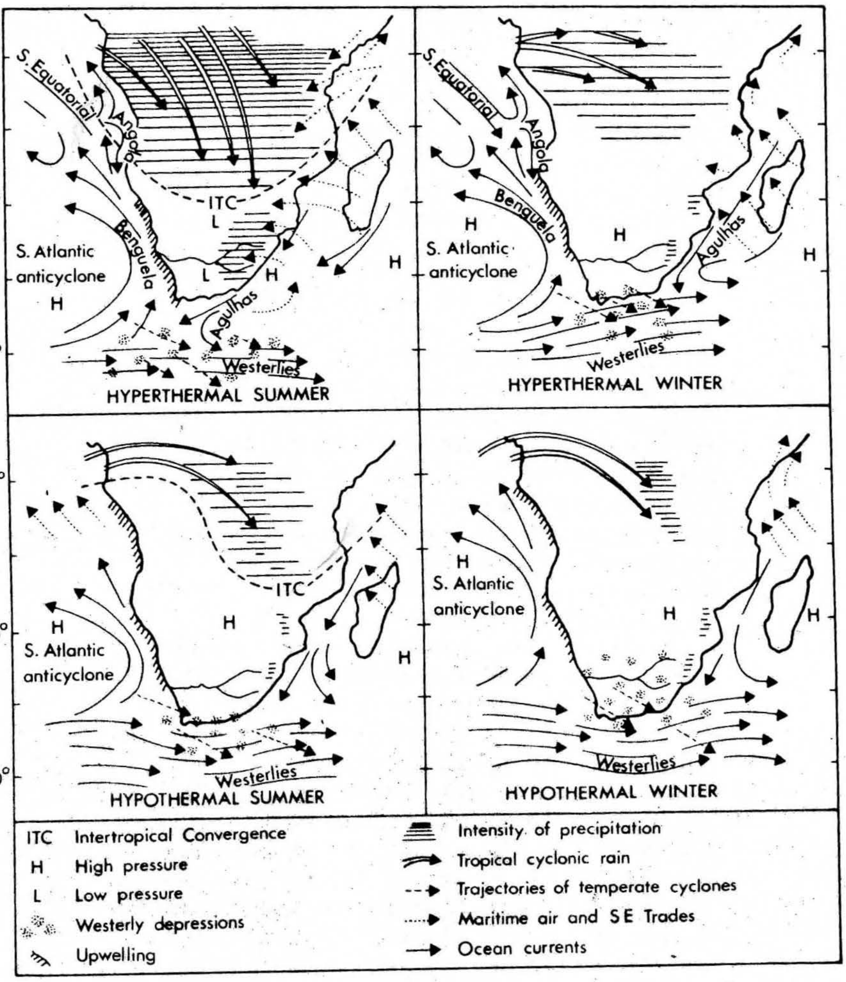

Fig. 3.—Atmospheric and oceanic circulation patterns during hyperthermal (interglacial) and hypothermal (glacial) times (after Van Zinderen Bakker, 1976; Tankard & Rogers, 1978).

from Van Zinderen Bakker (1976). It should be noted, however, that in his most recent interpretation of African palaeo-environments during the last glacial (Van Zinderen Bakker, 1982) this author ascribed the changes as being a consequence of worldwide cooling and a strengthening of the circulation systems rather than to a latitudinal shift in climatic belts. Deep-sea cores from the northern Benguela show a distinct layering of organic carbon concentrations corresponding to cycles of 30 000 to 50 000 years (Meyers *et al.*, 1983) which reflect changes in upwelling intensity in the area and/or of sea level.

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To summarize, therefore, while there is evidence of sporadic upwelling occurring during the Oligocene, the proto-Benguela system off Namibia was not initiated until the late Miocene. Upwelling was intensified progressively during the Pliocene, and the full development of the Benguela system and its southward extension to the South Western Cape dates from late Pliocene to early Pleistocene. Cyclical perturbations induced during glacials and interglacials resulted in latitudinal shifts in the extent of the system and changes in the intensity of upwelling and currents.

TOPOGRAPHY

The Cape and Angola Basins which comprise the abyssal plain in the South East Atlantic Ocean are separated by the Walvis Ridge, which runs from its abutment with the coast at about latitude 20° S in a southwesterly direction for more than 2500 km towards the Mid-Atlantic Ridge (Fig. 4). The Walvis Ridge forms a barrier to the northward and southward flow of water below a depth of 3000 m (Shannon & van Rijswijck, 1969; Nelson & Hutchings, 1983) and, as will be discussed later, exerts a major influence on the circulation in the South East Atlantic. Prominent geological features of the Cape Basin, which is bounded in the south by the Agulhas Ridge, are the numerous seamounts of volcanic origin (*e.g.* Discovery, Vema).

The western coast of southern Africa, which forms the eastern boundary of the Benguela system, is characterized by a relatively narrow coastal plain which rises to the main continental escarpment, situated between 50 and 200 km inland (see Fig. 4). Much of the coastal region is arid. The Namib Desert extends between about 14° S and 31° S and is at its widest in the central region, which comprises the main Namib Sand Sea (Ward *et al.*, 1983). The major part of the coastal belt is characterized by sand dunes, with occasional rocky outcrops. North of 32° S the coastline is regular and, except for Walvis Bay and Lüderitz, is devoid of significant embayments. The Orange and Fish River valley forms a major break in the escarpment and continental plateau at the Namibia–South Africa boundary, while the valleys of the Olifants and Berg Rivers in the south, that of Kunene River in the north, and the courses of several dry Namibian rivers provide secondary discontinuities. South of 32° S the coastline is irregular with several capes (formed by granitic outcrops, *e.g.* Cape Columbine, Cape Peninsula, Cape Hangklip) and bays (*e.g.* St Helena Bay, Saldanha Bay, Table Bay, False Bay). This southern region, *viz.* the South Western Cape is topographically different from the remainder of the area and has a Mediterranean type climate and vegetation.

The bathymetry of the western continental margin of southern Africa is variable, with narrow parts of the continental shelf situated off southern

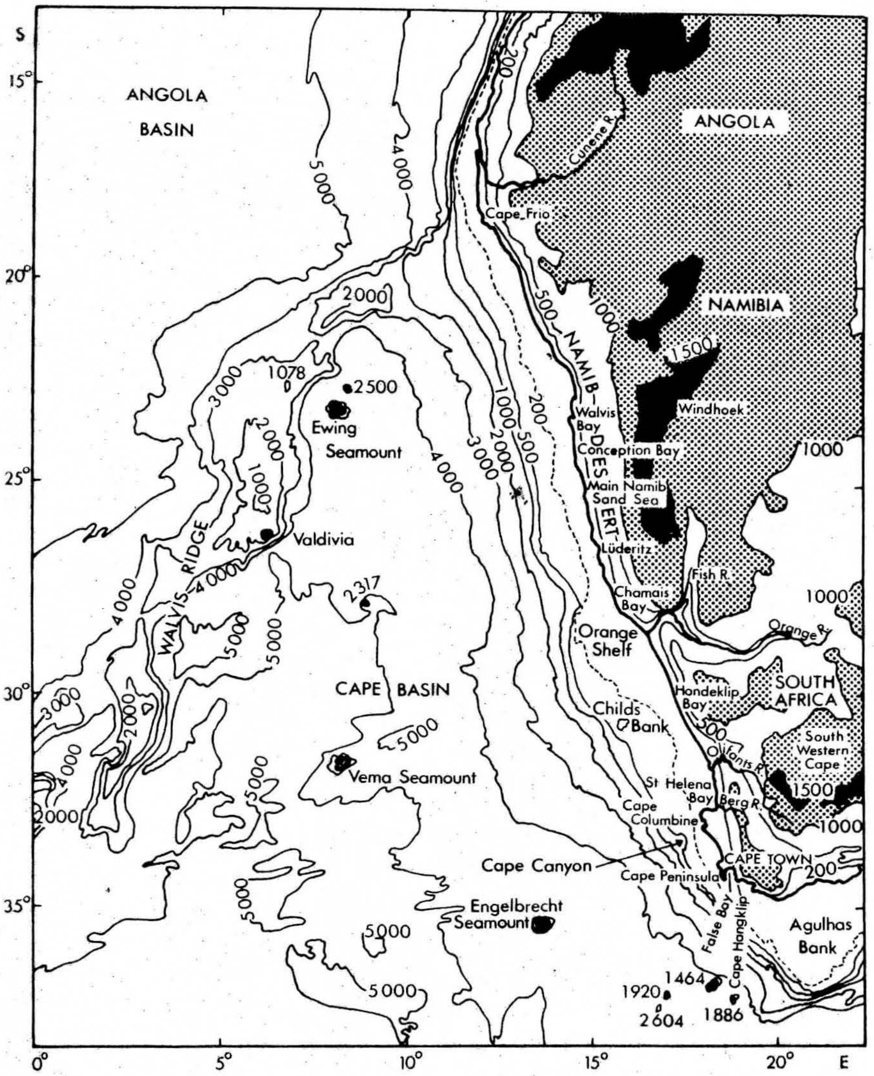

Fig. 4.—Bathymetry of the South East Atlantic Ocean and main orographical features of southwestern Africa.

Angola (20 km), south of Lüderitz (75 km) and off the Cape Peninsula (40 km) and the widest zones off the Orange River (180 km) and in the extreme south (Agulhas Bank). In the Kunene margin the shelf is narrow (about 45 km), about 200 m deep and the continental slope is relatively steep (Bremner, 1981). Between Cape Frio (18° S) and Chamais Bay (28° S), the Walvis Shelf, which is typically 140 km wide, is relatively deep on average, with the shelf break being at about 350 m on average (Birch, Rogers, Bremner & Moir, 1976). Double shelf breaks are, however, common off the west coast (Siesser, Scrutton &

Simpson, 1974), and around 23° S (Walvis Bay) there are very pronounced inner and outer breaks corresponding to depths of about 140 and 400 m, respectively (see Fig. 5). The broad inner shelf between 22° S and 23°30′ S probably plays an important rôle in the dynamics of the central Namibian region. Between Chamais Bay and Hondeklip Bay (30°30′ S) is the Orange Shelf which is at its widest, *viz.* 180 km, off the mouth of the Orange River. In this region the shelf break deepens from about 200 m in the north to 500 m in the south (Birch *et al.*, 1976). The outer shelf, *viz.* the Orange Bank, is shallow (160 m) while the mid shelf reaches 190 m in places. At about 31° S there is another shallow feature, Childs Bank, situated about 150 km offshore. Further south, in particular between 32° S and 35° S the coastline is irregular and the shelf is variable in width. Between 31° S and 33° S (Cape Columbine) there is an inner and an outer shelf break (200 to 380 m and 500 m, respectively) which merge south of 33° S to form a single, deep shelf break (500 m, Birch & Rogers, 1973). Between 31° S and 35° S several submarine canyons cut into the shelf, the most prominent of these being the Cape Canyon which is situated 60 km offshore between 33° S and 34° S. Its axis runs in a north–south direction, *i.e.* more or less parallel to the coast, and the canyon is thought to be a marine extension of the Olifants and Berg Rivers dating from the Palaeogene (Dingle & Hendey, 1984). The Agulhas Bank, a relatively wide and shallow feature, forms the southernmost margin of the continent. East–west bathymetric transects at selected localities in the Benguela region are shown in Figure 5, and these will be referred to in subsequent sections. For detailed charts of the bathymetry of the Benguela region and South East Atlantic, readers are referred to Dingle, Moir, Bremner & Rogers (1977), Rabinowitz, Shackleton & Brenner (1980), and Shackleton (1982).

The western continental margin of southern Africa between the Kunene River and Cape Agulhas is dominated by biogenic sedimentation and concomitant authigenesis resulting from the high productivity of the upwelled waters in the Benguela system (Birch *et al.*, 1976). Like the bathymetry, the composition and physical characteristics of the surficial sediments can be helpful in gaining a broad understanding of the dynamics of the system. The following very brief summary of the nature of the sea floor, sediment texture, calcium carbonate, and organic carbon content is based on Birch *et al.* (1976) and Birch & Rogers (1973), as is Figure 6. Much of the region, in particular in the south and between Lüderitz and the Orange River has a rocky bottom while there are substantial areas with sparse sediment cover. The sediments mantling the western continental margin form textural zones parallel to the

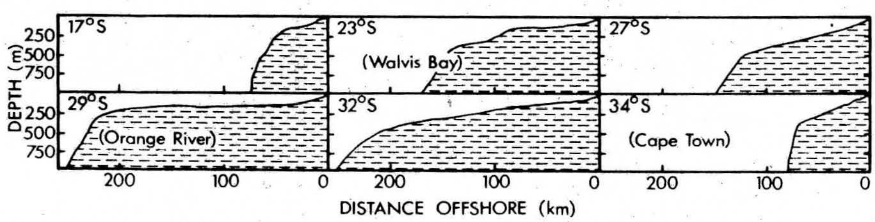

Fig. 5.—Shelf profiles at selected latitudes.

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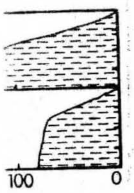

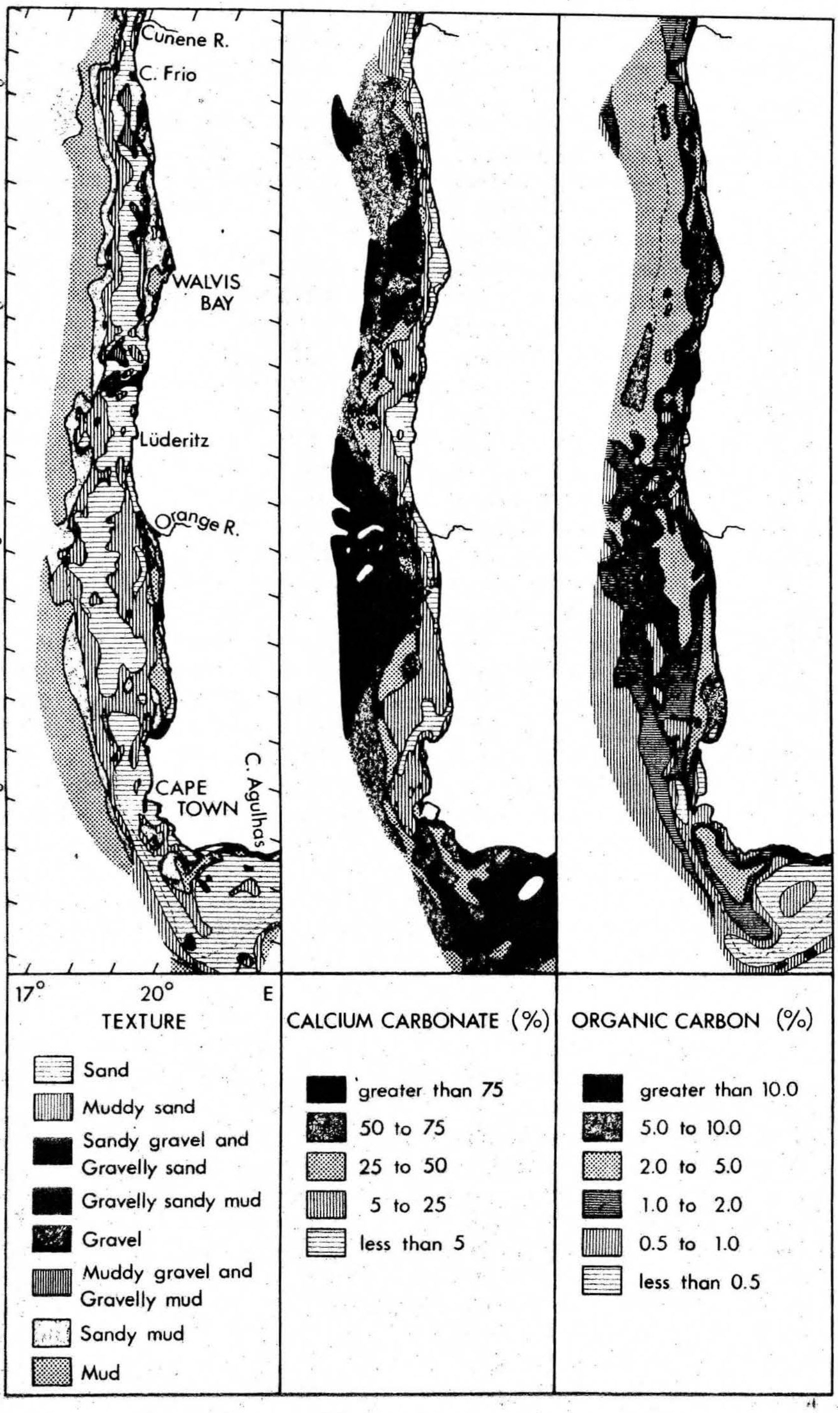

Fig. 6.—Sediment texture, calcium carbonate and organic carbon content along the west coast (from Birch *et al.*, 1976).

coast, and in general the sediments become finer seawards, changing from sands on the inner and middle shelves to muddy sands and sandy muds on the outer shelf and muds on the outer continental slope (Birch *et al.*, 1976). This pattern has, however, been considerably modified by localized river input and biological deposition. Significant features of the shelf in the Benguela region are the two extensive mud belts, each almost 500 km long. The southern belt which extends between the Orange and Olifants Rivers, up to 40 km wide and averaging 15 m thick, is situated over the outer edge of the middle shelf and is mainly of terrigenous (river) origin (Birch *et al.*, 1976). The northern belt which lies over the middle shelf between Cape Frio and Conception Bay comprises organic rich diatomaceous oozes ($>5\%$ organic carbon content). Other than two large deposits of dominantly fine sediment, the outer shelf is covered in sand-sized material (Birch *et al.*, 1976). According to these authors the calcareous flora and fauna which are the most important biogenic contributors to the shelf sediments in the Benguela region are mainly planktonic or benthic foraminiferans, while locally, siliceous phytoplankton dominate some near-shore sediments and polychaete worms have converted some of the terrestrially derived muds on the western coast into sand-sized faecal pellets. The outer and middle parts of the shelf are covered by carbonate-rich sediment ($>50\%$ $CaCO_3$). Birch *et al.* (1976) have shown that the highest organic carbon values on the shelf are related to the diatomaceous muds between Lüderitz and Cape Cross (15% C_{org}) and to the organic rich faecal pellet muds on the shelf break north of Lüderitz ($>5\%$ C_{org}), on the inner shelf and off Lamberts Bay ($>5\%$ C_{org}), and south of Cape Hangklip ($>2\%$ C_{org}). Sediment, depleted in organic carbon, in some regions is associated with deposits of glauconite and apatite mixed as pellets (Birch, 1979). The manner in which the organic-rich sediment curves eastward around the Agulhas Bank (see Fig. 6) is regarded by Birch *et al.* (1976) as an indication that this southernmost region lies within the sedimentological regime of the Benguela. Readers are referred to the following for further information: van Andel & Calvert (1971); Calvert & Price (1971a); Birch (1975, 1977, 1978); Scrutton & Dingle (1975); Rogers (1977); Bremner (1978, 1980a,b, 1981, 1983); Summerhayes, Bornhold & Embley (1979); Robson (1983). A synthesis of information about the sedimentology of the Benguela region is in preparation by Drs J. Rogers, J. M. Bremner and R. Johnson.

METEOROLOGY

S. Atl. anticyclone

The prevailing winds over the Benguela region are determined by the South Atlantic high pressure system (anticyclone), the pressure field over the adjacent subcontinent and by eastward moving cyclones across the southern part produced by perturbations on the subtropical jet stream (Schell, 1968; Nelson & Hutchings, 1983). The South Atlantic high is maintained throughout the year but undergoes seasonal shifts in position (it is approximately centred around 30° S : 5° E in summer and 26° S : 10° E in winter, Schell, 1968; van Loon, 1972a) and intensity (3 to 4 mb). The pressure over the African subcontinent changes radically from a well-developed low during summer to a weak high in winter as the continental heat low and Intertropical Convergence Zone moves northwards, and consequently the pressure gradient along the western coast is seasonally variable. The curved anticyclonic flow associated with the South Atlantic high is guided by the coastline owing to the desert-like nature of the coastal plain acting as a thermal barrier to cross flow (Nelson & Hutchings, 1983) and by the orography of the continental escarpment. As a result the winds along the western coast of southern Africa are predominantly southerly and upwelling favourable. The Ekman transport computed from an extensive set of data by Parrish *et al.* (1983) for 1° latitude and longitude rectangles during January to February (summer) and July to August (winter) is illustrated in Figure 7a, and the effect of the seasonal migration of the pressure systems is clearly evident. There are four features apparent in Figure 7a which are worth highlighting. First, the Lüderitz area is the principal potential upwelling centre in the system, while at Cape Frio a secondary region of winds highly favourable for upwelling exists; secondly, there is a pronounced offshore divergence zone north of 28° S, which supports the earlier findings of Stander (1964), of a progressive anticyclonic rotation in the wind field with increasing distance offshore; thirdly, the wind stress maximum is situated in a band away from the coast, evident particularly during winter; the fourth feature is that the difference between the summer and winter nearshore Ekman transport north of Lüderitz is not large (*cf.* Stander's 1964 wind data), but is substantial in the south. What is less evident in Figure 7a but is apparent in the work of Berrit (1976), from the plots of wind speed cubed in the paper of Parrish *et al.* (1984), of wind stress and wind stress curl in Duing, Ostapoff & Merle (1980), and of nearshore wind stress in Boyd, Husby & Norton (in prep.)—see Figure 7b—is that the 15° S parallel is the approximate northern boundary of the highly upwelling favourable wind field. North of this latitude winds tend to be lighter and more onshore. It should be noted that Figure 7a was intended by Parrish *et al.* (1983) to facilitate the inter-regional comparison of four upwelling systems, and probably Figure 7b is more suited to a discussion of the longshore variation of wind stress near the coast in the Benguela region as the 1° rectangles are taken adjacent to the coast. Figure 7a tends to over-emphasize the winter wind stress near 27° S (Lüderitz) and under-emphasize the summer upwelling favourable winds in the southern part of the system. In Figure 7b the wind stress maxima near 33° to 34° S and 31° S during spring and summer are shown quite clearly.

The essential differences in the seasonal wind regime between the northern and southern parts of the Benguela region were illustrated by Hart & Currie (1960) from a consideration of four coastal sites. Their diagrammatic representation is shown here in Figure 8. In winter, with the northward shift of the pressure systems, the effect is much more pronounced in the south where the frequency of winds with westerly components—*i.e.* non-upwelling favourable—is significant. In this southern region of the Benguela, wind-induced upwelling is highly seasonal and reaches a maximum during spring and summer (Shannon, 1966; Andrews & Hutchings, 1980) and the upwelling season extends from September to March (Andrews & Hutchings, 1980). North of about 31° S the macroscale wind field exhibits relatively less seasonal variation. Upwelling is perennial here, but with a spring–summer maximum and autumn minimal as far north as 25° S and a late winter–spring maximum north of this latitude (Stander 1964; Schell, 1968). While the wind off northern and central Namibia shows relatively little seasonal variation, there are nevertheless slight maxima in the upwelling favourable wind during April to

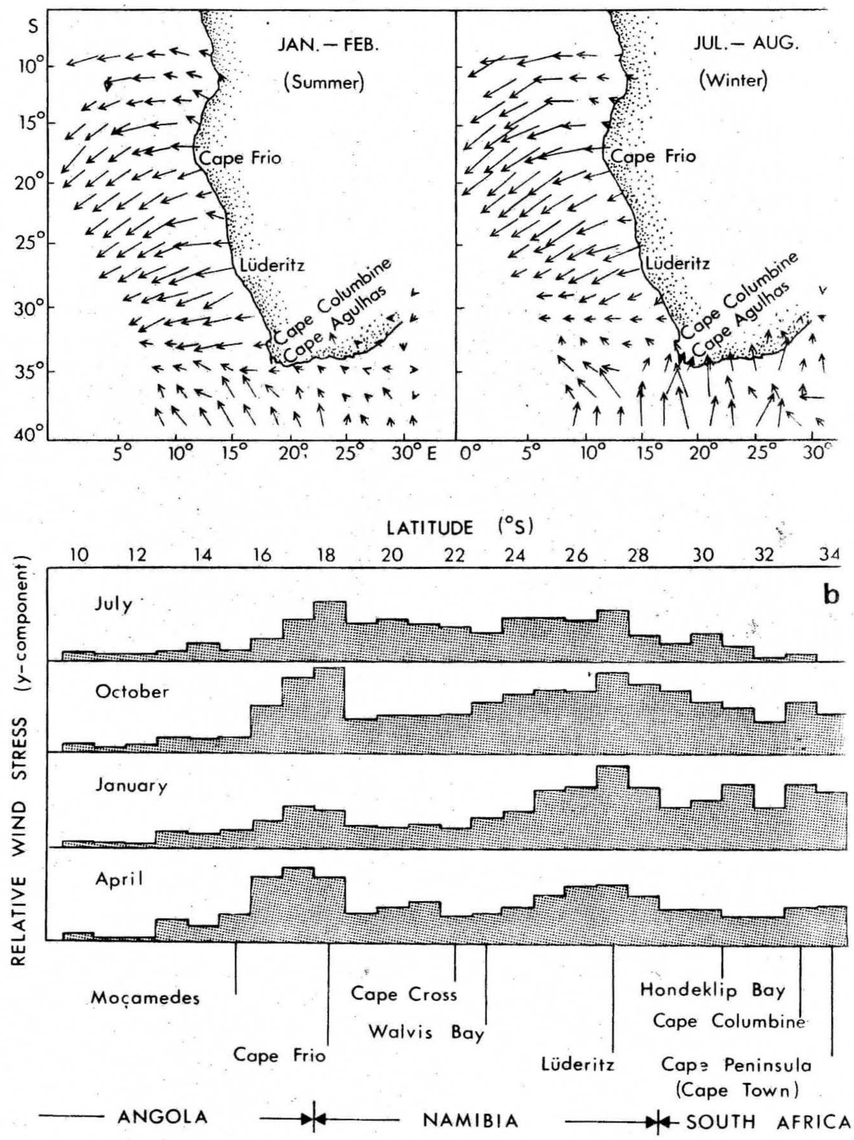

Fig. 7.—a, Ekman transport during summer and winter (after Parrish *et al.*, 1984). b, Y-component of wind stress per 1° rectangle adjacent to the coast (after Boyd *et al.*, in prep.).

May and October (Stander, 1958, 1963; Berrit, 1976; Boyd *et al.*, in prep.). Thus the Benguela region can be divided into two distinct regimes.

Land-sea breezes are common along the coast north of Cape Columbine (Jackson, 1947), and the diurnal modulation of the coastal wind has been described very adequately by Hart & Currie (1960) and Stander (1958, 1963

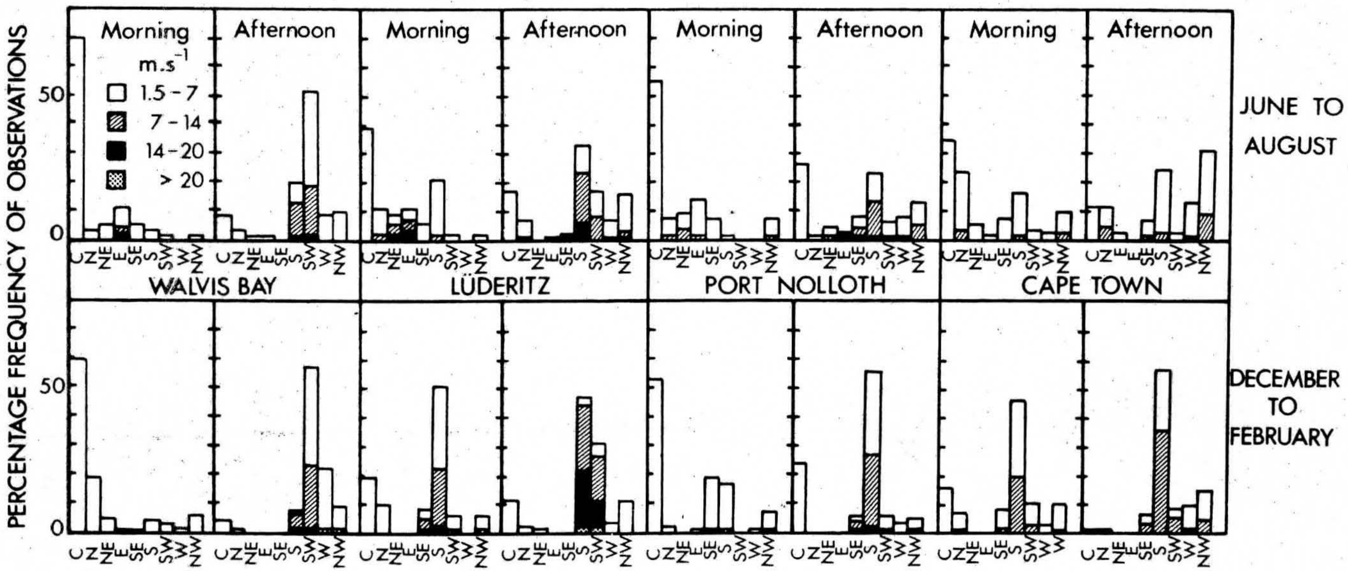

Fig. 8.—Percentage frequency and speeds of coastal winds in winter and summer (after Hart & Currie, 1960).

1964). The contrast between the morning and afternoon winds, in particular those at Walvis Bay, is illustrated in Figure 8. In this figure the marked intensification of the coastal winds during the day with a veering or backing of winds towards the south or southwest (depending on location and season) is evident. Distinct seasonal changes in the land-sea breezes were noted by Stander (1964). Hart & Currie (1960) have cited the view of S. P. Jackson ". . . that the sea-breeze probably has a fetch of 80–100 miles over the sea, that is from its divergence from the southeast trade". Obviously this diurnal pulsing of winds and the seasonal variation exhibited therein must be important for the coastal upwelling dynamics of much of the Benguela region, the northern part in particular.

In the southern Benguela an important modulation of upwelling with a longer period, *viz.* about a week, is provided by the wind relaxation or reversals associated with the passage of cyclones south of the continent during the upwelling season. Nelson & Hutchings (1983) summarized the situation very neatly as follows (see Fig. 9): "In the belt of westerly winds between 35° S and 45° S, low pressure cells form ahead of planetary waves in the subtropical jet stream. The associated cyclonic rotation of air produced as these cells advect eastwards, causes the wind field as far north as the Olifants River to be modulated with an intensity which increases southwards to Cape Point. In the summer months the effect is usually, but not necessarily, weak, manifesting

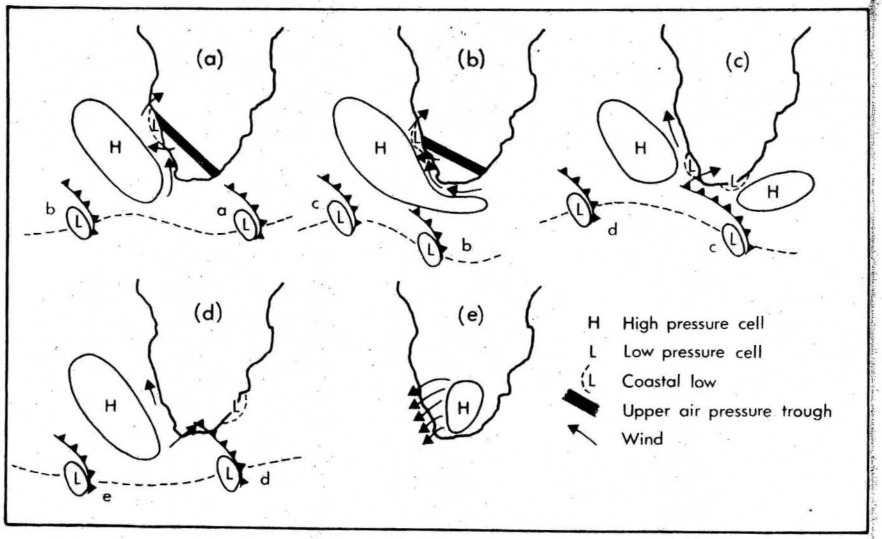

Fig. 9.—Cyclic weather pattern over the Benguela system typical of summer conditions (after Nelson & Hutchings, 1983). a, South Atlantic high established, coastal low at Lüderitz, southerly winds at Cape Town; b, South Atlantic high ridging, gale force winds at Cape Town, coastal low moves south; c, South Atlantic high weakens, northwest winds at Cape Town, following passage of coastal low; d, South Atlantic high strengthens, southerly winds along west coast; e, berg wind conditions.

itself as a periodic weakening of the South Atlantic High and slackening or abatement of south easterly winds along the coast. In the winter months, the effect may be, but is not necessarily, strong, bringing in its extreme form gale force north westerly to south westerly winds of some hours duration, in cycles of three to six days.... Associated with the approach of cyclonic systems, is the appearance of cells of low pressure which form near Luderitz (Taljaard, Schmidt and Van Loon, 1961) and travel round the subcontinent as trapped waves at a speed of about 750 km day^{-1}. . . . The migration of these cells southward along the coast seems to occur as a precursor to the approach of the cyclonic systems to the south (Nguyen Ngoc Anh and Gill, 1981) and is best observed in the summer months under conditions of weak modulation of the South Atlantic High pressure cell. The cyclonic rotation of air about these cells suppresses upwelling locally as the wave travels along the coast and the relaxation in the wind at the centre causes suitable conditions for the generation of inertial motions and possibly shelf waves."

Another large-scale feature of the meteorology of virtually the entire Benguela region is the occurrence of "berg" winds during autumn and winter (Jackson, 1947; de Wet, 1979). These occasional katabatic wind events are associated with the formation of a large high pressure system (a precursor to the coastal lows) over or just south of the southern or southeastern part of the subcontinent of several days duration. The anticyclonic circulation around the high results in a strong (up to 15 m·s^{-1} or more) easterly to northeasterly flow off the plateau of dry adiabatically heated air. Satellite imagery suggests that these winds are locally intensified by topographic features such as river valleys, and they may transport substantial quantities of sand and dust out to sea (Shannon & Anderson, 1982). This aeolian transport to a distance of 150 km offshore between 18° S and 30° S during a single berg wind event is shown in Figure 10 which emphasizes the directional coherency of the wind over a distance of nearly 1500 km. It has been suggested by Hart & Currie (1960) that these berg winds have little effect over the sea owing to their divergence above the marine boundary layer. On the macroscale, however, they appear to suppress upwelling in the southern region (Nelson & Hutchings, 1983). A. J. Boyd (pers. comm.) has indicated, however, that these winds can produce localized upwelling, but that this effect seems to be limited to within 10 km of the coast.

east wind.

Very little has been published on the interannual variations in the Benguela wind regime. An upwelling index based on the integrated longshore component of the wind at a single site (Cape Point) since 1961 has been cited in Hutchings, Nelson, Horstman & Tarr (1983), Nelson & Hutchings (1983), Crawford, Shelton & Hutchings (1984) and Hutchings, Holden & Mitchell-Innes (1984), and shows minima during 1966 and 1983 and a maximum in 1975, but no clear trends. The Cape Point site is, however, not characteristic of the whole Benguela region. Early work by Rawson (1908) has indicated a 19-year periodicity in the latitudinal position of the subtropical high pressure belt, which he tentatively suggested may be linked to the lunar period of 18·6 years, while Dyer & Tyson (1975) and Tyson (1981) have indicated a distinct periodicity in summer rainfall over the interior of South Africa. It seems probable that there may well be an interannual wind cycle over the Benguela regime, but long reliable sets of data are not readily available to verify this. It is also probable that the northern and southern Benguela regions will be affected

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Fig. 10.—Aerosol plumes of sand and dust due to a katabatic wind event: NIMBUS 7 CZCS, 670 nm band, 9 May 1979; (after Shannon & Anderson, 1982).

differently, as the recent studies on the southern Benguela warm event of 198: 1983 suggest (*e.g.* Shannon, 1983; Boyd & Agenbag, 1984b; Walker, Taunto Clark & Pugh, 1984).

As indicated earlier, much of the western coast is arid and has a low rainfa Rainfall over the sea in the Benguela region ranges from less than 10 cm·yr in the north to 50 cm·yr^{-1} in the south, while the respective annu evaporation ranges from less than 75 to 125 cm (Albrecht, 1960 as cited in Va Loon, 1972b). On average, the only significant inputs of fresh water into tl Benguela system are *via* the Orange and Kunene rivers (summer) in the nort and Olifants and Berg rivers (winter) in the south. The last three have mea annual run-offs of 7300×10^6, 708×10^6, and 528×10^6 m^3, respective (Department of Environment Affairs, South Africa). During periods of heav rainfall in the catchment areas the discharge is, however, manifest as a thin ler of low salinity water with mesoscale dimensions (Shannon, 1966). The impa of the episodic floods which occur in the Namib with a time scale of decade (*e.g.* during 1933–1934; Ward, Seely & Lancaster, 1983) on the oceanograph of the Benguela has not been well documented.

Fog is common over much of the region north of 32° S, in particular aroun Walvis Bay. Diagrams in van Loon (1972b) indicate the existence of a dee centre of dew point depression at 850 mb along the western coast with summer maximum (16 °C) centred around 27° S and in winter at 22° S (18 °C Total cloud cover generally increases offshore and from south to north, wit inshore values ranging from about 40 to 70% (Van Loon, 1972b; Parrish *et al* 1983). The summer minimum zone indicated by Parrish *et al.* (1983) lie between 25° S and 34° S while in winter it shifts north to between 22° S an 32° S in sympathy with seasonal movement of wind belts. The relatively clou free nature of the southern Benguela region makes it amenable to investiga tion using satellites (Shannon & Anderson, 1982).

WATER MASSES

The large scale hydrology of the South East Atlantic has been described by a number of authors. Clowes (1950) synthesized much of the early German (Meteor) and British (Discovery) work, while authors such as Fuglister (1960) Hart & Currie (1960), Darbyshire (1963), Stander (1964), Shannon (1966) Shannon & van Rijswijck (1969), Visser (1969a,b), Moroshkin, Bubnov & Bulatov (1970), Welsh & Visser (1970), and Henry (1975) have reported the results of several cruises undertaken in the region during the 1950s and 1960s.

There are several water masses present off the western coast of southern Africa, including *inter alia* tropical and subtropical surface waters, South Atlantic central, Antarctic Intermediate, deep and bottom water. The principal water masses are annotated on the suite of *T-S* plots for the Benguela area between latitudes 15° S and 35° S shown in Figure 11 and the general similarity between the *T-S* curves is evident. According to Clowes (1950), Stander (1964) and Shannon (1966) the water which upwells to the surface and to subsurface depths along the coast between Cape Frio (about 18° S) and Cape Point (about 34° S) is central water. This water mass which corresponds to the linear portion of the *T-S* curve connecting the approximate points 6 °C, 34·5‰ and 16 °C, 35·5‰ is found throughout the South East Atlantic, either

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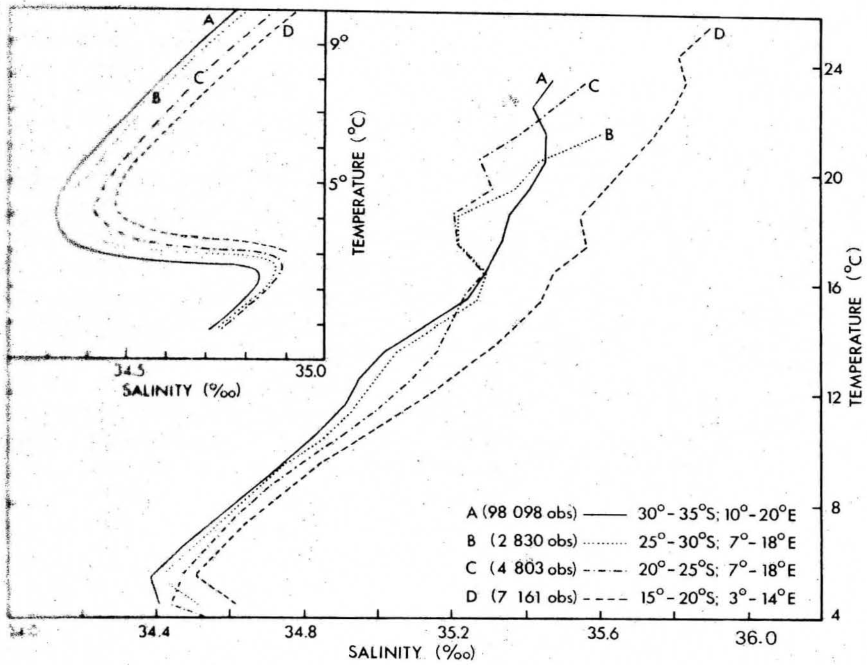

Fig. 11.—Mean salinity per 1 °C temperature interval between 15° S and 35° S, with the lower portions of characteristic *T-S* curves, drawn from scatter plots, inset.

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as a layer separating the surface and Antarctic intermediate layers in the oceanic region, or as the sole or main water mass present over the Benguela continental shelf. Central water is formed in the Subtropical Convergence region by the sinking and northward spreading of mixed subtropical and sub-antarctic water masses (Sverdrup, Johnson & Fleming, 1942; Orren, 1963, 1966). Central waters in both the South East Atlantic and South West Indian Oceans have very similar *T-S* characteristics (Orren, 1963; Shannon, 1966) and it is difficult to quantify the contribution of the latter to the Benguela system. Clowes (1950) and Shannon (1966) have suggested that some Indian Ocean central water enters the Benguela region at subsurface depths, although its contribution is probably small. Dietrich (1935) and Darbyshire (1963), however, have shown that the dynamic topography is favourable for the entrainment of this water around the Agulhas Bank. Clowes (1950), Shannon (1966), and Visser (1969a), by considering the properties of central water at the 26·70 to 26·75 sigma-*t* and 140 cl·t^{-1} isanosteric surfaces (approximately 10° to 11 °C; 34·9 to 35·0‰) deduced that in the oceanic region the flow of central water was predominantly northerly to northwesterly as far as latitude 20° S. North of 20° S the central water *T-S* curve is laterally displaced by about 0·1‰ (see Fig. 11) suggesting that in this region the central water is not of Benguela origin but has been advected southwards from lower latitudes. The lower upper portion of the *T-S* curves for the region between 20° and 30° S indicates the presence of a water mass characterized by a salinity minimum (18 °C, 35·2‰). The dissimilarity between the curves for the four 5° areas suggests that the origin of the salinity minimum may not be simply due to sun warming of upwelled water.

Overlying the central water west of the zone influenced by coastal upwelling are subtropical surface and subsurface waters with temperatures and salinities in the approximate ranges 15–23 °C and 35·4–36·0‰ (Clowes, 1950). Deacon (1937), Clowes (1950), Fuglister (1960), Shannon & van Rijswijck (1969), and Welsh & Visser (1970) have shown the existence of a subsurface current (salinity maximum) at depths between 50 and 200 m.

At its source Antarctic intermediate water has a characteristic temperature and salinity of 2·2 °C and 33·8‰, respectively (Sverdrup *et al.*, 1942), and the salinity minimum which marks the core gradually becomes less pronounced with distance from the source. In the South East Atlantic the minimum salinity is usually between 34·3 and 34·5‰ and the temperature between 4 and 5 °C (Clowes, 1950; Stander, 1964; Shannon, 1966). The meridional change in the minimum salinity is evident in Figure 11, and if the dilution curve generated by Welsh & Visser (1970) for the South East Atlantic is correct (it differs from that of Defant, 1961), then this implies that the percentage of true Antarctic intermediate water in the core decreases from about 50% in the southern Benguela region to slightly less than 40% north of 20° S. The salinity minimum lies at depths between 600 and 1000 m (Clowes, 1950; Fuglister, 1960; Shannon, 1966) but there is a tendency for it to be shallowest offshore at about latitude 24° S (refer to Fuglister, 1960) and immediately west of the shelf break in the southern Benguela region (*e.g.* Fig. 12). Isentropic analysis by Clowes (1950) and Shannon (1966) using sigma-*t* surface 27·25 suggested that there is a

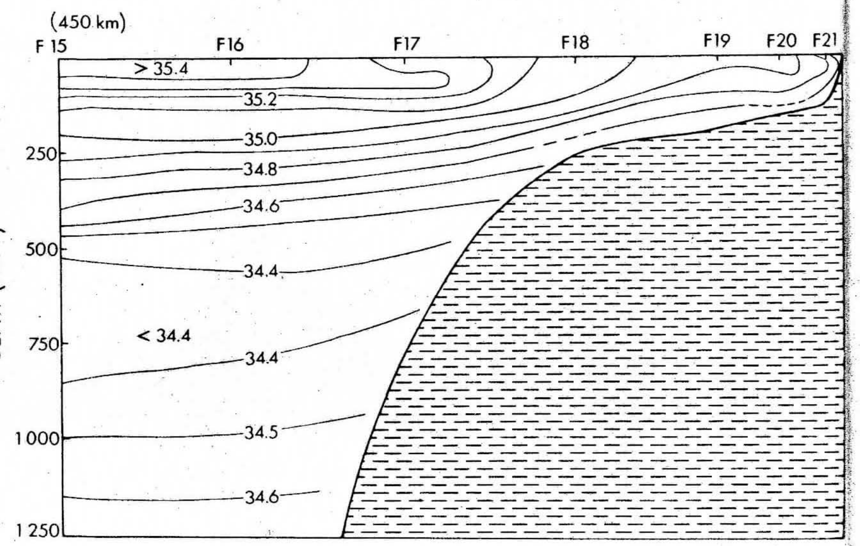

Fig. 12.—Salinity distribution off Roodewal Bay (30°21′ S), January 1959 (after Shannon, 1966).

northwesterly flow of Antarctic intermediate water south of latitude 25° S, a finding which was supported by Visser (1969a). Recent direct current measurements by G. Nelson (pers. comm.), however, show substantial poleward movement of this water mass in the Benguela region. Some Antarctic intermediate water may be advected from the Indian Ocean into the Atlantic around the Agulhas Bank (Clowes, 1950; Shannon, 1966). North of 25° S Visser's (1969a) data indicated a southward movement at the 80 cl·t^{-1} isanosteric surface (equivalent to sigma-t 27·28) within 500 km of the coast.

North Atlantic deep water is detected as a high salinity layer lying beneath the Antarctic intermediate water, and according to Clowes (1950) can be traced as far south as 56° S. Shannon & van Rijswijck (1969) noted the presence of the deep water core at depths ranging from about 2000 to 3000 m in the South East Atlantic, it being deepest at about latitude 32° S. Its temperature and salinity decreased from about 3·1 °C, 34·93°/oo at 16° S and 2·9 °C and 34·89°/oo at 24° S to 2·4 °C and 34·87°/oo at 32° S, suggesting slow southward movement. At 24° S in the Angola Basin the deep water is characterized by a slight salinity minimum, with the relatively uniform warm and saline North Atlantic bottom water lying below it. In the Cape Basin, however, the deep layer overlies the colder and less saline Antarctic bottom water which has a typical temperature and salinity of $<1·5$ °C and $<34·77$°/oo, respectively, and is present at depths deeper than 4000 m (Shannon & van Rijswijck, 1969). These authors considered that the Walvis Ridge effectively blocked the southward penetration of North Atlantic bottom water into the Cape Basin and northward flow of Antarctic bottom water into the Angola Basin, although they did detect some leakage at the break in the Walvis Ridge at 31° S; 2° E. The temperature profile along latitude 24° S (Fig. 13, from Fuglister, 1960) illustrates this blocking effect. Subsequent to Shannon & van Rijswijck's (1969) study the question of the penetration of Antarctic bottom water from the Cape into the Angola Basins was considered in more detail by Connary & Ewing (1974).

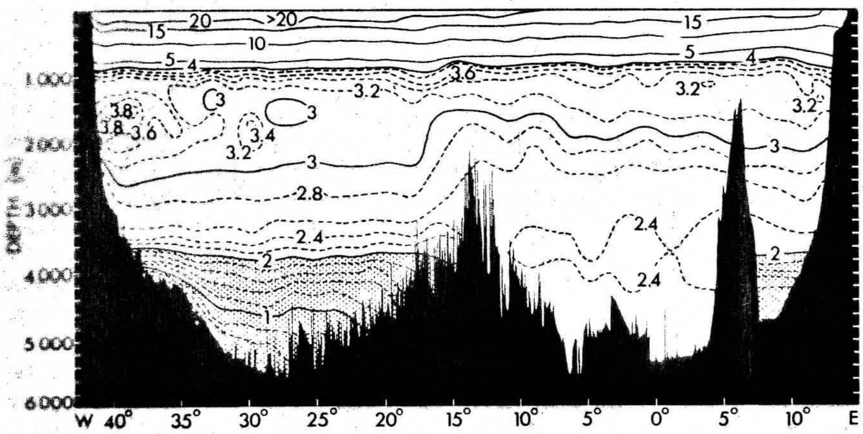

Fig. 13.—Temperature profile across the South Atlantic Ocean at 24° S, October 1958 (modified from Fuglister, 1960).

BOUNDARIES OF THE BENGUELA SYSTEM AND LARGE SCALE FRONTAL FEATURES

The "Benguela Current" was defined by Hart & Currie (1960) as the name applying "... to the region of cool upwelled coastal water along the South-west coast of Africa", *i.e.* water characterized by a pronounced negative surface temperature anomaly found mainly between 15° S and 34° S within 185 km of the coast, which forms the eastern periphery of the anticyclonic gyre in the South Atlantic. This definition was followed by *inter alia* Shannon (1966) while others have preferred to define the Benguela Current in terms of generally northward setting currents. Bang (1971) questioned both definitions; if considered on the basis of upwelled water the character of the "current" becomes absurdly discontinuous in both time and space; if defined in terms of surface flow, then the Benguela cannot be differentiated from the southeast Trade Wind Drift. Bang (1971) proposed that the term "Benguela Current" should be considered as "... that area east of the offshore divergence within which, as has been well established, oceanic processes are dominated by short-term atmospheric interactions. The divergence is thus, at least partly, a hydrodynamic discontinuity reflecting the geomorphological discontinuity of the continental slope and separating the weather-dominated Benguela system from what might be termed the climatic flywheel of the southeast Atlantic deep sea circulations." Several subsequent workers, *e.g.* Lutjeharms (1977), Nelson & Hutchings (1983) and Parrish *et al.* (1983) have tended to refer to the area as the Benguela current system, Benguela upwelling area, Benguela system or Benguela region. Throughout this review it is referred to as Benguela system or region. As processes taking place both seawards and shorewards of the shelf break and oceanic front are important for the understanding of the ecosystem as a whole, the western boundary of the system will, for the purpose of this review, be considered as being fairly open-ended.

While the question of definition of the seaward boundary of the Benguela system is somewhat academic, there generally exists over much of the area between Cape Point and Cape Frio a well-developed oceanic thermal front. South of Lüderitz (27° S) the front tends to be well developed and although spatially and temporally variable, approximately coincides with the run of the shelf break. The meandering nature of the oceanic front was first noted by Currie (1953) who suggested that this might be related to the existence of centres of upwelling and resulting mesoscale eddy systems as far north as Cape Frio. Little is known, however, about the oceanic frontal system off Namibia. Satellite-derived, sea-surface temperature and pigment (chlorophyll) maps indicate the existence of a frontal band which appears to be more diffuse than in the southern Benguela region. The width of the zone influenced by upwelling related processes off Namibia varies seasonally (see Fig. 14, which is from Parrish *et al.*, 1983; see also Stander, 1964; Boyd & Agenbag, 1984a). While the surface features, however, change in time and space, the oceanographic station spacing and sampling frequency has generally been inadequate to establish the existence and persistence of a baroclinic frontal zone near the shelf break off Namibia. In the southern Benguela region the oceanic frontal system is better documented, particularly between Cape Point and Cape Columbine thanks to the pioneering work of the late Dr Nils Bang

Fig. 15.—NOAA 1 enhanced infrared image of the Benguela, 15 June 1979, showing frontal features (from Van Foreest *et al.*, in press).

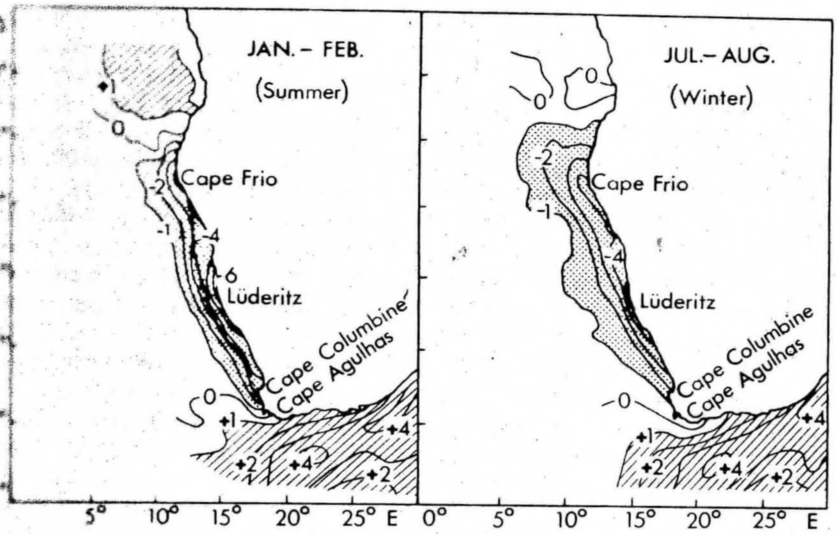

Fig. 14.—Sea surface temperature anomaly (from Parrish *et al.*, 1983).

(Bang, 1971, 1973, 1974; Bang & Andrews, 1974). Bang's mesoscale observations will be discussed later; at this stage it will suffice to record that Bang drew a clear distinction between the inshore frontal system and the strongly baroclinic offshore frontal zone (divergence) close to the shelf break. Another phenomenon which was observed by Bang (1971) in the vicinity of the shelf edge divergence between the Orange River and Hondeklip Bay was the existence of slicks, which he suggested might be related to internal wave activity. These features have been observed in some satellite images further south (Apel, Bryne, Proni & Charnell 1975; Nelson & Shannon, 1983), while they are commonly sighted off Namibia during research cruises (A. J. Boyd, pers. comm.).

Pronounced waves or meanders in the main oceanic thermal or pigment front have been illustrated by various authors, *inter alia* Currie (1953), Bang (1971), Lutjeharms (1981a), Shannon, Walters, Mostert & Anderson (1983), and Shannon, Hutchings, Bailey & Shelton (1984). In a recent paper Van Foreest, Shillington & Legeckis (in press) have suggested that the large-scale stationary features which have been observed at the thermal front from satellite imagery of the Benguela (see Fig. 15) may be related to the existence of a barotropic shelf wave with a 2·2-day period. These authors have pointed out, however, that while the wavelength of such a wave fits the observed structural features well, it does not explain their baroclinic and stationary nature.

The northern and southern boundaries of the Benguela system are reasonably well defined, although in the literature there is some disagreement as to what constitutes the boundaries. While Copenhagen (1953), Shannon, Nelson & Jury (1981), and Nelson & Hutchings (1983) have cited Cape Frio (18° S) as the effective northernmost boundary of the upwelling system, Hart & Currie (1960), Stander (1964), and Parrish *et al.* (1983) have shown that upwelling does take place north of this latitude. Hart & Currie (1960) provided evidence of coastal upwelling extending to 17° S and 15° S during March and October 1950, respectively, while Stander (1964) recorded upwelling off the Kunene River on three of his quarterly surveys during 1959, the most pronounced upwelling being during July 1959. The surface temperature anomaly maps from Parrish *et al.* (1983), shown in Figure 14, suggest that the upwelling influence extends as far north as about 15° S, and support the concept of a seasonal shift in the extent of the northernmost zone of upwelling. Parrish *et al.* (1983) have provided a sigma-*t* profile along a line situated approximately 200 km offshore around southern Africa (reproduced as Fig. 16). From this it is suggested that the northern boundary of the broader Benguela system lies at about 16° S, although from the work of Moroshkin *et al.* (1970) there is evidence for the northward penetration of components of the Benguela as far as 12° S to 13° S. What is important is that, while there are changes in the wind field and orientation of the coastline around 15° S, the northern boundary of the Benguela system is largely an oceanographic one. Although coastal upwelling does occur over a three- or four-month period (June–September) further north off equatorial west Africa (Berrit, 1976; Picaut, 1981), it is clear that this is not related to local winds, and is not a northward extension of the Benguela system. While the wind field north of 15° S is not highly favourable for upwelling, some uplift of warm saline waters close inshore from above the intense pycnocline is nevertheless possible.

Hart & Currie (1960) regarded the southernmost extent of the Benguela upwelling area as about 34° S and Andrews & Hutchings (1980) endorsed the view that the Cape Peninsula is the southernmost significant upwelling site. During the summer the upwelling zone can, however, extend as far south and east as Cape Agulhas (35° S: 20° E) which is regarded by Harris (1978) and Shannon *et al.* (1983) as a more appropriate boundary of the western coast system than Cape Point. At both of these capes the orientation of the coastline changes by about 45°. In addition, there is a marked change in the wind field

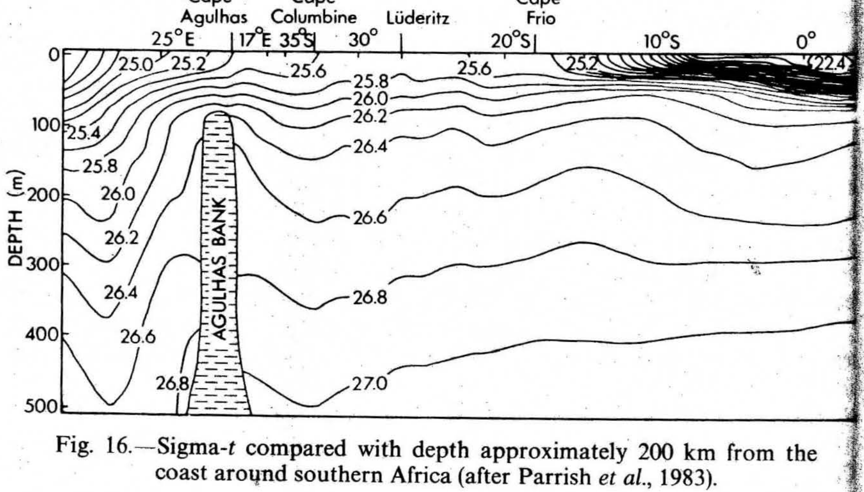

Fig. 16.—Sigma-*t* compared with depth approximately 200 km from the coast around southern Africa (after Parrish *et al.*, 1983).

south of about 35° S (see Fig. 7). Shannon *et al.* (1981) broadly interpreted the southern boundary of the Benguela system as the Agulhas retroflection area, which implied the inclusion of the Agulhas Bank region within the system (see also Fig. 16). That water of Agulhas origin contributes to the offshore part of the Benguela is well established (Clowes, 1950; Darbyshire, 1963; Shannon, 1966; Bang, 1973). As Bang (1973) stated, ". . . vestiges of Agulhas Bank or Agulhas Current water are almost always found off the Cape upwell cell". Reference to Figure 6 shows that organically rich sediments occur over the western Agulhas Bank, which is ecologically an integral part of the productive western coast regime. Thus, the southern boundary of the Benguela is produced by a combination of meteorological, oceanographical, and topographical factors.

The mesoscale processes associated with the northern and southern boundaries of the Benguela system and the oceanic front are discussed in more detail in subsequent sections.

MACROSCALE CIRCULATION

The following is a brief discussion of the large scale circulation in the upper layers of the South East Atlantic between latitudes 13° S and 38° S and the 0° and 20° E meridians. Readers are referred to the section (pp. 122–125) on water masses for information on the movement of the deeper water masses, while the literature pertaining to the highly variable and complex dynamics of the continental shelf region will be reviewed later.

SURFACE CURRENTS

Prior to the work undertaken during the Meteor and Discovery expeditions earlier this century, knowledge of the surface currents in the region was based on ships' drift measurements. Defant (1936) as cited in Hart & Currie (1960) analysed an extensive set of data of Dutch current observations and averaged them seasonally into one degree rectangles. His charts show the existence of a well-defined current in a band 200–300 km wide running in a north northwesterly direction, adjacent to the coast at 34° S and moving progressively offshore northwards. North of 20° S the streamlines bend westwards. Between 35° S and 20° S Defant (1936) indicated a one-sided divergence line, west of which the flow was predominantly westwards. It is significant that two satellite-tracked spar buoys (FGGE type) with drogues set just below the surface, released west of Cape Town during March 1977 (Harris & Shannon, 1979) and February 1979 (Nelson & Hutchings, 1983) followed similar general paths to that indicated by Defant's "Benguela current" streamlines. Nelson & Hutchings (1983) suggested that the current is topographically steered, accelerating in areas of steep topography and meandering over the planes. Harris & Shannon (1979) noted a predominantly westerly movement between 25° S and 18° S in accordance with the dynamic topography of the area (Stander, 1964; Shannon & van Rijswijck, 1969; Moroshkin, Bubnov & Bulatov, 1970). It is perhaps also significant that the drifter released in March 1977 crossed into the Angola Basin over the gap in the Walvis Ridge at about 22° S. In this respect Shannon & van Rijswijck noted that the Ridge appeared to influence local currents significantly. According to Harris & Shannon (1979) their drifter indicated a mean velocity of 17 $cm{\cdot}s^{-1}$, calculated over the shortest distance between the first and final fixes, which is similar to the mean velocity of drift cards passing through the region (Stander, Shannon & Campbell, 1969; Shannon, Stander & Campbell, 1973). Harris and Shannon (1979) recorded a 15° deviation of drifter trajectory to the left of the cumulative wind stress vectors, a finding supported by the inferred drift card trajectories in the region (see Shannon *et al.*, 1973). The implications of this are that, away from the coast and in the absence of strong gradient currents, the surface currents will be closely related to the prevailing wind. The anticyclonic curvature of both wind and surface current fields in the area tends to support this (compare Figs 7 and 17).

From drift card returns, Shannon, Stander & Campbell (1973) estimated a rotational period of the South Atlantic gyre of about 38 months. Although relatively few drift cards released in the Atlantic between 30° S and 50° S have been recovered along the west coast of southern Africa, those which have, suggest a mean speed in the West Wind Drift of about 15 $cm{\cdot}s^{-1}$. On the basis of drift card data Stander *et al.* (1969) speculated on the probable transport of *Jasus tristani* larvae from Tristan da Cunha to the Vema Seamount, and from their work the meandering nature of the currents between 30° S and 40° S and 10° W and 10° E can be inferred, and which has subsequently been confirmed by the trajectories of three satellite-tracked drifters through the area (Lutjeharms & Heydorn, 1981; Lutjeharms & Valentine, 1981). Their oscillating paths which abruptly change direction from eastwards to northwards on approaching 10° E were in agreement with the dynamic topography of Dietrich (1935).

The work of Stander (1964), Moroshkin *et al.* (1970) and Dias (1983a) has greatly contributed to the understanding of the complex flow patterns in the region north of 23° S. Moroshkin *et al.* (1970) showed a substantial westward flow between 23° S and 15° S at the surface with a large cyclonic gyre (r = 300 km) separating this "west (main) branch of the Benguela current" and the eastward flowing South Equatorial Countercurrent further north. These authors proposed the existence of three branches of the Benguela north of 20° S with a "Benguela divergence" separating the main (west) branch from the more easterly branches. They also showed the merging of the swift southerly flowing Angola Current inshore and the Benguela. It should be noted that the study of Moroshkin *et al.* (1970) related to autumn, a season which, as will be seen in the subsection on seasonal and interannual variability, is not typical of 'average' conditions in the Benguela region.

Different authors have measured or inferred surface or near-surface currents using various techniques, including drift cards, ships drift, satellite-tracked drifters, dynamic topography and isentropic analysis, and an attempt to synthesize the data has been made in Figure 17. It is based on the following: Rennell (1832); Dietrich (1935); Defant (1936); Clowes (1950); Hart & Currie (1960); Darbyshire (1963); Stander (1964); Shannon (1966); Duncan & Nel (1969); Shannon & van Rijswijck (1969); Stander *et al.* (1969); Visser (1969a); Moroshkin *et al.* (1970); Shannon *et al.* (1973); Harris & Shannon (1979); Lutjeharms & Heydorn (1981); Lutjeharms & Valentine (1981); Boyd & Agenbag (1984a); Nelson & Hutchings (1983); Dias (1983a); and Parrish *et al.* (1983).

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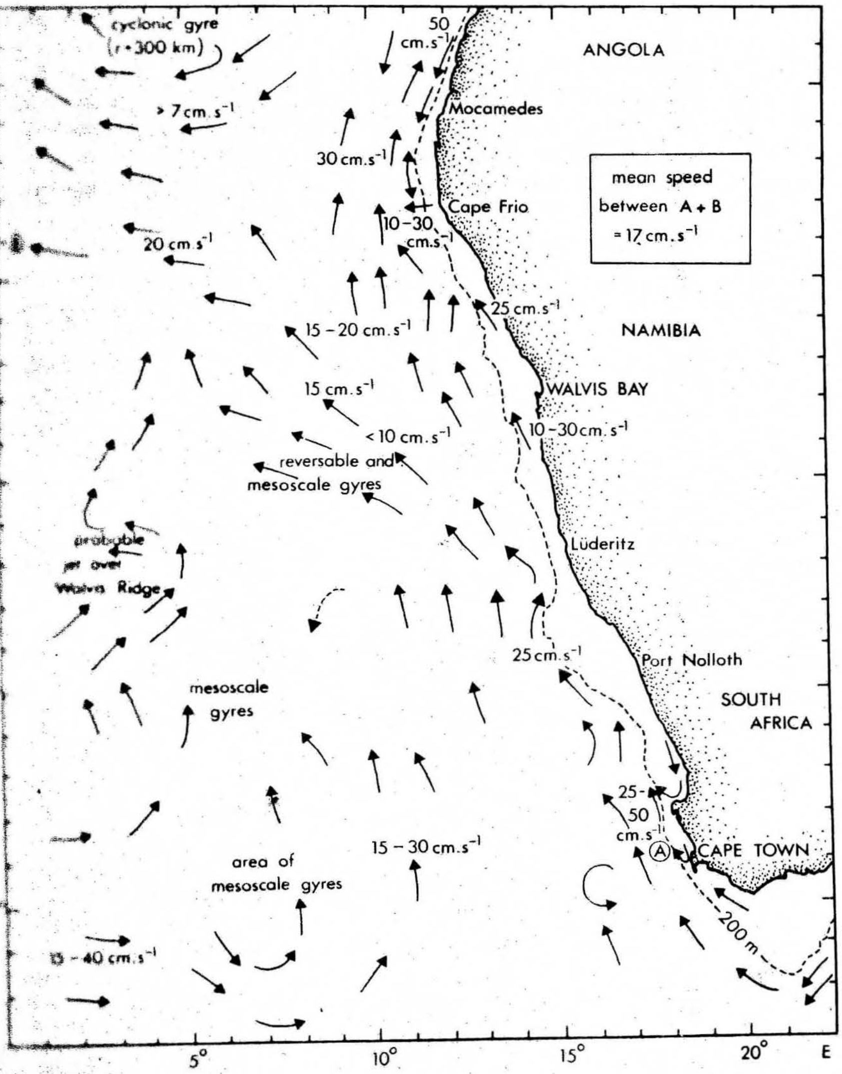

Fig. 17.—Surface currents—composite of work of several authors (refer to text).

CURRENTS BETWEEN 200 AND 300 M

The circulation of South Atlantic central water between 200 and 300 m is illustrated schematically in Figure 18. This figure is a composite based on the dynamic topography of Dietrich (1935), Stander (1964), Shannon & van Rijswijck (1969), and Moroshkin *et al.* (1970) and the isentropic analysis of Shannon (1966) and Visser (1969a). A note of caution must, however, be sounded. The papers of Shannon & van Rijswijck (1969) and Visser (1969a) relate to winter, while that of Moroshkin *et al.* (1970) was based on an autumn cruise, with the resultant bias of some of the features in Figure 18.

Excluding the 250 km wide coastal band, the direction of flow in the area south of 20° S is not significantly different from that at the surface (Fig. 17), *e.g.* the convergence of the West Wind Drift and the main equatorward flow between 10° E and 15° E; the topographic control exerted by the Walvis Ridge and the westward flow between 20° S and 25° S. North of 20° S the circulation

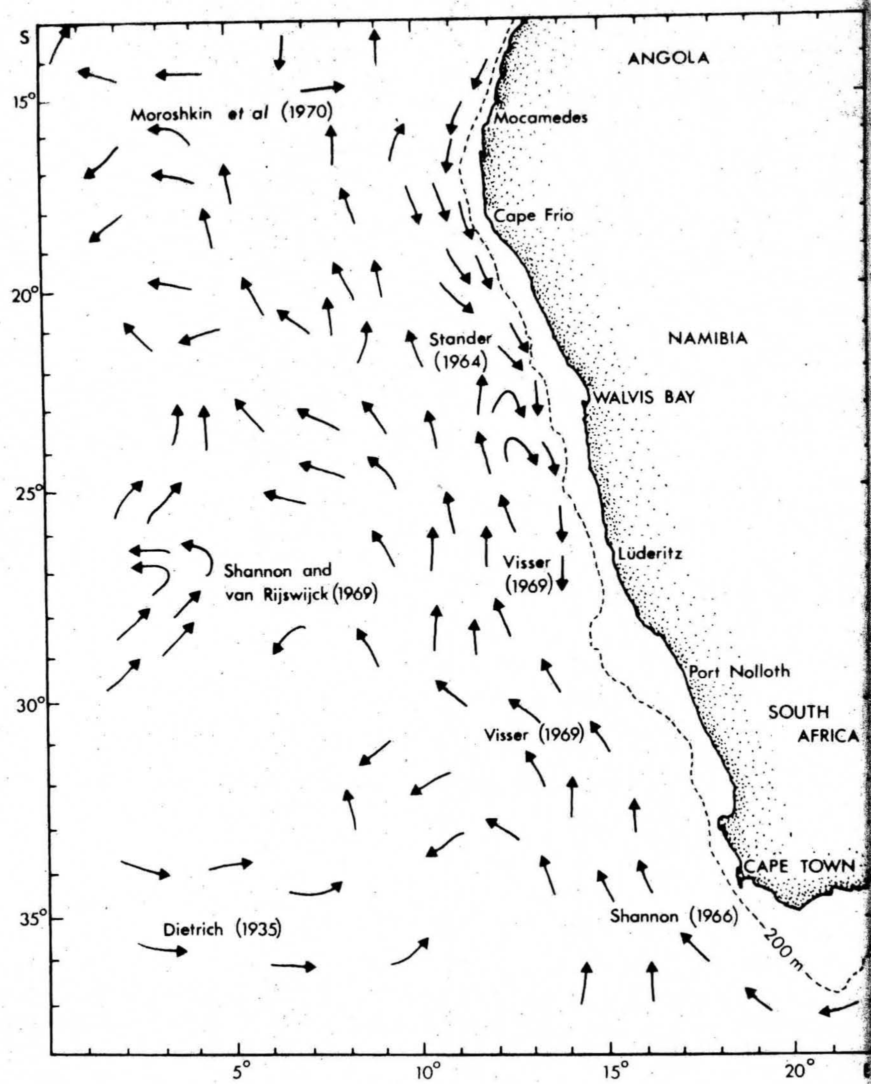

Fig. 18.—Probable movement of central water between 200 m and 300 m—composite based on work of Dietrich (1935), Stander (1964), Shannon (1966), Visser (1969a), Shannon & van Rijswijck (1969), and Moroshkin *et al.* (1970).

is evidently more complex. Moroshkin *et al.*'s (1970) data suggest a predominantly westward meandering movement west of 10° E and south of the anticyclonic gyre which is centred around 13° S: 4° E. Various authors, including *inter alia* Hart & Currie (1960), Stander (1964), Visser (1969a), De Decker (1970), Moroshkin *et al.* (1970), Bailey (1979), and Nelson & Hutchings (1983) have shown the existence of a poleward undercurrent flowing parallel to the coast west of the shelf break and penetrating as far south as Lüderitz (Visser, 1969a) and the Cape of Good Hope (De Decker 1970, Andrews & Hutchings, 1980). This undercurrent is at times characterized by its low dissolved oxygen content—typically $< 2\ ml \cdot l^{-1}$, and $< 1\ ml \cdot l^{-1}$ at the core—and was postulated by Hart & Currie (1960) as a deep compensation current, which according to Stander (1964) results in the poleward extension of the equatorial oxygen-poor zone. Nelson & Hutchings (1983) have suggested that a weak cyclonic gyral motion which exists in the region with the Walvis Ridge as its northern boundary would tend to trap the oxygen-deficient water. The available literature on the formation and advection of the oxygen-depleted layers is reviewed in Chapman & Shannon (1985).

VOLUME FLUXES

TABLE I

Volume fluxes in the Benguela system

Author	Latitude (° S)	Equatorward flux (Sv)	Lateral input flux (Sv)	Lateral output flux (Sv)	Comments
Defant & Wüst (1938)	25	14·9			Above 1000 db
Defant & Wüst (1938)	28·5	15·7			,, ,,
Sverdrup *et al.* (1942)	30	16			Upper water (above intermediate)
Bang & Andrews (1974)	34	7			Shelf-edge jet only
Bang (1976)	32–34·5	10*	0·5–0·7		*Shelf-edge jet only, at 32·5° S
[illegible] & Aagaard (1977)	20–25		1·7+	1·7	+1 Sv Ekman transport plus 0·7 Sv baroclinic transport above 200 m
Dias (1983b)	12	−1·2 to −3·7			Angola system above 400 m relative to 800 db, between coast and 9° E

The equatorward volume transport in the Benguela system appears to be o the order of 15 Sv (1 Sv = 10^6 $m^3 \cdot s^{-1}$), a figure which is comparable with th other major eastern boundary currents (Wooster & Reid, 1963). Bang & Andrews (1974) and Bang (1976) estimated fluxes of 7 Sv and 10 Sv respectively, for the equatorward shelf-edge jet in the southern Benguela which represents a substantial proportion of the gross transport in the South Atlantic. Nelson & Hutchings (1983) cited recent unpublished work which suggests that in the vicinity of the shelf-edge jet the flux may be locally mor intense.

The (few) published estimates of the meridional and lateral volume fluxes i the Benguela system are summarized in Table I.

SEASONAL CHANGES AND INTER-ANNUAL VARIABILITY

The seasonal distribution of temperature and salinity along the western coa has been described by several authors *inter alia*, Clowes (1954), Buys (195 1959), Stander (1958, 1963, 1964), Shannon (1966), Schell (1970), Wooste (1973), Christensen (1980), O'Toole (1980), Boyd & Agenbag (1984a), Parris *et al.* (1983), Stetsjuk (1983), and Strogalev (1983). Readers are also referred t the section on macroscale meteorology.

Christensen (1980) showed the mean monthly surface temperatures aroun southern Africa south of 25° S from data provided by commercial shippin during the period 1968–1978, while more recently Boyd & Agenbag (1984 have discussed the seasonal surface structure of the Benguela between 17° and 34° S and between 37 and 300 km offshore using a similar set of data b from a slightly longer period (1968–1980). Boyd & Agenbag have compare their seasonal (three months per season, *viz.* December–February, *etc.*) map with those generated from a more extensive data set, but with a coarser gri spacing, by Parrish *et al.* (1984) for the two two-month periods, January t February and July to August. The seasonal averages from Boyd & Agenba (1984a) are shown in Figure 19. These authors drew attention to the gener similarity between the winter and spring distribution with water cooler tha 16° C along the entire coast between 18° S and 34° S extending up to 300 k offshore. During summer and autumn (very similar—see Fig. 19) the area o cool water contracts meridionally and zonally. Longshore temperatu gradients are weak (1 °C or less per 1° of latitude) compared with the offsho gradients, the latter being strongest in summer and autumn in the area south o Walvis Bay (Boyd & Agenbag, 1984a). North of Walvis Bay the offsho gradients weaken slightly and the isotherms bend more towards the coast. Th large-scale seasonal surface temperatures broadly reflect changes in ins lation, upwelling, vertical mixing and horizontal advection. They do no however, show the intense mesoscale coastal upwelling events.

The changes in temperature and salinity in the upper 50 m based o monthly sampling by research vessels in areas off Namibia (21° S–24° S) an off the Cape (32° S–33° S) have been described by Buys (1957, 1959), Stand (1958, 1963), Stander & De Decker (1969) and Boyd & Agenbag (1984a,b). Th

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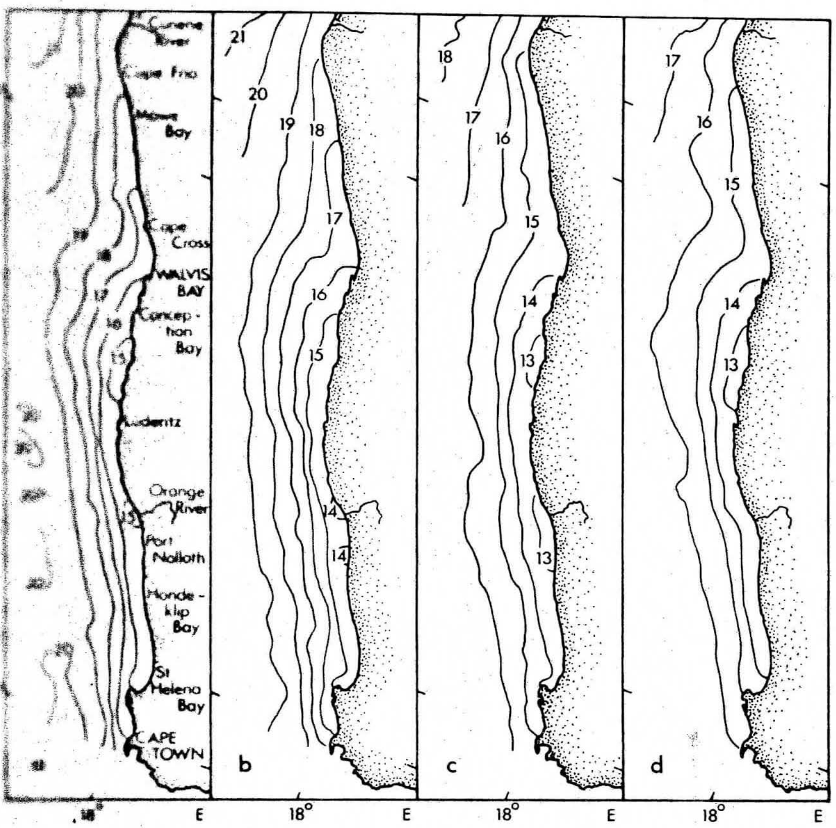


Fig. 19.—Average seasonal sea surface temperature (°C) (after Boyd & Agenbag, 1984a): a, summer; b, autumn; c, winter; d, spring.

The monthly variation in temperature in the two regions is shown in Figure 20, and indicates lowest values during August and close agreement between the 20 m and 0–50 m averaged temperatures. The seasonal signal appears to be more pronounced off Namibia than off the Cape where, excluding May, there is little change in the temperature at 20 m and 0–50 m during the year. Both Namibia and the Cape curves show peak temperatures during late summer to autumn with a two-month lag between the two regions (March and May, respectively). Off Namibia the salinity tends to follow the temperature curve but with a lag of about one to two months (see Stander & De Decker, 1969), whereas off the Cape the situation is more complex. (A note of caution: the area between 32° S and 33° S encompasses a strong baroclinic frontal zone and is highly variable. The variability off the Cape is evident from the time series of monthly measurements made by Andrews & Hutchings, 1980, near Cape Town—see Fig. 21.)

Two studies on the occurrence of thermoclines in the Benguela system have been made, one by Duncan (1964) and the other by du Plessis (1967) for the

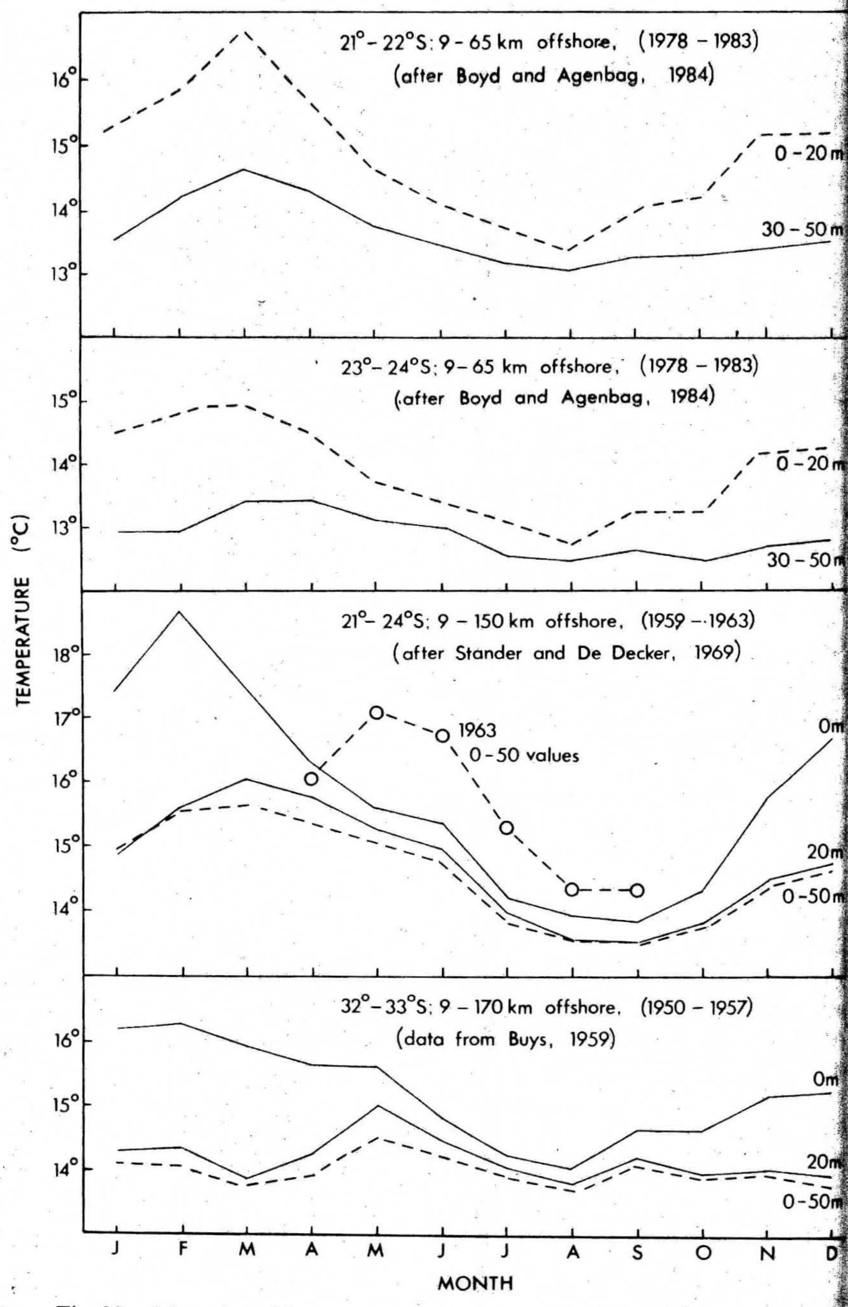


Fig. 20.—Mean monthly temperatures in the upper 50 m in two areas of the Benguela system.

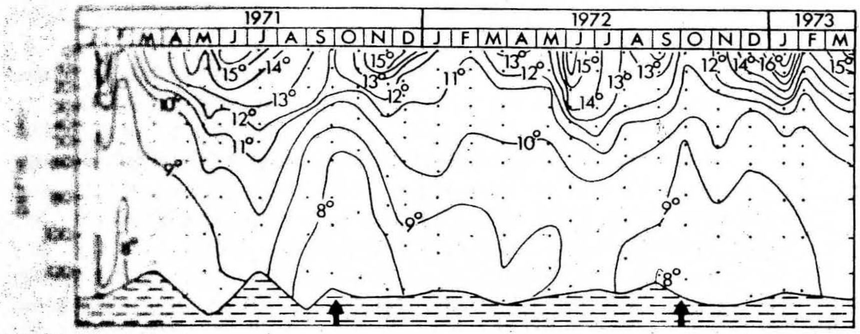


Fig. 21.—Monthly temperature distribution 25 km west of Cape Town showing intrusions of cold water near the bottom during September–October (after Andrews & Hutchings, 1980).

areas between 32° S and 33° S and 21° S and 24° S, respectively. The results are summarized in Table II. Off Namibia thermoclines tended to be shallower and except during summer were less frequent than in the south, and were associated on the average with slightly warmer less stratified water. Boyd & Agenbag (1984a) have suggested that the warming of the layer below 30 m off Namibia during late summer to early autumn is not primarily caused by

TABLE II

*Occurrence of thermoclines at two sites in the Benguela system: *estimated from diagrams*

Area, author, etc.	Season	Frequency occurrence (%)	Depth of top of thermocline (m)	Temperature at top of thermocline (°C)	Temperature drop across thermocline (°C)
21° S–24° S, < 150 km offshore 1959–1965, from du Plessis (1967)	Summer	86	10	17·5	3·1
	Autumn	52	22	16·8	2·2
	Winter	9	29	15·4	1·5
	Spring	24	8	15·5	2·0
32° S–33° S < 170 km offshore 1955–1961, from Duncan (1964)	Summer	87	17	15·6*	4·2*
	Autumn	87	22	15·0*	4·0*
	Winter	58	39	13·8*	3·0*
	Spring	60	33	14·2*	3·2*

downward heat transfer across the thermocline, and that the water which upwells there during this period is warmer and more saline.

Readers are referred to the sections dealing with mesoscale processes and to the work of Stander (1964), Schell (1970), and O'Toole (1980) off Namibia, and of Clowes (1954), Shannon (1966), De Decker (1970), and Andrews & Hutchings (1980) off the Cape for further information about seasonal changes in the vertical structure of the Benguela system.

Surface isotherms in the Atlantic Ocean during winter (*e.g.* Mazeika, 1968) show upwelling off the coasts of Angola and Gabon, *i.e.* north of the main Benguela region. In an analysis of data from the region Berrit (1976) concluded that north of 15° S the winds were not favourable for Ekman upwelling and that the strong upwelling signal evident in the temperature data had other causes. Upwelling occurs in the eastern equatorial region of the Atlantic and along the Gulf of Guinea coast during the austral winter, and Moore *et al.* (1978) attributed part of this upwelling to an internal Kelvin wave generated by increased easterly winds off northern Brazil. This wave travels along the equatorial wave guide and reflects at the African coast polewards and westwards in the form of coastal Kelvin and Rossby waves. Subsequent studies by Picaut (1981) have indicated that the wave propagates polewards between 1° S and 13° S with a phase velocity of about 0·7 $m \cdot s^{-1}$ and then propagates further south, but with some distortion. If Picaut's results are extrapolated then it would imply that the wave could reach Namibia during August and Cape Town (34° S) about a month or two (allowing for changes in stratification) later. Although there is no definite proof that this does happen, the data below 30 m in Boyd & Agenbag (1984a) and the appearance on the shelf near Cape Town of cold (< 8 °C) water during September to October (refer to Fig. 21—from Andrews & Hutchings, 1980) seem to support the wave concept. A comparable rate of progression of a "warm pulse" could possibly be inferred from a comparison of the diagrams presented by Berrit (1976) and Figure 20. It should also be noted that a warm pulse is evident off Abijan during April which is approximately the same time as the Benguela subsurface temperature maximum. Furthermore, Hirst & Hastenrath (1983) postulated a link between the (austral) summer relaxation of the westward wind stress in the equatorial western Atlantic, the subsequent warm pulse off Angola and the Angolan rainy season (March to April). It is tempting to speculate whether the upwelling system throughout the Benguela is not perhaps 'primed' by this coastal trapped Kelvin wave which then facilitates Ekman upwelling. The lack of a strong seasonal wind signal off Namibia and the oceanographic data from both the Cape and Namibia seem to support the idea of priming. If the Benguela upwelling is primed and then terminated by Kelvin waves, then this implies that a major causative mechanism of the inter-annual variability of the system would have its origin in the equatorial region of the Atlantic. Variations due to changes in the local wind field would be superimposed on this.

Very little has been published on the inter-annual variability in the Benguela system, and it is evident from what literature is available that long term records, where they exist, have yet to be adequately analysed. That perturbations occur on the time scale of years and decades seems probable from the biological record (Shannon, Crawford & Duffy, 1984), but the inherent short term and spatial variability of the system has presented programmes for monitoring. Buys (1957, 1959), Stander (1958, 1963), Stander & De Decker (1969), Strogalev (1983), and Boyd & Agenbag (1984b) have all analysed relatively short oceanographic sets of data in the northern and southern Benguela regions and have suggested certain trends. In the Cape, the work of Buys (1959) indicated that for the period 1950 to 1957 there was a cool interval during 1954 to 1955 and 1955 to 1956, and he suggested the possibility of a seven-year cycle, a view supported by Stander (1963) in his analysis of the 1954 to 1961 Namibian records, where a cool period was noted during 1957 to 1958. Strogalev's (1983) data indicated that 1972 to 1974 and 1976 were warmer than normal years off central Namibia. The recent study of the 1982 to 1983 southern Benguela warm event (reported in *S. Afr. J. Sci.*, Vol. 80, No. 2, February 1984) has highlighted the inter-annual variability in the system since 1950. Walker, Taunton-Clark & Pugh (1984), see Table III, reported on monthly sea surface temperature data for the southern Benguela for the post 1950 period and concluded that, while warm events in this region corresponding to Pacific events peaked during the summer, they had a distinctly different character to the El Niño, being related to variations in the wind field rather than to the advection of warm water into the system from the north. These authors showed that the early 1960s were 1 to 2 °C warmer than average years with 1963 (see last paragraph) being the warmest. 1967, 1971, and 1979 were cool years in the southern region and Walker *et al.* (1984) indicated a possible relationship between the position of the Subtropical Convergence and local sea temperatures (see also Gillooly & Walker, 1984). While the summer of 1982 to 1983 was abnormal in the southern Benguela with respect to temperature (Walker *et al.*, 1984; Duffy, Berruti, Randall & Cooper, 1984; Gillooly & Walker, 1984), wind (Nelson & Walker, 1984; Schulze, 1984; Hutchings, Holden & Mitchell-Innes, 1984), sea level (Brundrit, de Cuevas & Shipley, 1984), and biological characteristics (Shannon, Crawford & Duffy, 1984; Shannon, Brundrit *et al.*, 1984; Branch, 1984; Duffy *et al.*, 1984; Shannon & Chapman, 1983b; Hutchings *et al.*, 1984), off Namibia the work of Boyd & Agenbag (1984b) has shown no significant anomaly, although the preceding autumn and winter (1982) were characterized by cooler than usual conditions.

Probably the most satisfactory approach to the question of inter-annual variability is the examination of long-term tidal records. Recent studies on sea level, adjusted for atmospheric pressure and tides, by Brundrit, Shipley, de Cuevas & Brundrit (1983) and Brundrit *et al.* (1984) have shown coherency in the monthly records from nine sites along the coast (22°57′ S to 34°35′ S), with the inter-annual contribution, which has a very large spatial structure, revealing the same trend at each site. Their results suggested a decline in sea level since 1979 with 1982 showing a lower level in the inter-annual cycle. Anomalously high values were observed by Brundrit *et al.* (1984) at the southern sites during the 1982 to 1983 spring and summer. On the longer term, Brundrit *et al.* (1983) showed peaks during 1963 and 1968 to 1969 and troughs during 1965 to 1966 and 1971 (Table III).

Since the early 1950s, although there have been several warm and cool periods in the Benguela, only two events approximating to major El Niño type situations have occurred *viz.* in 1963 and 1984. During the 1963 event temperatures 2–4 °C and salinities 0·1–0·2‰ above normal were recorded in the upper 50 m off Namibia (Stander & De Decker, 1969). These authors noted that the southward intrusion of warm saline Angolan water was not accom-

TABLE III

*Comparison of upwelling indices for the southern Benguela region: *, 1960s were generally warm; ‡, units of relative wind displacement*

Year	Sea temperature: Shannon (1976) St Helena Bay area	Sea temperature: Walker *et al.* (1984) Table Bay & west coast*	Wind: Nelson & Hutchings (1983)	Wind: Nelson & Walker (1984)	Sea level: Brundrit *et al.* (1983, 1984)
1953–1954	Warm				
1955–1956	Cool				
1957	Warm	Warm			
1958	Cool				
1959					Low (2nd half of 1959)
1960		Warm			
1963	Warm	Very warm			High (mid-1963)
1964	Warm				
1966			Low (14)‡		Low (also 1965)
1967		Cool			
1968					High
1969			Low (18)‡		High
1970		Cool			
1971		Cool			Low (early 1971)
1972		Warm			
1973		Slightly warm			
1974		Warm	1974–1976		
1975			High (22)‡		
1976–1977		Summer warm			
1978		Cool			
1979		Cool			
1982–1983		Warm		Abnormally low	Low in 1982, [illegible] of 1982 to 1983 anomalous

caused by a decrease in upwelling favourable winds. A recent event of similar magnitude occurred during the late summer of 1984 (Boyd & Thomas, 1984). The effect of the 1963 event was felt subsequently in the Cape (Brundrit *et al.* 1985, and Table III). From data presented by Walter (1937), it is evident that events of similar or greater magnitude occurred during 1934, when monthly mean sea temperatures were 2–3 °C above the long-term average from March through July at Swakopmund. The anomaly was accompanied by a reversal or weakening of the usual northerly flow of the Benguela, and the flood waters from the Orange River were reported as moving southwards instead of northwards. Although not yet confirmed by the physical data, biological records (e.g. Shannon, Crawford & Duffy, 1984), suggest that Benguela El Niños may also have occurred in 1950 to 1951 and around the turn of the century. These perturbations do, however, seem to be less frequent in the Benguela system, than in the eastern Pacific.

THE NAMAQUA–LÜDERITZ UPWELLING AREA

The principal upwelling centre of the Benguela, as borne out by the work of several authors, *inter alia* Copenhagen (1953), Stander (1964), Boyd & Agenbag (1984a), Parrish *et al.* (1983), and Stetsjuk (1983) is in the vicinity of Lüderitz (27° S) –approximately equidistant from the northern and southern boundaries of the system. Defant (1936) noted that the zone of greatest negative surface temperature anomaly was situated between 23° S and 31° S an observation which, although based on relatively few data, has been substantiated by subsequent analyses by Wooster (1973), Parrish *et al.* (1984) and Stetsjuk (1983). Copenhagen (1953) identified on the basis of temperature and bathymetry three main centres of upwelling in the Benguela, *viz.* Lüderitz, Saldanha Bay, and Cape Point with secondary centres at Hondeklip Bay and Walvis Bay. Boyd & Cruickshank (1983) indicated maximum negative surface temperature anomalies 37 km offshore at 25° S and 29° S, which corresponded to the positions of two cool tongues noted by Hart & Currie (1960) on both their cruises, by Stander (1964), and by Bang (1971). These tongues which are apparently related to both the bathymetry and the wind field have as their bases Lüderitz and Hondeklip Bay and, together comprise a major environmental barrier in the Benguela, effectively dividing the system into two. Relatively few oceanographic stations have been occupied in the central Benguela, probably because most of the research has been focused on the important pelagic fishing areas which lie to the north and to the south of the 'cold' region. In the following paragraphs published work relating to mesoscale upwelling processes in the Namaqua (28° S–31° S) and Lüderitz (24° S–28° S) zones is discussed.

NAMAQUA ZONE

A cool wedge-shaped zone extending northwards and broadening from Hondeklip Bay to the Orange bight is evident from satellite thermal infrared imagery, while Stander (1964) and Shannon (1966) have shown the presence of cold water in the bight during most seasons. Nelson & Hutchings (1983) and Shannon-Clark (in press) considered that this tongue was largely determined

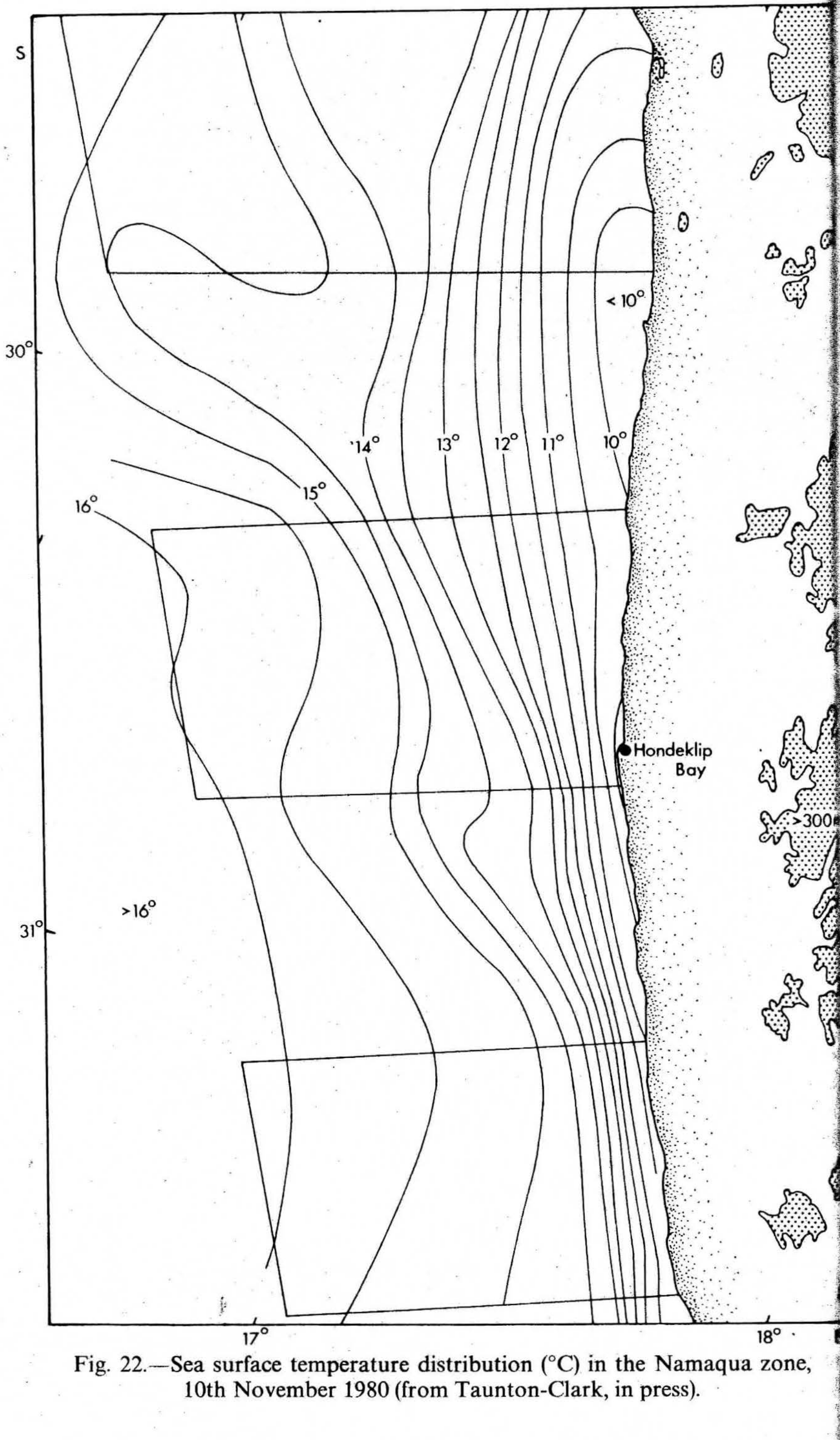
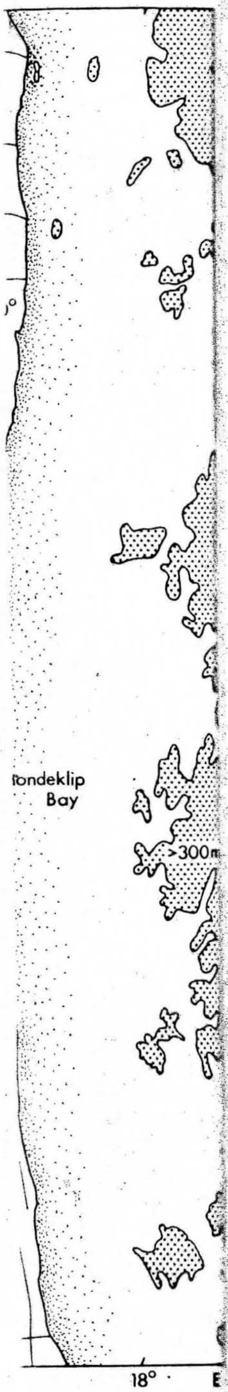


Fig. 22.—Sea surface temperature distribution (°C) in the Namaqua zone, 10th November 1980 (from Taunton-Clark, in press).



Namaqua zone,
ress).

by the bathymetry. Southwest of Hondeklip Bay the shelf is narrow and deep, and in this region cold water would be readily available close inshore. Further north the shelf broadens and the inner shelf is marked by a progressive shallowing towards the coast (Taunton-Clark, in press). Moreover the mean orientation of the coast changes slightly at Hondeklip Bay. The progressive northward shoaling of the deeper (8–12 °C) isotherms over the mid-shelf between 31° S and 29° S is evident from Bang's (1971) sequence of sections. Winds in the area are predominantly longshore with a diurnal sea-breeze modulation and a strong perturbation as coastal lows pass through and the South Atlantic high weakens (Nelson & Hutchings, 1983). Nelson, Kamstra & Walker (1983) have examined the mean annual winds and surface temperature on a half degree rectangle grid and have shown the existence of an area of maximum negative wind stress curl adjacent to the coast at Hondeklip Bay, with the coolest water to the north in the Orange bight. Taunton-Clark (in press) has recently reported on the results of a series of meteorological and surface temperature measurements made on aerial surveys of the Namaqua zone during the last quarter of 1980. Figure 22 illustrates what this author described as a typically developed upwelling tongue. Taunton-Clark (in press) recorded a wind speed maximum offshore, with a local maximum northwest of Hondeklip Bay coinciding with the plume base and the local thermal low pressure area. Lowest temperatures were recorded north of Hondeklip Bay and the author felt that the upwelling response in the area was markedly stronger than off the Cape Peninsula. Drift data in Clowes (1954) is consistent with the configuration of the Namaqua upwelling tongue.

Although Stander (1964) and Shannon (1966) found no evidence to support the classical cellular structure (Fig. 23) proposed by Hart & Currie (1960),

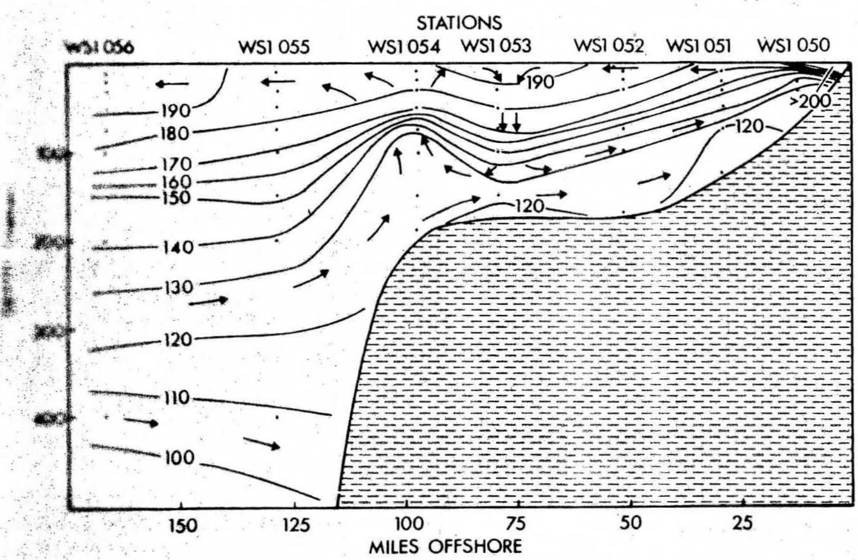


Fig. 23.—Distribution of specific volume anomaly, Orange River line, September 1950, showing cellular structure (from Hart & Currie, 1960).

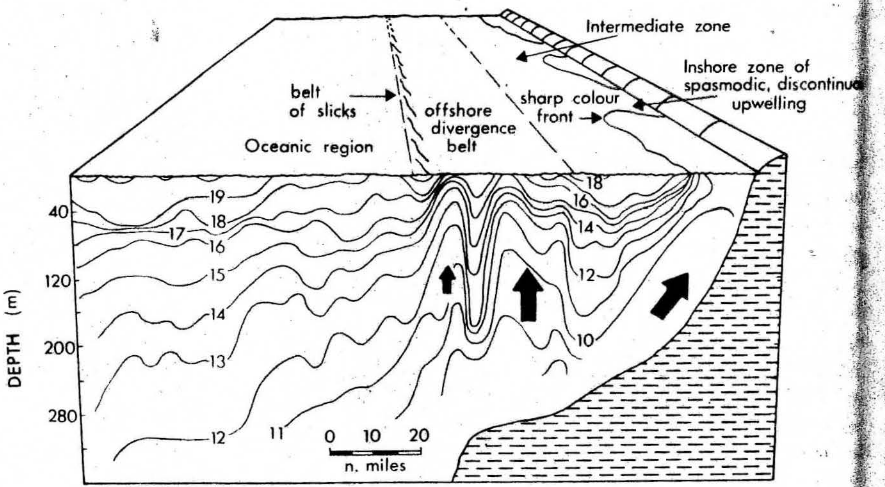


Fig. 24.—Main structural features of the Benguela region between 29° S and 32° S identified by Bang (1971).

probably on account of their wide station separations, Bang's (1971) vertical temperature sections of the area between 29° S and 31° S during February 1966 show marked wave-like features over the outer shelf and shelf break, which he suggested were indicative of surface divergence. This author's annotated schematic representation of the main structural features of the Benguela between 29° S and 32° S is shown in Figure 24. The presence of surface slicks noted by Bang (1971) over the shelf break, and which were probably related to internal waves, were subsequently observed in LANDSAT MSS imagery of the region between 30° S and 32° S by Apel *et al.* (1975). The slick spacing recorded by Bang (1971) *viz.* 1–2 km, was similar to that of Apel *et al.* (1975) offshore, *viz.* 1·6–1·8 km. The wave packets evident from the LANDSAT scene appeared to radiate out with a spacing of 20–40 km from a source near Child's Bank.

Shannon, Schlittenhardt & Mostert (1984) showed, from NIMBUS-7 CZCS imagery, the existence of an S-shaped band of elevated chlorophyll over the Orange Banks (approximately 150 km offshore between 28° S and 30° S) which they suggested might be associated with a semi-permanent shelf-break divergence zone. These authors, however, did not exclude the possibility of chlorophyll-rich water being advected offshore from the Hondeklip Bay upwelling centre. Whether shelf-edge upwelling, as suggested by Shannon (1966) actually takes place is debatable.

Although there have been no published measurements to date of the shelf-edge jet in the Hondeklip Bay–Orange River area, its existence seems probable from Bang's (1971) sections. Likewise, while there is no definite evidence of an undercurrent or deep compensation current in the area (Nelson & Hutchings, 1983) the possibility of these being present should not be excluded.

LÜDERITZ ZONE

Following the work of Copenhagen (1953), Currie (1953), and Hart & Currie (1960) which identified the region around Lüderitz as an important upwelling and the subsequent studies by Stander (1964), Bailey (1979) and Stetsjuk (1983) demarcated the spatial and seasonal extent of upwelling in the area. According to Stander (1964), "Low surface temperatures are most conspicuous between 26° and 28° S. It would undoubtedly appear that the area from the Orange River mouth to Lüderitz is a region where upwelling occurs repeatedly, probably more frequently and intensely than elsewhere along this coast." Stander's (1964) work which was based on nine quarterly surveys (1959 to 1961) showed persistent upwelling throughout the year with a slight maximum in spring and a minimum during autumn, which is in agreement with the trend evident in Bailey's (1979) analysis of the mean monthly upwelling favourable winds at Lüderitz from 1971 through 1977. Bailey described the wind field in some detail and showed that the southerly coastal winds were fairly consistent throughout the year with a tendency towards a maximum during the last quarter and a minimum (half to two-thirds of average speed) between May and July. (The wind speed maximum is situated 50 km or more offshore.) This is in general agreement with Stetsjuk's (1983) "cold advection" values which indicate minimum upwelling in the region 24° S to 28° S between July and August and maximum upwelling between October and December. Bailey (1979) also found good agreement between the monthly average sea-surface temperatures within 80 km of the coast between 24° S and 29° S and the winds at Lüderitz from 1969 through 1977, and he considered that the upwelling event scale for the Lüderitz zone was relatively long—more analogous to the Peru and North West African situations rather than to those off Oregon and in the southern Benguela. The sea temperature and wind speed records from Walter (1937) for Lüderitz during November 1927, which showed a good inverse correlation between the two variables, suggest, however, a fairly rapid response (Fig. 25). Somewhat curiously Bailey (1979) observed that during certain years in January to March warmer water appeared following stronger southerly winds (recorded at Lüderitz) in December to January. This response may be due to the large eddy, centred around 26° S to 27° S, which is characteristic of the Lüderitz zone upwelling, noted by Hart & Currie (1960), Stander (1964), Bang (1971), and Bailey (1979) and which is associated with a southward flow north of 25° S and an eastward intrusion of oceanic water between 27° S and 28° S, possibly as some form of compensation.

The temperature and salinity distribution at 0, 200, and 400 m off Namibia during January 1960 is shown in Figure 26 (from Stander 1964), and the spatial scale and impact of the Lüderitz upwelling site on the Benguela system is immediately evident as is the convergence zone between the tongue and the northern Namibian regime at 22° S to 24° S. Nelson & Hutchings (1983) have suggested that the coherency in the tongues of water between 200 and 400 m moving slightly offshore during the summer, points to the possibility of wind-induced upwelling being enhanced in the area by the bottom topography. Stander commented that cold water (7–10·5 °C) was consistently present in the region adjacent to the shelf at these depths. As in the Namaqua zone, the availability of this cold water near to the coast between 27° S and 28° S (the shelf is deep here with the shelf break at about 500 m—see Fig. 5, p. 114) coupled with the orientation of the coast and the shelf break appear to be important considerations for the upwelling dynamics.

Although unfortunately Stander (1964) did not show vertical profiles along the Lüderitz line, seasonal mean vertical sections from the same set of data are

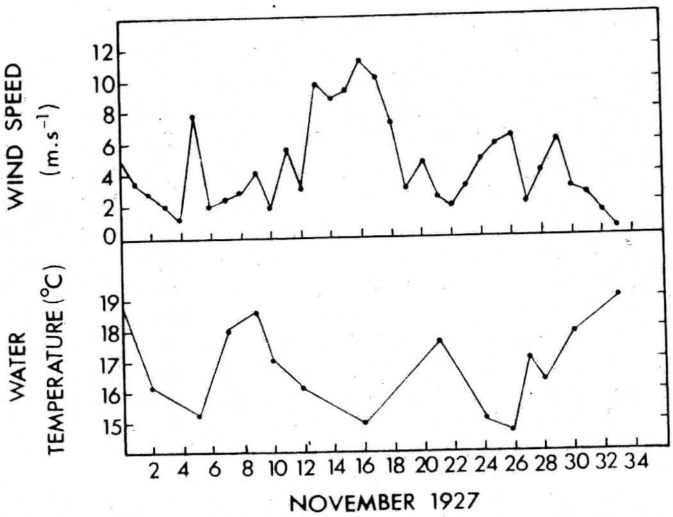


Fig. 25.—Variations in wind speed and water temperature at Lüderitz during November 1927 (after Walter, 1937).

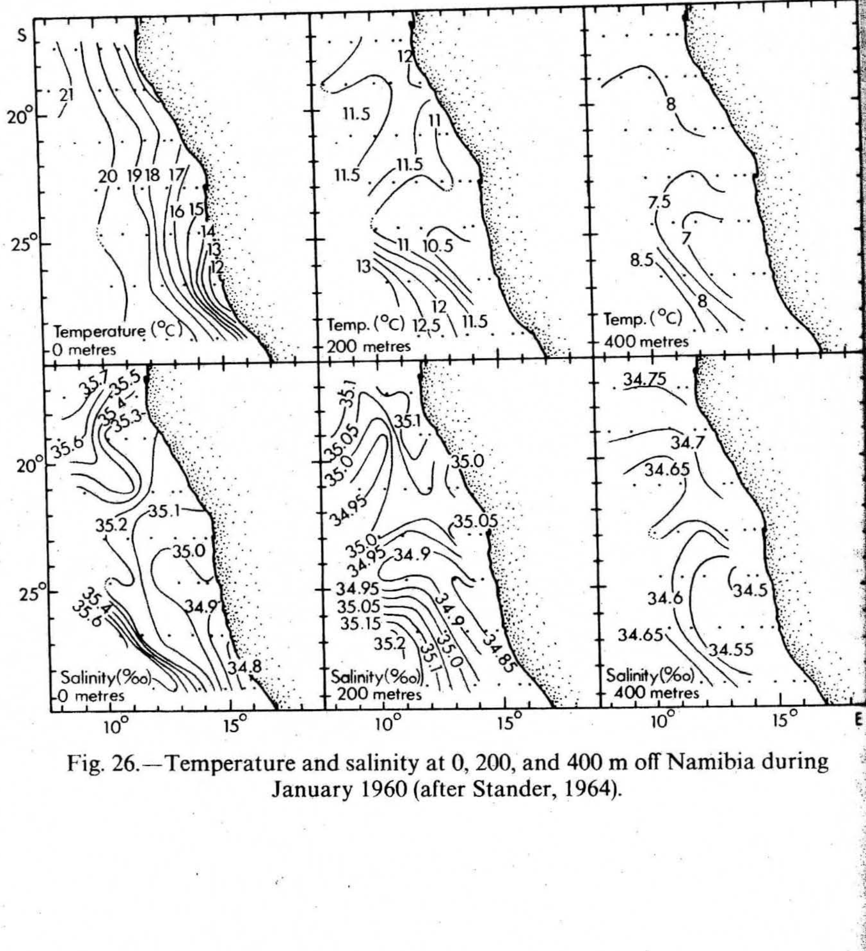


Fig. 26.—Temperature and salinity at 0, 200, and 400 m off Namibia during January 1960 (after Stander, 1964).

available in Schell (1970) and these confirm the pronounced uplift of the isolines over the shelf. Hart & Currie (1960) and Stander (1964) found that the upwelled water off southern Namibia originated from a depth of 200–300 m: Calvert & Price (1971b) have quoted a depth of 220 m for their Sylvia Hill line (25° S—their southernmost line and one on which maximum active upwelling was noted) during October 1968, while Bailey (1979) recorded depths of 300 and 330 m for lines off Chamais Bay (27°56′ S) and Marshall Rocks (26°21′ S), respectively, during an upwelling event in November 1976. It should be noted also that Latun (1962), using a two-layer steady state model, obtained good agreement with Hart & Currie's (1960) observations of upwelling on their Orange River line. He found strongest upwelling (3·5 to $4{\cdot}7 \times 10^{-4}$ $cm{\cdot}s^{-1}$) close to the shelf break and that water could not rise to the surface layer from levels below 325 m. Bailey (1979) has compared this maximum upwelling

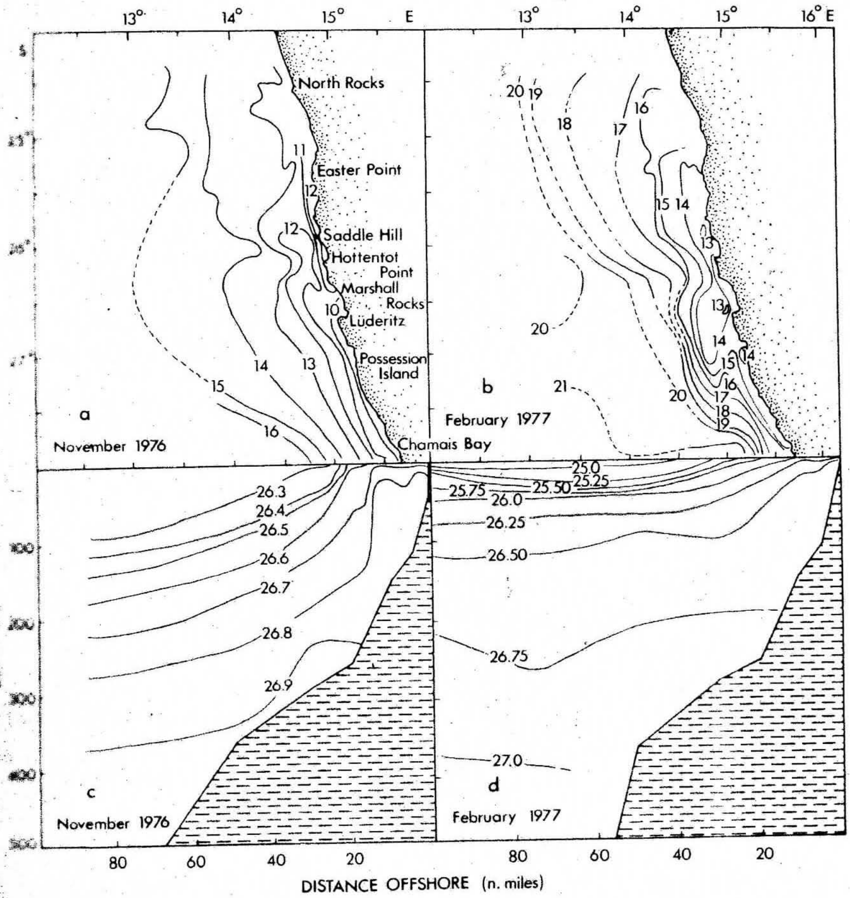


Fig. 27.—Surface temperature (°C) distribution in Lüderitz zone and sigma-*t* profile along the Marshall Rocks line (26°21′ S) during upwelling period (a,c) and quiescent period (b,d) (from Bailey, 1979).

condition with the quiescent case (February 1977) and his surface temperature charts of the area and vertical sigma-*t* profiles along the Marshall Rocks line for these two extreme cases are shown dramatically in Figure 27.

The only study of the response of the Lüderitz system to a mesoscale wind event is that by Bang (1971) who illustrated the effect of 3–6 days of gale force southeasterly wind. (His dramatic diagram is shown in Figure 28.) Bang noted that the mixing effects of the gale were smaller than anticipated, the thermocline depth having increased by 5–15 m. The 16 °C isotherm was displaced up to 80 km seawards over a few days, and temperature changes of 0·5–1 °C well below the thermocline at depths between 100 and 250 m were evident. Bang postulated the existence of a "pivot" or point of little change in the system at about 27°40′ S: 14°55′ E from his and Stander's (1964) data. Bang felt that, while upwelling occurs to the north and to the south, the vicinity of the pivot was one of comparative stability. The position of the pivot coincides with the central Benguela "environmental basin" proposed by Boyd & Cruickshank (1983) and may be the effective boundary between the Cape and Namibian subsystems.

Evidence for the existence of a deep compensation—poleward undercurrent in the Lüderitz zone was examined by Bailey (1979). Although no direct current measurements have been made in the area, Bailey deduced from the dynamic topography of the region (*e.g.* Fig. 29) together with isentropic analysis and an examination of the oxygen distribution that, during autumn, a poleward undercurrent was present 100–150 km offshore at a depth of 200–400 m, and over the shelf in the south between 50 and 100 m. Visser (1969a)

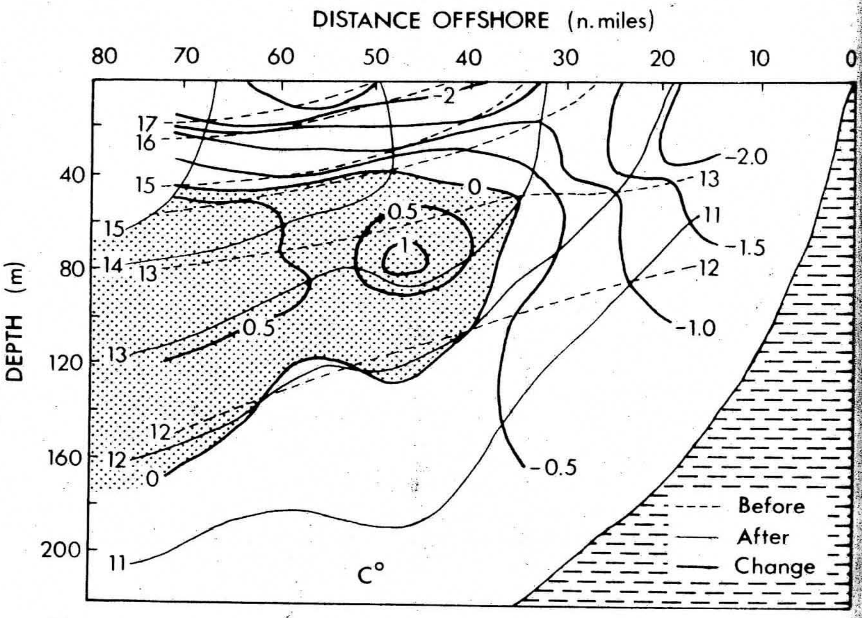


Fig. 28.—Temperature structure (°C) at 26° S (from Bang, 1971) before and after a southeasterly gale during February 1966.

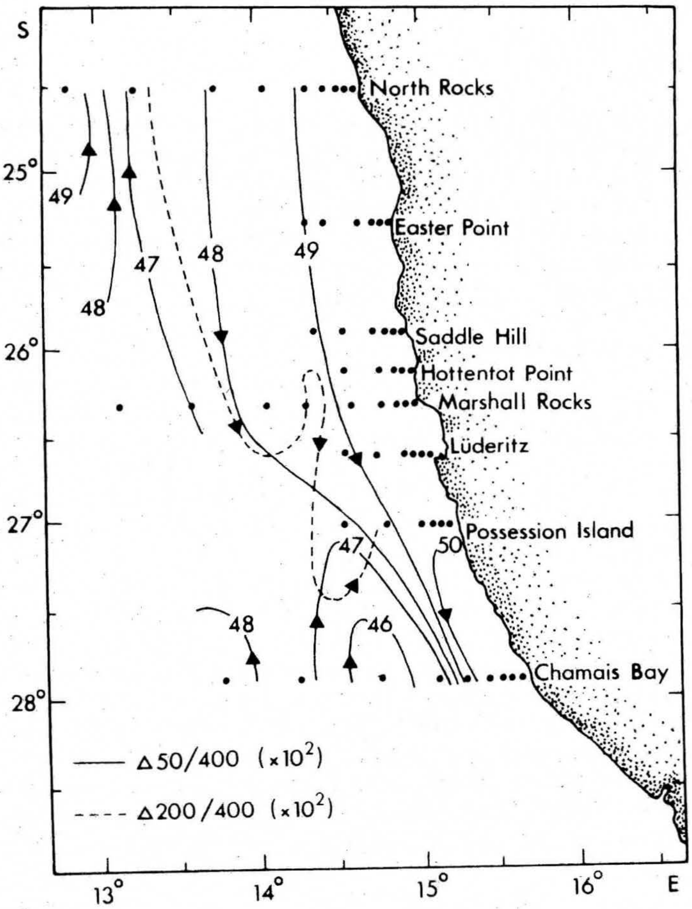


Fig. 29.—Water movement inferred from dynamic topography, May 1976 (from Bailey, 1979).

also showed an undercurrent 100–150 km west of Lüderitz at a depth of about [illegible] m during July 1959. The fact that it has not been detected during other seasons should not exclude it as a permanent feature, but may rather be symptomatic of the station spacings and measurement techniques.

MESOSCALE PROCESSES IN THE NORTHERN BENGUELA

The literature on the northern Benguela region is mainly of a descriptive nature (*e.g.* Currie, 1953; Hart & Currie, 1960; Stander, 1964; Calvert & Price, 1971b) and relatively little is known about the mesoscale dynamics off central and northern Namibia. Reference to the wind field and bathymetry and the work of Hart & Currie (1960), Stander (1964), O'Toole (1980), Boyd (1983a), Parrish *et al.* (1983) and Van Foreest, Shillington & Legeckis (in press) suggests that there are centres of upwelling around Conception Bay (24° S) and near 18° S–19° S and 20° S–21° S, *i.e.* south of Cape Frio and Palgrave Point, respectively. The region between Walvis Bay (23° S) and 21° S is a transitional area between the Lüderitz and northern Namibian zones and is generally characterized by a lower upwelling intensity (see Stander, 1964; Calvert & Price, 1971b; Boyd & Agenbag, 1984a).

O'Toole (1980) characterized three main surface water types in the region, *viz.* cool upwelled water (12–18 °C, *S* 34·9–35·2‰), warm, saline Angola water (17–22 °C, *S* 35·5–35·9‰) which periodically advances towards the southeast, mainly during summer and autumn, and water of oceanic or mixed origin (16–20 °C, *S* 35·2–35·5‰) which appears to advance from the west towards the coast between 19° S and 22° S during summer. Boyd (1983a) identified a fourth saline water type (*S* 35·3–35·5‰) having a temperature lower than 15 °C which is upwelled off central Namibia in autumn, the origin of which appears to be from the north.

The geostrophic circulation during nine cruises off Namibia has been discussed by Stander (1964) in some detail. Surface and subsurface flow patterns often differ substantially, *e.g.* Figure 30. The dynamic topography is not a good indication of surface currents (Boyd & Agenbag, 1984a) as the upper 20 m is primarily wind driven (Moroshkin, Bubnov & Bulatov, 1970; Boyd & Agenbag, 1984a; Hagen, 1984) and it is in this surface layer that most of the short-term variability in the system occurs (Stander, 1963). The upper 50 m is well mixed during winter and spring, the main upwelling season (Stander, 1964; O'Toole, 1980) but stratification is increased during the summer and autumn due to insolation, advection, and a partial relaxation of the wind.

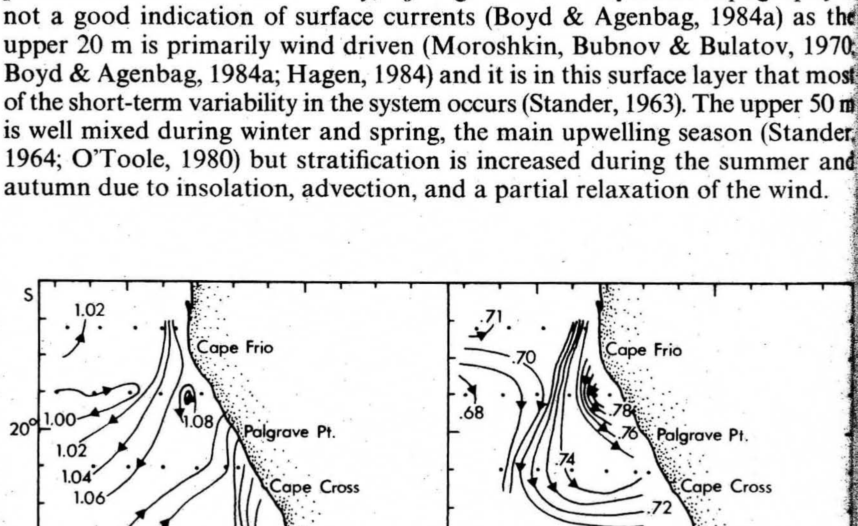


Fig. 30.—Dynamic topography of the 0 and 200 decibar surfaces relative to 1000 decibars, October 1959 (from Stander, 1964).

Except near the northern boundary of the Benguela or during wind reversals, the surface currents are predominantly longshore towards the northwest. Drogue studies by Boyd & Agenbag (1984a) at 10 m depth 46 km offshore (approximately over the 200 m isobath) at 21° S and 24° S showed relatively consistent motion to the northwest (10–30 cm·s^{-1}) in accordance with the prevailing winds. These authors suggested that this flow probably extended as far as about 18° S where the drogues indicated a pronounced offshore movement near to the Angola front. The work of Stander (1964), Yelizarov (1967), and Filippov & Kolesnikov (1971) revealed a fairly complex system of currents between 50 and 300 m. Yelizarov (1967) showed pronounced poleward flow west of the shelf break between 20° S and 28° S and he postulated the existence of a mesoscale anticyclonic gyre centred around 20°30′ S: 12°30′ E. Several of Stander's (1964) diagrams indicate two large eddies, one centred around Lüderitz and the other around 19° S to 20° S, with eastward flow between 21° S and 23° S (see also Fig. 30). The latter feature is difficult to explain—the area between Walvis Bay and Cape Cross is not a main upwelling site—unless a narrow jet is present over the shelf break. Nelson & Hutchings (1983) postulated the existence of a large cyclonic gyre between 18° S and 25° S, with poleward flow near the shelf break while Stander (1964) indicated a subsurface cyclonic gyre of smaller dimensions. Whether or not this gyre extends to the surface is, however, a matter of speculation, as is its permanence. That there is a perennial poleward jet or subsurface compensation current extending down to 300–400 m in the northern Benguela region, often but not always associated with low oxygen water, seems to be well established (Hart & Currie, 1960; Stander, 1964; Yelizarov, 1967; De Decker, 1970; Moroshkin *et al.*, 1970). The sediment texture and composition maps (Bremner, 1981; also Fig. 6, p. 115) show the existence of a narrow band of relatively coarse material deficient in organic carbon on the outer part of the inner shelf (*i.e.* west of the diatomaceous mud belt) extending between Cape Frio and Walvis Bay, and which may be associated with a subsurface jet. In this respect Hagen (1979) and Hagen *et al.* (1981) have shown evidence of an equatorward jet in the upper 100 m over or near the upper shelf break at 20° S. Its speed and persistence, however, have yet to be confirmed by direct measurement.

Data presented by Hart & Currie (1960), Stander (1964), Schell (1970), and Calvert & Price (1971b) indicated that the maximum depth affected by upwelling off central Namibia was generally little more than 200 m. Stander (1964) showed that the depth tended to increase through the upwelling season, occasionally reaching about 300 m near the end of the season. At 17° S (Kunene River) it seems that the upwelled water originates from a depth of 100 m or shallower, probably on account of the increased stratification in the vicinity of the northern boundary of the Benguela (see Fig. 16, p. 128). The only published estimate of the rate of upwelling off central Namibia is Stander's value of 1·9 m·day^{-1} (2·2 × 10^{-3} cm·s^{-1}) for periods of vigorous upwelling.

THE CENTRAL NAMIBIAN REGION

The central Namibian region (21° S to 24° S) can broadly be divided into two areas. Between Conception Bay (24° S) and Walvis Bay (23° S) the coastline runs approximately from north to south, while north of about 22°30′ S the orientation of the coast is about 330°. A double shelf break is characteristic of much of the region (see Fig. 5, p. 114) and is most marked west of Swakopmund where the upper break occurs at 140 m some 100 km offshore and the lower break at 400 m about 160 km offshore. The broad shallow Swakop shelf probably plays an important rôle in the dynamics of the region. Except near Conception Bay and at 21° S where the shelf is narrow, water from depths greater than 150–200 m is not readily available close inshore. The extremities of the central region seem to be upwelling centres (see figures in Calvert & Price, 1971b; O'Toole, 1980; Boyd, 1983a) although A. J. Boyd (pers. comm.) feels that they only appear as such in view of the reduced upwelling intensity in the area between 21°45′ S and about 23° S.

Diagrams in Stander (1962, 1964), O'Toole (1980), Parrish *et al.* (1983), Boyd (1983a), and Boyd & Agenbag (1984a) strongly suggest the existence of a semi-permanent convergence zone off central Namibia between the Lüderitz and northern Namibian systems. It is most pronounced during the summer between latitudes 22° S and 23° S (*e.g.* Fig. 25), and is at times more evident at subsurface depths (*e.g.* Fig. 30). Also the mean annual "cold advective values" of Stetsjuk (1983) showed a well-defined maximum (warm) at 22° S. A. J. Boyd (pers. comm.) prefers to consider the central region as a transitional zone with local bathymetry influencing conditions rather than an area whose dynamics are determined by convergence. That the configurations of the large upwelling tongue which emanates from Lüderitz to an extent determines the structure of the region nevertheless seems probable from Stander (1964), Bang (1971), and Parrish *et al.* (1983). Bang (1971) noted southward motion of the front just south of 24° S and inshore south of Walvis Bay following a gale, indicating cyclonic circulation and possibly nearshore compensation flow around the main Lüderitz upwelling tongue. Bang's data also seem to suggest a secondary centre of upwelling off Conception Bay.

The thermohaline characteristics of the central Namibian region are not well understood. The salinity fluctuations lag the temperature by about one to two months (Stander & De Decker, 1969) and this complicates the interpretation of upwelling (Boyd, 1983a). During autumn warmer, higher salinity water (15 °C; S 35·2 to 35·4$^{\circ}/_{\circ\circ}$) is upwelled in the region (Boyd, 1983a; Boyd & Agenbag, 1984a), and Boyd (1983a) suggested that this upwelled water had a more northerly or offshore origin than at other times of the year. Alternatively, instead of upwelling, the surface water which is removed through Ekman transport may be replaced to an extent by horizontal advection of warm saline water from the north during periods when stratification is at a maximum, but this would tend to contradict the observations of equatorward currents at 10 m by Boyd & Agenbag (1984a). (These authors suggested a poleward compensation just below the thermocline when stratification was maximum.) Nevertheless surface compensation may occur during periods of weak coastal winds, and whatever the origin of the water upwelled, the topography of the region between 22° S and 23° S would not tend to favour upwelling of water from deeper than 150 to 200 m.

Temperatures reach a maximum during March (see Fig. 20, p. 136; Strogalev, 1983), and during late summer water of oceanic or Angolan origin tends to move in from the north or west and then to retreat as the upwelling is intensified later through autumn when stratification decreases. This process may be what happened in the sequence shown in Figure 31. What should not

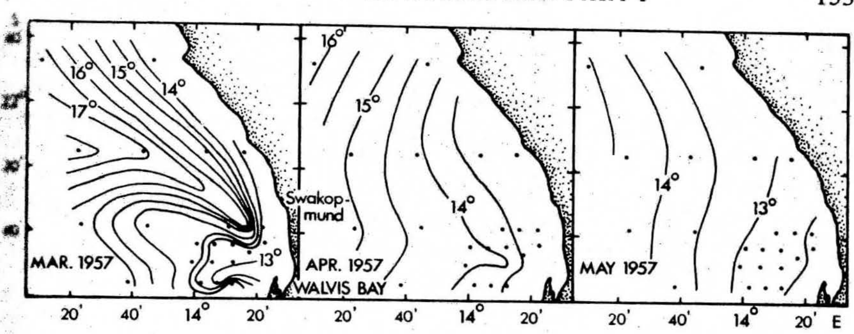


Fig. 31.—Temperature distribution (°C) at 20 m near Walvis Bay (after Stander, 1962).

be lost sight of is that the Tropic of Capricorn passes through Walvis Bay and during summer insolation is substantial (Copenhagen, 1953; du Plessis, 1967; Boyd & Agenbag, 1984a).

Although the station spacing in most published studies is inadequate to deduce the nature of the frontal structure over the shelf breaks, the work of Hart & Currie (1960), Stander (1962, 1964), Visser (1969a), Schell (1970), and Hagen *et al.* (1981) suggest the existence of a baroclinic jet. A subsurface salinity maximum at 50–100 m west of the upper shelf break is well defined during summer (Stander, 1962), but whether this is indicative of an undercurrent or merely an advanced stage of upwelling is not clear. Another feature which may be characteristic of summer is the presence of high salinity water (> 35·0‰) at the bottom on the mid-upper shelf off Walvis Bay (Stander, 1962) and also noted by Hart & Currie (1960) during September to October 1950.

In an analysis of winds at Pelican Point and temperatures at selected sites near Walvis Bay, Stander (1963) found a good inverse relationship between the southerly wind component and 0 and 20 m temperatures (Fig. 32) and fair agreement between downwelling favourable winds and the occurrence of

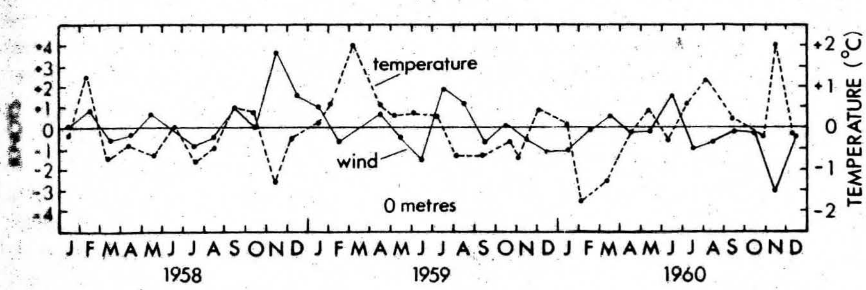


Fig. 32.—The relationship between sea water temperature anomaly (broken lines) at selected stations and the southerly wind component (solid lines) at Pelican Point, Walvis Bay (after Stander, 1963).

warm water. His view was that most of the short term variability was confined to the upper 20 m. What is somewhat surprising, considering the spatial scales, was the degree of coherency noted by Stander (1963) between the surface temperatures in the Walvis Bay (23° S) and St Helena Bay (32° S to 33° S) areas during the period 1954–1957. Stander's single 23-h anchor station indicated the existence of internal waves with a half-tidal period. Although Boyd (1983a) found little change in the vertical structure over a two-week period near 22° S, data from Schultz, Schemainda & Nehring (1979) showed distinct short-term periodicity, possibly related to internal tides with an amplitude of about 10 m (see Fig. 17 in Chapman & Shannon, 1985).

The only direct measurements of currents in the areas are those of Boyd (1983b), Boyd & Agenbag (1984a), Boyd, Potgieter & Buys (1983), and Hagen (1984). The drogue tracking study of Boyd (1983b) over the shelf around 22° S suggested that diurnal land–sea breezes control the currents in the upper 5 m, with a response time of a few hours, but that deeper currents were not directly influenced by the wind. The average surface current (10–15 $cm{\cdot}s^{-1}$) was 1·7% of the wind speed. Boyd noted marked shear between the surface currents and those at 20 and 30 m. At the latter depth Boyd observed a meandering poleward undercurrent with an average velocity of 3·5 $cm{\cdot}s^{-1}$. Boyd *et al.* (1983) identified three main time scales from a current meter mooring near Walvis Bay between October 1981 and March 1982, *viz.* 2–4 days; tidal or diurnal; those with periods of less than half a day. These authors obtained a fair correlation between the mean daily southerly wind component and the onshore subsurface currents, the latter evidently compensating for the cyclonic circulation and net surface flow out of the Bay noted by Pieterse & van der Post (1967).

Hagen (1979) suggested that the rhythms with periods of several days in the current field over the shelf were due mainly to the dynamics of these waves, while Hagen (1981) concluded that the energy-rich barotropic shelf waves occasionally impressed their space–time structures upon that of the local baroclinic mass field. His model, which ignored stratification, indicated that, for the shelf configuration at 20°30′ S there was a northward propagation of barotropic mesoscale eddies of 9 $km{\cdot}day^{-1}$, while the second and third modes of the solution yielded wavelengths and periods of 600 km and 350 km and 5 days and 7·2 days, respectively, *i.e.* similar longshore spacings to the features observed by Van Foreest *et al.* (in press). In a follow-up investigation Hagen (1984) compared the space–time patterns arising from the cross-shore modal structure of a free barotropic continental shelf wave with those of the baroclinic structures from the relative pressure fields. His study was supported by data from four current meter moorings and a cross-shelf transect between 20° S and 21° S comprising 15 stations 10 km apart which was repeated 15 times at 36-h intervals during the autumn of 1979. Hagen concluded that the local temporal cross-shelf variations in the structure of the observed mass field could be explained in terms of the linear theory of continental shelf waves. Hagen (1984), furthermore, found that within 100 km of the coast the longshore current and the appropriate mass fields responded hydrostatically with less than a day lag to the variations in the long-shore component of the local wind. These events had a mean period of 5·6 days. Seawards of the 100-km wide coastal zone Hagen indicated the possible existence of an internal Rossby wave with a period of 13 days.

INTERACTION BETWEEN THE BENGUELA AND ANGOLA SYSTEMS

The region between 15° S and 20° S is extremely complex and not well understood (see Currie, 1953; Stander, 1964; Filippov & Kolesnikov, 1971; Moroshkin *et al.*, 1970; O'Toole, 1980; Parrish *et al.*, 1983; Nelson & Hutchings, 1983; Boyd & Thomas, 1984). Usually evident from the temperature and salinity distribution is a well-defined convergence which demarcates the approximate boundary between the surface waters of the Angola and Benguela systems. This front which is positioned approximately over the Walvis Ridge abutment extends from close to the coast near 17° S offshore in a southwesterly direction and appears to migrate seasonally over about 3° of latitude (18° S in March, 15° S during winter). The surface temperature and salinity changes associated with it are about 4 °C and $0{\cdot}4\,^{\circ}/_{\circ\circ}$, respectively—the distinct differences in the thermohaline characteristics of Angola water and that of Benguela origin are illustrated by the upper portions of the mean *T-S* curves (see Fig. 12, p. 124).

The interpretation of the mesoscale oceanographic processes in the region are complicated by the fact that the majority of cruises tended to terminate near the Angola–Namibia geographic boundary, and by the generally inadequate spatial and temporal scales of measurement. The timing of the cruises, many of which have taken place during autumn when the southward flowing Angola Current (Fig. 33) is near a maximum and upwelling off

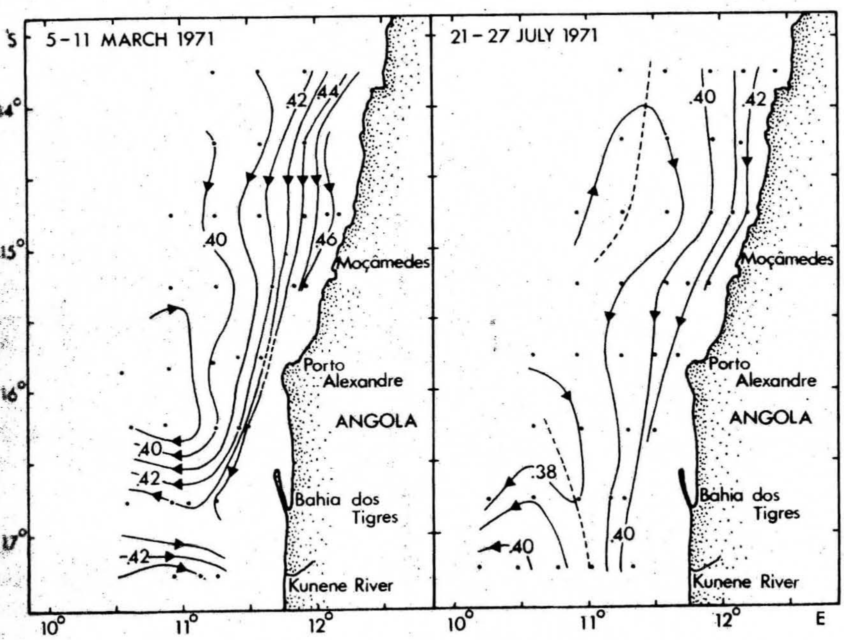


Fig. 33.—Dynamic topography of the 100 relative to 400 decibar surface during March and July 1971 off Angola (after Dias, 1983b).

Namibia near a minimum may also have led to some bias in the published interpretations.

Except for the work of Stander (1964) little appears to be known about the seasonal changes at 17° S. His figures indicate, that at this latitude, stratification reaches a maximum during summer west of the 100-km wide coastal zone, and a minimum during winter. Stander (1964) showed clear evidence of upwelling off the Kunene River at times. His diagrams suggest that during summer, when it occurs, it does so very close inshore and the upwelled water originates from depths shallower than 50 m. During winter the depth affected by upwelling may extend to 100 m. From the wind stress (Wooster, 1973; Berrit, 1976; Duing, Ostapoff & Merle, 1980; Parrish *et al.*, 1983) and topography it might be expected that the area north of Cape Frio would be a key area for upwelling throughout the year (with slight maxima during May and October), but this does not seem to be the case. Either the stratification is such, particularly during summer, that relatively warm saline water is recycled at very shallow depths or otherwise the surface water transported offshore (*cf.* drogue studies

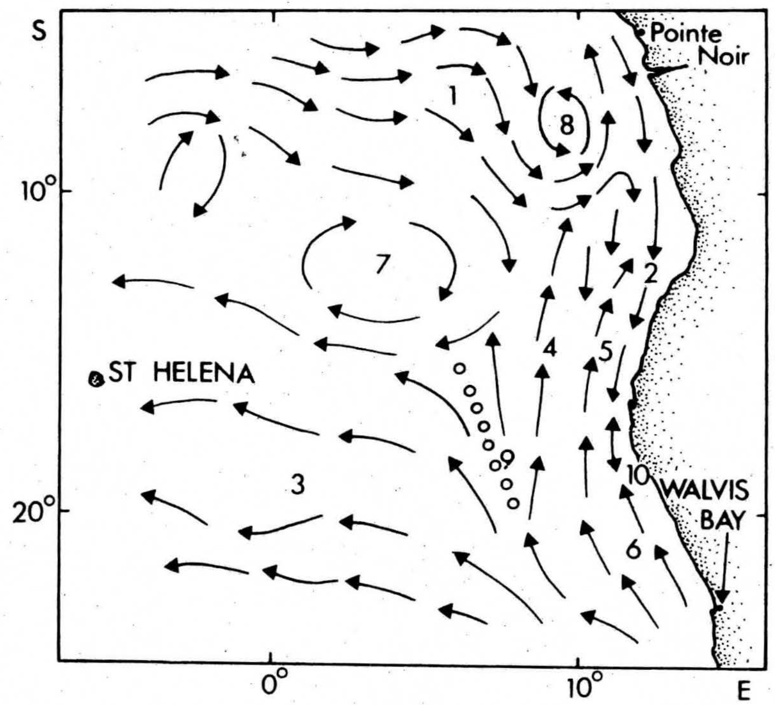


Fig. 34.—Geostrophic water circulation in the 0–100 m layer off Angola and Namibia (after Moroshkin *et al.*, 1970): 1, South Equatorial Countercurrent; 2, Angola Current; 3, West (main) branch of Benguela Current; 4, 5, 6, North branches of Benguela Current; 7, eddies in inner region of cyclonic gyre; 8, anticyclonic curl; 9, Benguela Divergence; 10, merging zone of Angola Current and north littoral branch of Benguela Current.

of Boyd & Agenbag, 1984a) is replaced by the southward horizontal advection of Angola water, as intimated by Hart & Currie (1960). The latter would explain the strong coastal downwelling and southward flow observed by Stander (1964, his plate 69) during October 1959 when upwelling was taking place further south (see also Fig. 30).

Stander (1964) noted that the surface flow becomes more complex and variable with decreasing latitude off Namibia, in particular between 23° S and 17° S, and his dynamic topography also shows similar complexity and variability at subsurface depths (100 m). The most comprehensive single investigation of the circulation off Angola and northern Namibia was undertaken between April and June 1968 (Moroshkin *et al.*, 1970). These authors identified northward filaments of the Benguela west of the coastal Angola Current, extending to 12° S to 13° S, and they postulated the existence of a divergence zone between these northerly and the main (westerly) branch of the Benguela north of 20° S. The schematic representation of flow in the upper 100 m proposed by Moroshkin *et al.* (1970) is shown in Figure 34 which, although for autumn, is in substantial agreement with the studies of Dias (1983a,b) off southern Angola and Stander (1964) and Robson (1983) off Namibia. The surface southerly flow inshore is most marked north of 17° S, while at this latitude Stander (1964) indicated a semi-permanent northerly flow between 100 and 200 km offshore and more variable flow near the "divergence". Filippov & Kolesnikov (1971) have, however, questioned whether the equatorward intrusions off Angola are of Benguela origin and whether the poleward flow off Namibia is an extension of the Angola Current. Nevertheless the available evidence does seem to suggest that subsurface meridional interaction between the Angola and Benguela systems can occur over a distance of about 1000 km.

MESOSCALE PROCESSES IN THE SOUTHERN BENGUELA

The general features of the oceanography around the South West Cape have been described by Isaac (1937), Clowes (1950), Darbyshire (1963, 1966), Shannon (1966), Bang (1973), Andrews & Hutchings (1980), and others, while Nelson & Hutchings (1983) have reviewed recent work on mesoscale upwelling processes in the region.

The dynamics of the southern Benguela upwelling system are governed largely by three factors, *viz.* mesoscale atmospheric perturbations, the topography, and the influence of the Agulhas Current system. The periodic relaxation in upwelling caused by the free zonal passage of easterly moving cyclones to the south of the continent, described by Nelson & Hutchings (1983), was summarized earlier in this review and readers are referred to Figure 9 (see p. 120). Typically wind reversals modulate the system with a period of about six days. These modulations are superimposed on the seasonal cycle, which is much more pronounced in the Cape Peninsula and Agulhas Bank areas than in the north. The seasonal wind stress curl diagrams in Kamstra (in press) illustrate this quite dramatically. In winter a broad convergence zone exists south and east of the Cape Peninsula with smaller zones in Table and St Helena Bays, whereas in summer the wind field around the South West Cape is predominantly divergent and strongly so immediately west of the Cape Peninsula. (These features are also evident from the study of Shannon, Chapman, Eagle & McClurg, 1983.) The topography of the southern Benguela region is complex. The coastline is irregular with several bays and capes and changes its general orientation markedly at Cape Point, just south of 34° S, while in places mountain ranges are present close to the coast (see Fig. 4, p. 113). These topographic features shear the wind stress field, giving rise to alongshore variability in the coastal upwelling (Jury, 1984, in press). Intrusions of water of Agulhas Current origin, which appear off the western coast either as shallow tongues or as advected eddies (sheared off from the current in the Agulhas retroflection area) provide a modulation with time scales ranging from a few days to several months.

Although Copenhagen (1953) identified both the Cape Peninsula and Cape Columbine areas as important upwelling centres, the tongue-like nature of the upwelling was first noted by Andrews & Cram (1969) in their classic aerial and shipboard study. Subsequent investigations have shown localized upwelling off other capes (*e.g.* off Cape Hangklip—Cram, 1970; Jury, 1980, 1984, in press). The mesoscale structure of the upwelling tongues is particularly noticeable from satellite thermal infrared and ocean colour imagery (*e.g.* Harris, 1978; Lutjeharms, 1981a; Shannon & Anderson, 1982; Shannon, Mostert, Walters & Anderson, 1983; Shannon & Lutjeharms, 1983). Figure 35 (from Harris, 1978) shows two broad but discrete tongues emanating from the Cape Peninsula and Cape Columbine systems during an advanced stage of upwelling, and well-developed upwelling can be seen extending as far east as Cape Agulhas.

Prior to the work of Bang & Andrews (1974) knowledge of the circulation around the South West Cape was based on data from drift card returns (Clowes, 1954; Duncan, 1965; Duncan & Nell, 1969), Pisa tubes (Duncan, 1966), ships' drift (*e.g.* Rennell, 1832; Defant, 1936; Tripp, 1967) and movement inferred from isentropic and *T-S* analysis (Clowes, 1950; Shannon, 1966; Darbyshire, 1966; De Decker, 1970) and the dynamic topography (Dietrich, 1935; Darbyshire, 1963). From the above a relatively simple picture emerged with cyclonic curving flow around the Cape, predominantly wind driven at the surface during much of the year, current reversals during periods of northwesterly winds, intermittent countercurrents close inshore, and an ill-defined poleward undercurrent. Much of this work has been synthesized by Harris (1978). The complexity of the current field was only fully appreciated after the advent of relatively recent current metering and buoy tracking programmes.

The area has a high degree of variability. Lutjeharms (1977) addressed the question of time variations in the spatial scales and intensities of ocean circulation around the South West Cape, and noted changes in the amplitude of the 26·4 sigma-*t* surface of up to 50 m. His energy distribution over the spatial spectrum showed two configurations; the most common one having an abrupt change in slope at 150 km (*i.e.* a predilection for disturbances with dimensions of less than 150 km) which compared well with the characteristic dimensions (120 km) of frontal eddies between 30° S and 34° S (Lutjeharms, 1981a). The second configuration, which showed a monotonic increase for all distances measured, was tentatively attributed by this author to sporadic extensive upwelling.

In the following paragraphs an attempt will be made to synthesize the present knowledge of mesoscale upwelling processes in the Cape Peninsula and Cape Columbine areas as well as the influence of the Agulhas system on the Benguela regime. Fine-scale investigations undertaken in semi-enclosed bays or close inshore and which relate primarily to tidal or wave driven regimes (*e.g.* Shannon & Stander, 1977; Gunn, 1977; Bain, 1983) have not been included.

INFLUENCE OF THE AGULHAS CURRENT SYSTEM

The southernmost part of the Benguela system, like the northern extremity, is influenced by a warm water regime. In many ways both the northern and southern boundaries can be considered as mirror images, pulsating seasonally, but out of phase by a few months (Shannon, 1977). The warmer surface waters present offshore in the south (>20 °C in summer, >16 °C in winter at 20° E) have properties of both South Atlantic and South Indian Subtropical surface water as well as Agulhas Current water. These waters are advected northwards around the Cape during the upwelling season (Darbyshire, 1963, 1966; Shannon, 1966; Duncan & Nell, 1969) and may result in the intensification of the horizontal frontal thermal gradients in the Cape Peninsula and Cape Columbine upwelling areas (Bang & Andrews, 1974; Lutjeharms & Valentine, 1981).

There has been some difference in opinion as to the extent of the penetration of the Agulhas Current water into the Benguela region (*e.g.* Shannon, 1966; Darbyshire, 1966; Orren & Shannon, 1967; Jones, 1971), although this can in part be attributed to the definitions of the water types. The surface water of the Agulhas Current is warm (generally >21 °C, but somewhat cooler in winter and typically >23 °C during summer—Shannon, 1970; Pearce, 1977) with a salinity of 35·1–35·4‰ (Pearce, 1977) which is lower than South Indian Subtropical surface water. It has a subsurface salinity maximum of about 35·5‰ (Clowes, 1950; Darbyshire, 1964; Pearce, 1977) which corresponds to a temperature of around 18 °C—*i.e.* similar *T-S* to that of much of the subtropical surface water in the South East Atlantic between 30° S and 35° S. Central water in the Agulhas likewise has similar *T-S* characteristics to that in the South Atlantic (Clowes, 1950; Orren, 1963; Shannon, 1966). Thus, except at the surface, it is extremely difficult to assess the degree of interaction between the Agulhas Current and Benguela on the basis of their thermohaline characteristics. Relatively pure Agulhas Current water does, however, penetrate into the South East Atlantic at times (usually during summer months when the wind regime facilitates this advection) in the form of shallowing warm filaments, generally less than 50 m deep or as eddies. Generally the tongues of warm Agulhas surface water are situated 180 km or further away from the coast (Shannon, 1966) and seldom penetrate much north of 33° S. One such tongue moving around the Agulhas Bank is evident in Figure 36a (from Bang, 1973, 1976).

In the area south of the Agulhas Bank the Agulhas Current executes an abrupt turn (Dietrich, 1935; Duncan, 1968; Bang, 1970; Gründlingh & Lutjeharms, 1979) and it is in this retroflection zone, which is one of high shear, that various independent circulation features may be spawned (Lutjeharms, 1981b). Lutjeharms & Valentine (1981) and Lutjeharms (1981c) have shown

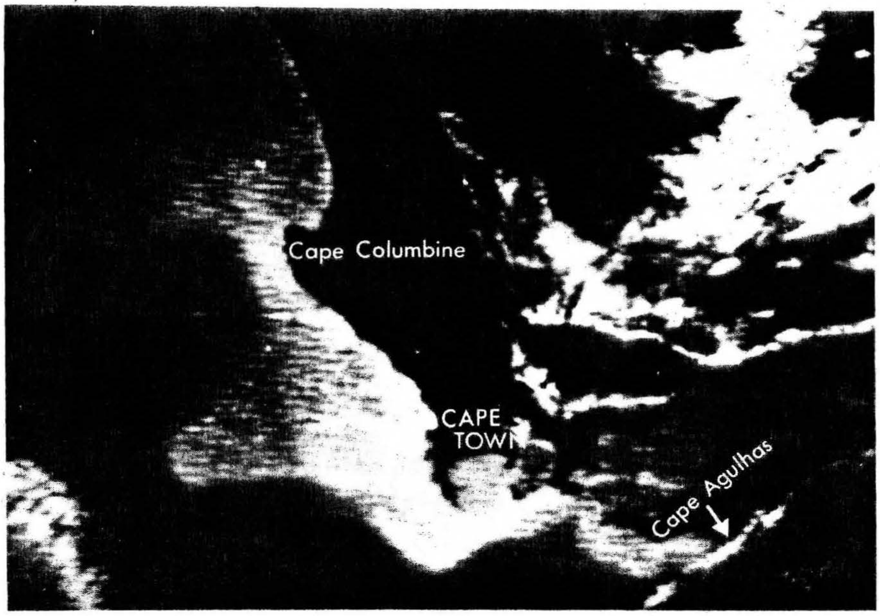


Fig. 35.—NOAA thermal infrared image of the S.W. Cape on 15th March, 1977 showing well-developed upwelling off Cape Columbine and the Cape Peninsula and extending as far east as Cape Agulhas (after Harris, 1978).

that at least two mechanisms exist for the advection of pockets of warm Agulhas Current water into the South Atlantic, *viz.* first, the growth of instabilities and the shedding of eddies on the northern border of the Agulhas Current and their subsequent advection across the Agulhas Bank and around the Cape and secondly, the formation of rings at the retroflection point southwest of the continent (similar to Gulf Stream rings). A drifting buoy placed at the centre of one such ring during May 1979 followed a meandering northward path west of, but in sympathy with, the Benguela thermal front, with the ring slowly losing its character due to mixing with colder surrounding water (Lutjeharms & Valentine, 1981). A conceptual image of the formation of eddies and rings is shown in Figure 36(b). Nelson & Hutchings (1983) suggested that the advection of vortices into the Benguela system could significantly alter the current flow patterns on a time scale of a few days.

Several authors, *inter alia* Dietrich (1935), Clowes (1950), Shannon (1966) and Darbyshire (1963, 1966) have shown that the movement of subsurface and central water components from the Agulhas Current around the Agulhas Bank and into the Benguela region is probable, and it has been suggested (Shannon, 1966) that modified Agulhas central water may upwell along the western coast at least as far as 30° S, and probably further north.

While there is some uncertainty as to the contribution of the Agulhas Current *per se* to the Benguela system, there is general agreement that, that of Agulhas Bank water is substantial, and that the region is ecologically of utmost

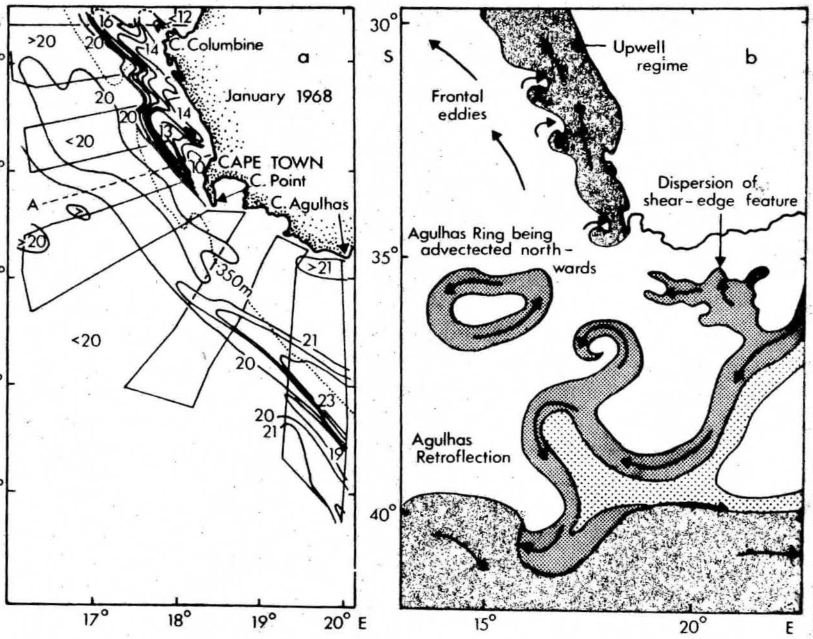


Fig. 36.—a, surface temperatures (°C) during January 1968 showing strong frontal feature pivoted on Cape Point and filament of Agulhas Current in the south (after Bang, 1973); b, conceptual image of Lutjeharms (1981b,c) showing formation of shear-edge eddies and rings.

importance to the Benguela. The formative mechanism of Agulhas Bank water is not properly understood but it appears that it comprises modified Agulhas Current and Atlantic water types. The Agulhas Bank is characterized by very strong winds during winter (Parrish *et al.*, 1983) which together with reduced stability facilitate mixing to a depth of at least 75 m (Pugh, 1982) and occasionally throughout the water column. During summer and autumn a strong thermocline is established at about 50 m which requires wind speeds greater than 20 $m{\cdot}s^{-1}$ for its erosion (Pugh, 1982). Net currents over the Bank are weak but indicate a strong diurnal activity (Welsh, 1964; Schumann & Perrins, 1983), and Shannon & Chapman (1983a) have suggested that a residence time of water on the Bank may be as long as several weeks. On the western side, cold water moves up onto the shelf during spring (Tromp, Lazarus & Horstman, 1975; Nelson & Hutchings, 1983), and remains on the shelf until autumn (the pronounced uplift in spring and early autumn evident in Tromp *et al.*, 1975, may be related to the bimodality in the southeasterly wind maxima in the area) and upwelling of central water can occur close inshore west of Cape Agulhas (Tromp *et al.*, 1975; Harris, 1978) and occasionally as far east as Algoa Bay (Schumann, Perrins & Hunter, 1982). Recent studies by Shannon & Chapman (1983a) on surface currents and winds and Schumann & Beekman (1984) on the thermal structure suggest that there is a divergence or transitional zone over the Bank around 21° E, which coincides with a break in the bathymetry, changes in sediment structure and composition (see Fig. 6, p. 115) and composition of fauna. It is also possible that internal waves of tidal character which are indicated by the strong surface signals in satellite images over the western Agulhas Bank (Nelson & Shannon, 1983) may play an important rôle in the mixing of subsurface and deeper waters in this region. Ships' drift (Harris, 1978), isopycnal and isentropic analysis (Shannon, 1966; Tromp *et al.*, 1975), drift cards (Duncan & Nell, 1969; Shelton & Kriel, 1980; Shannon *et al.*, 1983), *Physalia* (Shannon & Chapman, 1983b), and fish eggs and larvae (Shelton & Hutchings, 1982; Shelton, 1984) indicate a net westward, then northward movement around the Cape of surface Agulhas Bank water in the area west of 20° E during the summer (October to March) and an intermittent eastward movement in the inshore region further east. During winter and during periods of prolonged westerly winds the drift card returns suggest a net inshore eastward flow of surface water in the area between 18° E and 27° E (Algoa Bay), which implies that South Atlantic Subtropical surface water is seasonally important along much of the southern coast of the continent.

From the above it appears that, as in the northern Benguela–Angola system zone, the interactions between the Benguela and Agulhas systems can extend over a substantial distance, *i.e.* around 1000 km. Whether Agulhas water can maintain its identity as far north as 21° S on the western coast as suggested by Darbyshire (1966) is, however, questionable. Perhaps studies using stable isotope ratios or radionuclides (*e.g.* Shannon, Cherry & Orren, 1970) would be useful here.

THE CAPE PENINSULA UPWELLING AREA

The Cape Peninsula must surely rate as one of the most spectacular upwelling sites in the World. It is near the southern extremity of a subcontinent whose eastern boundary current system is bounded on its poleward end by a warm water regime. It is an area where bottom topography and orographically induced wind curl are at least as important as Ekman divergence in governing the upwelling process. It is a region where the wind field is modulated on a time scale of days as well as seasonally and where upwelling occurs on spatial scales from less than one kilometre to tens of kilometres, responding in a matter of hours to changes in the wind. It is also an area of immense natural beauty. The upwelling season off the Cape Peninsula extends from September to March (Andrews & Hutchings, 1980), with maxima in upwelling favourable winds in November and March at Cape Point. The temperature and salinity time series of Andrews & Hutchings (1980) showed that the properties of both the offshore water and subsurface (shelf) water followed a simple seasonal cycle. Distinct bimodality in the annual cycle was, however, exhibited by the nearshore surface (upper 30 m) waters, which may be related to the corresponding bimodality in the longshore winds.

The shelf break which is convoluted by the Cape Point Valley (Nelson, in press) is deep and the shelf itself is narrow with the result that cold water is available at subsurface depths close inshore, in particular near Cape Point, which Bang (1971, 1973) regarded as the southern pivot point of the Benguela system. The Table Mountain range which stands isolated to the flow of southerly wind creates local zones of strong wind curl and divergence (Nelson, in press). According to Nelson (in press) there are four definable classes of atmospheric events, *viz.* deep southeaster, shallow southeaster, northwester, and coastal low. Following on from the work of Andrews & Hutchings (1980), the wind field around the Cape Peninsula was studied during CUEX (Cape Upwelling Experiment) by means of aerial surveys augmented by coastal weather stations and data from shipping (Jury, 1980, 1981, 1984, in press; Nelson, Kamstra & Walker, 1983; Kamstra, in press). The characteristics of this wind field have been described in some detail by Jury (1980) and these have been highlighted in Shannon, Nelson & Jury (1981) and Nelson & Hutchings (1983).

The Cape Peninsula upwelling tongue or "plume" (see Nelson, 1981) which is characteristic of the alongshore variability in upwelling in the region was first documented by Andrews & Cram (1969), and was subsequently shown by Andrews & Hutchings (1980) to be a semi-permanent feature of the area during the summer months. Its surface features are masked during northwesterly winds but it only disappears entirely during the winter months (Andrews & Hutchings, 1980). These authors demonstrated that the pronounced short-term variability of the tongue and in the subsurface structure is imposed on the seasonal cycle, which they suggested was also wind-generated. Changes in sea surface temperature distribution in response to mesoscale wind events were examined by Jury (1980, 1984). His work has resulted in a substantially improved understanding of the dynamics of the southern Benguela, and the essential features of his study of the Cape Peninsula are briefly as follows.

The onset of southerly winds with wind speed maxima north of the Cape Peninsula, following the passage of a cold front, initiates upwelling off the northwestern Peninsula and entrains oceanic water and compresses the oceanic thermal front against the coast in the south. The travelling anticyclones passing south of the region create deep (2000 m) air flow from 160° when they are supported by an upper air ridge. This "deep southeaster" becomes topographically accelerated over the Cape Peninsula in a classic cape effect resulting in the growth of upwelling and the formation of pronounced tongues below the wind jets in Table Bay and immediately south of the Table Mountain chain (*i.e.* at Bakoven) as well as at other localities further south along the 60 km Peninsula mountain range. With the South Atlantic high ridging around south of the country the orographic influences are magnified due to the capping effect of the depressed inversion layer which results in the formation of a calm zone (wind shadow) north of the Peninsula and cyclonic wind vorticity. During this "shallow southeaster" phase the upwelling tongue and thermal oceanic front progress westwards and southwards. Finally, with the establishment of a coastal low and approach of the next cold front, wind reversals occur and northwesterly winds cause a relaxation in the upwelling and a compaction of the oceanic thermal front over the shelf region.

Jury (1984) has demonstrated that vertical wind shear controls the interaction of topography and winds in the area and his results indicated that vertical shears of $-2 \times 10^{-2} \cdot s^{-1}$ produced horizontal wind vorticities of $-6 \times 10^{-4} \cdot s^{-1}$ and alongshore sea surface temperature gradients of 1 °C per 10 km during the summer upwelling season. Two of Jury's (in press) case studies which illustrated the growth and decay of upwelling tongues off the Cape Peninsula are shown diagrammatically in Figure 37. Using data collected during CUEX, Taunton-Clark (in press) attempted to quantify the growth and decay of the Cape Peninsula upwelling tongue by comparing the areas encompassed by selected isotherms (measured by airborne radiation thermometry) with wind records from coastal weather stations. His results, although showing a fair degree of scatter, indicated a linear relationship between the plume area and the wind displacements, with a response time of between 12 and 24 h. The latter should be compared with the "almost immediate" response noted by Andrews & Hutchings (1980), and the rapid changes in superficial frontal gradients noted by Bang (1973) over 12 and 16 h. On a larger scale, Lutjeharms (1981a) obtained a fair correlation ($r = 0.73$) between wind stress and the area of upwelled water as inferred from infrared satellite images.

The existence of a well-developed front positioned approximately over the shelf break west of the Cape Peninsula and which forms the seaward boundary of the coastal upwelling region has been noted by various authors *inter alia* Duncan (1966), Andrews & Cram (1969), Bang (1973), Bang & Andrews (1974), Andrews & Hutchings (1980), and Shannon *et al.* (1981). According to Brundrit (1981) it could be established in perhaps five days of vigorous upwelling, but once established, its main features could persist over an entire upwelling season, particularly at subsurface depths. Indeed Duncan (1966), Shannon *et al.* (1981) and Hutchings, Holden & Mitchell-Innes (1984) have shown that the subsurface front is maintained during winter and during periods of sustained downwelling. Off the Cape Peninsula water of 9–10 °C is almost always present shorewards of the front at relatively shallow depths (50 to 100 m or less) even during winter, and this dynamic priming of the system implies that upwelling can be rapidly switched on and off by the wind. Andrews & Hutchings (1980) nevertheless show that cooler, lower salinity water is available at these depths in summer than in winter. Associated with the front and the shelf break off the Cape Peninsula is a well-developed equatorward jet (Duncan, 1966; Bang & Andrews, 1974) which appears to be

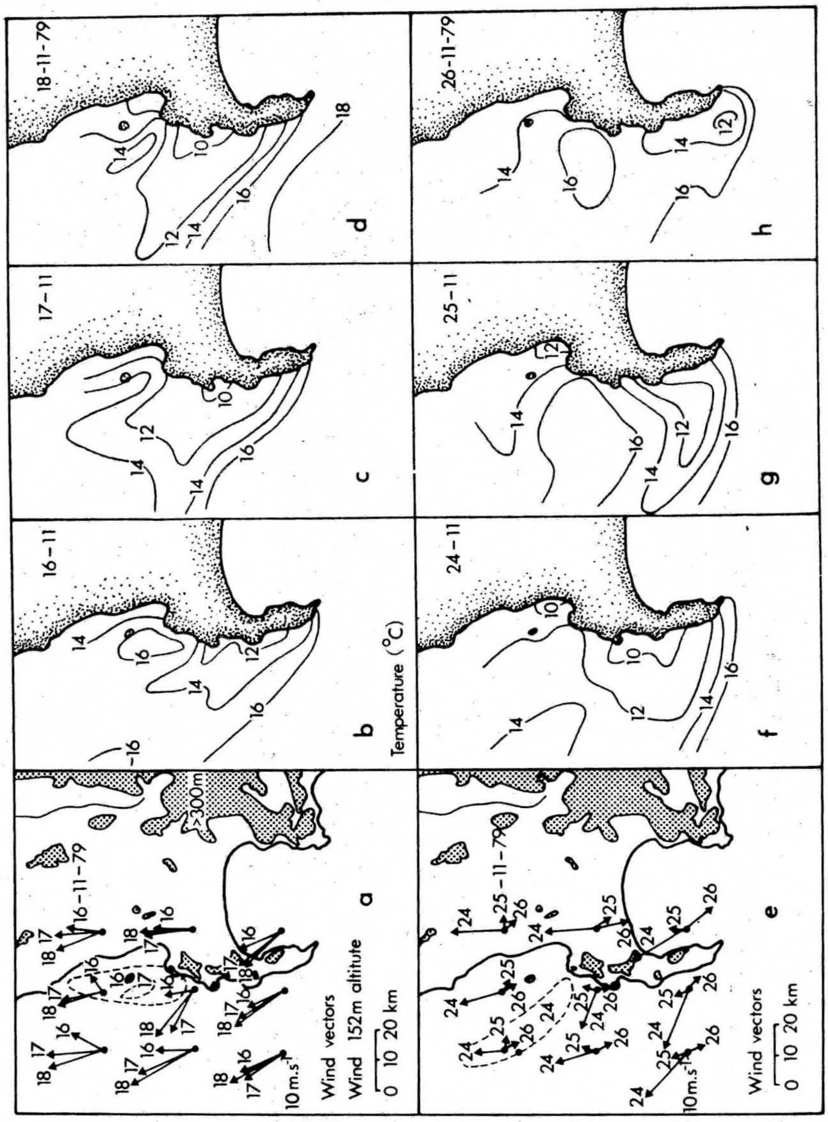


Fig. 37.—Growth and decay of upwelling off the Cape Peninsula (after Jury, in press).

more intense below the surface layer. Bang & Andrews (1974) have described how the front gradually moves seawards during periods of prolonged favourable wind stress as a result of potential energy being accumulated faster than the frontal jet can utilize it, and once it reaches the edge of the shelf it can expand downwards—*i.e.* an equilibrium position exists in the vicinity of the shelf break. These authors have suggested that the influx of warm water from the Agulhas Bank outside the Cape upwelling cell, by intensifying frontal gradients, favours the maintenance of both front and jet. The intense front and jet measured by Bang & Andrews (1974) during January 1973 is shown in Figure 38. Bang (1973), who first described the front in detail, drew a

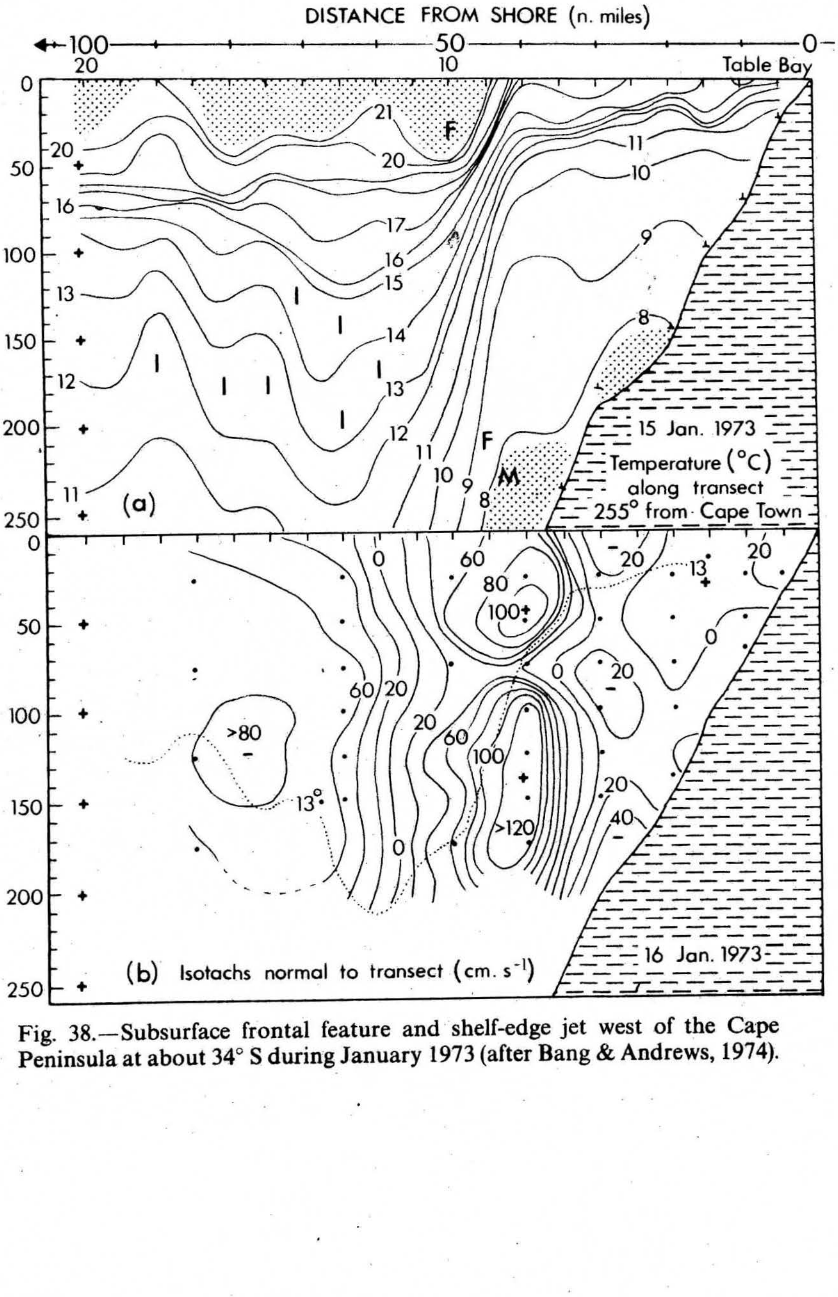


Fig. 38.—Subsurface frontal feature and shelf-edge jet west of the Cape Peninsula at about 34° S during January 1973 (after Bang & Andrews, 1974).

Fig. 37.—Growth and decay of upwelling off the Cape Peninsula (after Jury, in press).

distinction between the inshore frontal zone, formed by the outcropping of the seasonal thermocline 18–28 km offshore and which was dominated by the activity of mixing cells (shallow pockets imbedded in the front), and the main frontal system further offshore—his "offshore upwell region"—which had strong subsurface thermal gradients. Brundrit (1981) showed that the frontal jet is close to instability, and it has been suggested that small perturbations in the front can rapidly grow into fully fledged mesoscale eddies, as evident in satellite imagery of the region (Lutjeharms, 1981a; Shannon & Lutjeharms, 1983). These authors suggested furthermore that the cascading of potential energy at an intensive front caused by baroclinic instabilities could provide a potent mechanism for cross-frontal mixing. In this respect, Bang (1973) postulated that the observed mixing cells could provide a mechanism for cross-frontal advection of small particles, and that the action of the mixing cells within the front combined with overturning events to produce transient subthermoclinical sheets of relatively homogeneous water.

Andrews (1974) and Andrews & Hutchings (1980) identified four water types off the Cape Peninsula during the upwelling season, which they categorized as follows.

(a) Oceanic water lying outside the front and above the thermocline with a temperature greater than 18 °C and a salinity around 35·4°/ₒₒ.
(b) Upwelling water with a temperature of 8–10 °C and, an approximate salinity of 34·7°/ₒₒ.
(c) Mixed water, formed by the mixing of upwelling and oceanic waters and which generally exists as a wedge between the coast and the frontal zone, narrowing in the south.
(d) Shelf water with a temperature less than 8 °C underlying the upwelling water.

A possible criticism of this categorization is that it implies that only water of 8–10 °C upwells off the Cape Peninsula and that the water between 10 and 18 °C is mixed. While this may be the case during strong upwelling events, there is no reason to suppose that warmer more saline central water or mixed water does not upwell during periods of less intense winds. Bang (1973) suggested that the isothermal lens which he recorded immediately beneath and inshore of the surface front and which formed part of the shelf water indicated bottom turbulence. Shelf water is, however, normally isolated from the upwell circulation (Bang, 1973; Andrews & Hutchings, 1980). The oxygen-depleted water which appears near the bottom in late summer (De Decker, 1970) has the same *T-S* characteristics as upwelling water (Andrews & Hutchings, 1980). Nelson (in press) commented on the difficulties associated with *T-S* analysis in an upwelling region but noted that the characteristics of the central water off the Cape Peninsula differed from those of South Atlantic central water as defined in Sverdrup, Johnson & Fleming (1942). His work showed that the upwelled water originated from a maximum depth of 200 m and that its subsurface persistence was longer than two days.

Shannon (1966) intimated that breaking internal waves could be important in the local upwelling process, while Bang (1971, 1973) suggested that they might be responsible for large scale vertical mixing near the shelf break. Recently, internal tides with large amplitudes have been measured near Cape Point, and the question of these waves breaking in areas of steep bathymetry such as exist off the Cape Peninsula has been addressed by Nelson (in press). His work also showed evidence of turbulent mixing between 250 and 400 m west of the front.

Assuming a steady wind of 5 m·s^{-1} Shannon (1966) showed from theoretical considerations that water could upwell off the Cape at a rate of 11 m·day^{-1}. From the displacement of isohalines Andrews, Cram & Visser (1970) estimated a rate of around 21 m·day^{-1} with winds of about 10 m·s^{-1}, while Andrews (1974) and Andrews & Hutchings (1980) noted a linear relationship ($r = 0{\cdot}95$) between the rate of upwelling and the longshore wind speed, with a mean uplift during active upwelling conditions of 21 m·day^{-1} and a maximum of 32 m·day^{-1}. It should be noted, however, that these rates refer to part of the system for part of the time and as such are not true averages. From nutrient data Andrews & Hutchings (1980) obtained a mean upwelling volume flux for the Cape Peninsula cell of 9×10^9 m^3·day^{-1} which equates to an upwelling rate of 4·5 m·day^{-1} over the whole study area (2000 km^2). By comparison, Bang's (1976) estimate of a flux of $0{\cdot}5 \times 10^6$ m^3·s^{-1} for the Cape Columbine and Cape Peninsula zones (15 000 km^2) would give a flux of 6×10^9 m^3·day^{-1} for the Cape Peninsula. Bang (1976) also suggested that localized springs or fountains of upwelling of crystal clear ice-blue water of 8–9 °C with displacements of 30 m·h^{-1} might exist in the system. Andrews & Hutchings (1980) felt that coastal downwelling was a minor feature off the Cape Peninsula. Bang (1976) noted, however, that the concepts of upwelling systems in general are so pervaded by upwelling that the requirements of large scale sinking have received scant attention. Sinking has been noted near the front (Bang, 1973; Andrews, 1974; Andrews & Hutchings, 1980; Nelson, in press) and can occur there both as a slow diffuse process and in fairly dramatic overturning events (Bang, 1973). In addition, Bang (1976) suggested that parcels of cold water could break away from the cell and be dissipated by mixing and sinking. Both he and Nelson (in press) have drawn attention to the dissipative rôle which Langmuir circulations could play in the system.

The surface current patterns inferred from drift card recoveries (Duncan & Nell, 1969; Shannon *et al.*, 1983) have been substantiated by direct measurements using drogues and current meters. The earliest direct measurements of currents west of the Cape Peninsula (excluding nearshore pollution-related work) were made by Duncan (1966) using Pisa tubes (jelly bottles). His study showed the existence of three mesoscale cyclonic cells at a depth of 100–200 m in water of about 10 °C. The main eddy was situated off the northern part of the Peninsula and there was a suggestion of an equatorward jet of 27 cm·s^{-1} about 30 km offshore with southerly flow further east and a retroflection zone near 34°10′ S (Slangkop). His surface drifts under northwesterly winds were equatorwards (>50 cm·s^{-1}) over the shelf break west of Cape Point. The existence of a strong shelf-edge jet (Fig. 38) 60 km west of the Peninsula at 34° S, having velocities of about 60 cm·s^{-1} at the surface and 120 cm·s^{-1} at 150 m, was demonstrated by Bang & Andrews (1974). Their work showed poleward flow east and west of the jet with a poleward undercurrent of about 40 cm·s^{-1} near the bottom on the shoreward side of the jet. Bang (1974) suggested that the jet could be expected to be closer inshore off Cape Point, over the Cape Point Valley. The findings of Bang & Andrews (1974) and Bang (1974) were substantially confirmed by Shannon, Nelson & Jury (1981) using a profiling current meter during CUEX. (Nelson, in press, has discussed the CUEX results in more detail.) Except close inshore where shear was evident (noted also by Boyd, 1982), currents were generally barotropic. The jet (about 90 cm·s^{-1}) was most pronounced between 34°30′ S and 34°10′ S just west of the 230 m isobath, and there was evidence of a convergence zone around 34°10′ S, *i.e.* in a similar position to the area of retroflection noted by Duncan (1966). The results of an intensive buoy and drogue tracking experiment undertaken in the upper 50 m during CUEX have been reported by Nelson (in press) and again confirmed the existence of an equatorward jet of about 50 cm·s^{-1} over the shelf break. His study also showed the cyclonic eddy and the convergence zone near 34°10′ S. A buoy with a drogue set at 10 m was released southeast of Cape Point by Shelton & Hutchings (1982) and showed rapid movement (55 cm·s^{-1}) around the Cape along the oceanic front, accelerating to 70 cm·s^{-1} west of the Cape Peninsula. Other studies involving the tracking of drifters by ships and satellites have confirmed the existence of topographically controlled equatorward flow in the southern Benguela near the shelf break (Harris & Shannon,

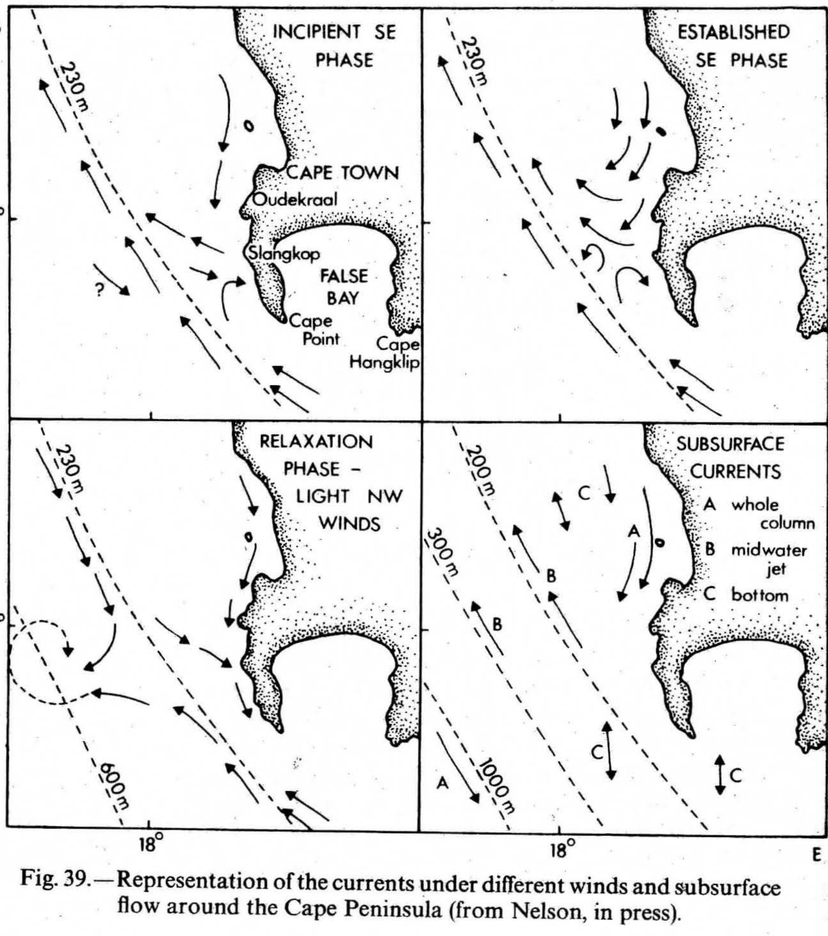


Fig. 39.—Representation of the currents under different winds and subsurface flow around the Cape Peninsula (from Nelson, in press).

1979; Nelson & Hutchings, 1983; Brown & Hutchings, 1984) and variable currents closer inshore (Van Ieperen, 1971; Boyd, 1982; Brown & Hutchings, 1984). The two mode numerical model of Van Foreest & Brundrit (1982) predicted the equatorward jet west of the Cape Peninsula in a similar position to that measured by Bang & Andrews (1974) and Shannon *et al.* (1981), as it did the poleward flow west of the jet. Their model also predicated that velocities would decrease with depth contrary to the observations of Bang & Andrews (1974) but in accordance with Shannon *et al.* (1981).

The only published information from moored current meters is that of Shannon *et al.* (1981) and Nelson (in press) who had two meters positioned within 50 m of the bottom in 253 m of water near the head of the Cape Point Valley during CUEX. The measured currents showed abrupt changes which correlated with local winds, the response time being a few hours, but evidently caused by pressure adjustments occurring on a scale many times larger than the Cape Peninsula upwell cell. Since the CUEX work, several moorings have been positioned in the vicinity of the Cape Peninsula. The results of this work are largely unpublished, but Nelson (1983) shows the existence of a southward coastal current and north–south direction switching over periods of typically six days, presumably as a barotropic response to atmospheric forcing. This appears to be a characteristic of the whole shelf zone as far as Cape Columbine (G. Nelson, pers. comm.).

The diagrammatic representation of near-surface and deep currents around the Cape Peninsula given by Nelson (in press) is shown in Figure 39, and illustrates a very plausible mechanism for the generation of frontal eddies as suggested by the author.

The dynamics of False Bay which forms the eastern seaboard of the Cape Peninsula does not appear to have any substantial influence on the upwelling system and, therefore, will not be discussed here. Readers are, however, referred to the following papers dealing with the Bay: Atkins (1970), Cram (1970), Harris (1978), Shannon, Walters & Moldan (1983), Jury (1984, in press), Van Foreest & Jury (in press).

THE CAPE COLUMBINE UPWELLING AREA

Upwelling in the Cape Columbine area, as off the Cape Peninsula, is controlled to a large extent by the bathymetry (see p. 174) and the influence of the orography on the wind field. Cape Columbine itself is a headland jutting seawards at 33° S consisting of low smooth hills rising to about 250 m (Jury, in press). The main shelf break is convoluted by a major submarine valley, namely the Cape Canyon (see Fig. 4, p. 113 and 41, see p. 174), the axis of which runs approximately from north to south. West of Cape Columbine the 300 m isobath is only about 30 km offshore, and this means that deep water is available close to the coast. Between 33° S and 32° S the shelf broadens substantially with the outer shelf break (400 m) running in a northwesterly direction, and the shallower isobaths tending to curve eastwards and northwards around the Columbine peninsula.

The surface wind vectors, stress and curl in the area between 31° S and 33° S, averaged over half degree rectangles of latitude and longitude, and described by Kamstra (in press) showed distinct seasonal variation. During summer the wind vectors indicated cyclonic curvature around Cape Columbine and into

St Helena Bay, with a zone of strong divergence evident from the wind stress curl north of 31° S. The pronounced curving of the wind has also been shown by the records from coastal weather stations (Hutchings, Nelson, Horstman & Tarr, 1983) and from aircraft drift measurements (Jury, 1984, in press). The last author also showed a distinct wake in the lee of the Columbine peninsula during a period of shallow southerly winds (inversion between 60 m and 350 m). Kamstra's (in press) winter wind data showed a rather confused pattern, which could have been expected as the area between 31° S and 33° S is intermediate between the zone of winter westerlies in the south and the perennial southerly winds north of 31° S. The two case studies of Jury (in press) showed the surface expressions of upwelling characteristic of periods of shallow and deep (inversion between 350 m and 500 m) southerly winds. In the shallow wind case the lowest sea surface temperatures occurred immediately west of Cape Columbine, approximately coinciding with a sausage-shaped zone of strongly negative wind stress curl, *i.e.* the classic cape effect off this headland. Reduced cyclonic curvature of the wind occurred under deep southerly wind conditions, with no significant headland wake, although a small convergence zone was evident in St Helena Bay (the latter could have implications for the retention of pollutants in the bay—see Shannon *et al.*, 1983). Evident in this case study was a cool tongue extending north from the Columbine peninsula and Ekman upwelling along the coast further north and east. The growth and decay of the Columbine upwelling tongue and coastal upwelling during a summer wind cycle was illustrated by Shannon, Schlittenhardt & Mostert (1984)—see Figure 40. The curving of the narrow tongue around the peninsula is evident—a feature often noted in the satellite infrared and ocean colour images of the region (Shannon, Walters & Mostert, in press). Several satellite images have indicated a subsequent anticyclonic curving of the tongue north of 32° S—*i.e.* an inverted 'S' configuration—suggesting topographic control. At times the base of the tongue or plume was around 33°20′ S (Shannon & Anderson, 1982).

The water masses present in the area have been described by Clowes (1954), Buys (1957, 1959), Duncan (1964), De Decker (1970), Bailey & Chapman (in press), and others. The properties of the surface layer (upper 20–30 m) are modified locally by the inflow from the Berg river during winter and in summer by insolation (Clowes, 1954), although recent studies have shown that the latter cannot account for the rapid warming of the water noted on occasion (G. Hughes, pers. comm.). Obviously advection, cross frontal mixing, and the entrainment of water in the area must be important considerations. According to Clowes (1954) a subsurface salinity maximum was common in the area during his study (1951 and 1952) at a depth of about 30 m, most marked during summer. The author attributed this to a subsurface current, although it may have been characteristic of advanced upwelling (Shannon, 1966). No estimates have been made of the rate of upwelling between 31° S and 33° S, although from the rapid response of the surface expression thereof (*e.g.* Fig. 40) it is probably of the same order as off the Cape Peninsula *viz.* a maximum of around 20 m·day^{-1}. The upwelled water originates from depths of 100–300 m (Clowes, 1954). An eight-year time series of monthly measurements in the upper 50 m at two stations, one just north of Cape Columbine, the other 120 km further west, showed distinct bimodality in the uplift of near-surface water (30–50 m), with maxima during October–November and March–April (Mostert, 1970).

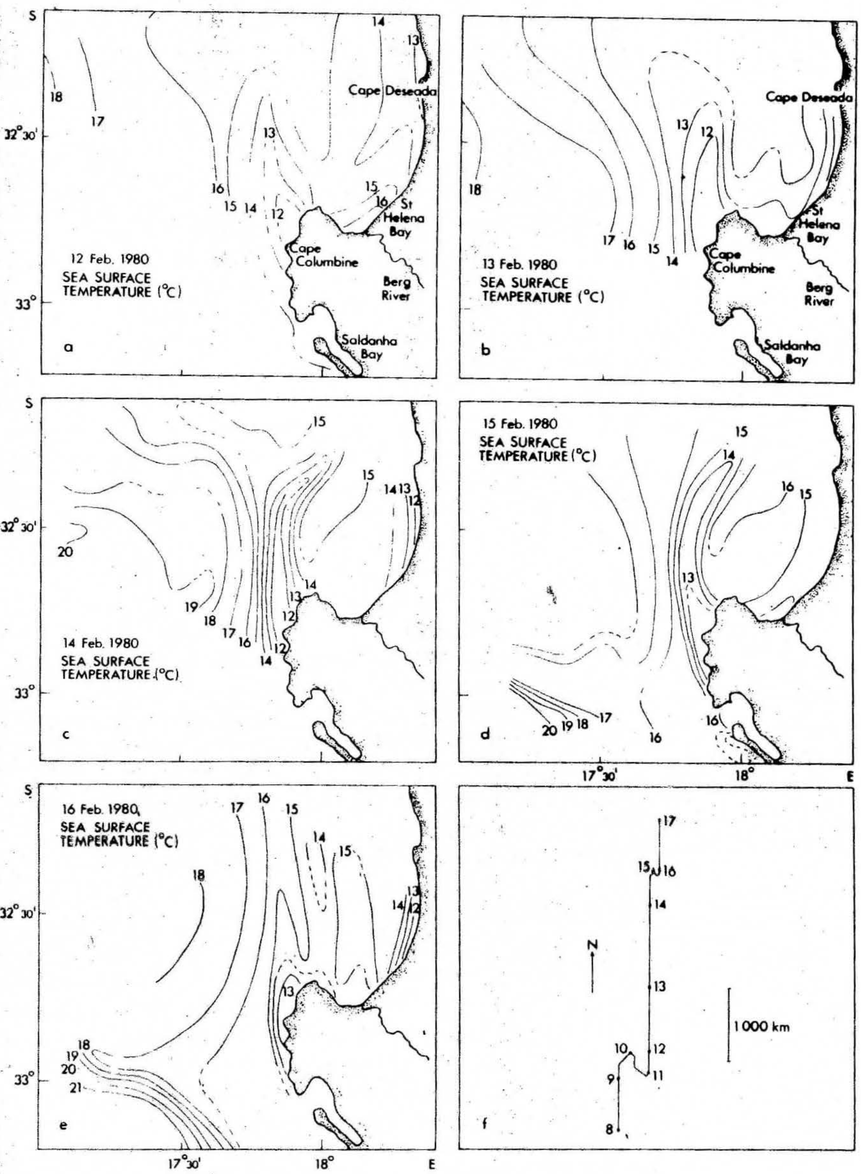


Fig. 40.—Changes in the Columbine upwelling tongue in response to local winds (after Shannon *et al.*, 1984).

The subsurface characteristics of the frontal systems in the Cape Columbine–St Helena Bay area have not been studied in any detail. The figures in Clowes (1950), Buys (1957, 1959), and Duncan (1964), however, showed the existence of a front in the upper 100 m around longitude 17°30′ E, *viz.* in the vicinity of the upper shelf break (20 km to 30 km offshore from Cape Columbine and further offshore north of 33° S) which suggest the existence of the shallow jet predicted by Van Foreest & Brundrit (1982). Few closely spaced measurements have been made in the region of the outer shelf break north of 33° S, except those of Bang (1976) along a line at 32°30′ S. His work showed clear evidence of a major subsurface front near the shelf edge during February 1966 (Bang, 1971; Jones, 1971), which he suggested was evidence of a baroclinic jet with an estimated surface velocity of about 60 cm·s^{-1}. His figure showed a second shallower jet at 17°30′ E. Also significant is that satellite thermal infrared images of the area (Shannon *et al.*, in press) showed close agreement between the position of the oceanic thermal front and the outer shelf break. Thus, there is a suggestion that, north of 33° S, there are two fronts and two jets, *viz.* the main oceanic front with which is associated the main shelf-edge jet (400 m isobath) and a second front immediately west of the Columbine upwelling tongue with which is associated a shallow ephemeral jet (200 m isobath). Satellite thermal infrared images (*e.g.* Figure 35, facing p. 158) sea-surface temperature data from aircraft (*e.g.* Fig. 40) and ships (Clowes, 1950; Shannon, 1966) often show a distinct westward bend in the oceanic front near 33° S. Thus, it seems that a divergence zone may exist at this latitude between the main shelf-edge flow and the shallower cyclonic curving jet west of the Columbine upwelling tongue. Inspection of figures in Clowes (1954) and Shannon (1966) seems to suggest that the divergence is most marked during late summer. From published diagrams it seems likely that the Cape Canyon may exert a substantial influence on both the circulation in the area and upwelling off Cape Columbine, although its precise rôle has still to be quantified.

Three questions arise. First, do direct measurements of currents support the concept of the postulated Columbine divergence; secondly, is there evidence of a poleward undercurrent; and thirdly, does sinking occur at the fronts? With respect to the last question the answer is probably yes. Clowes (1954) has shown strong evidence for this in some of his sections, while Shannon *et al.* (1983) showed a zone of high radiance in NIMBUS-7 CZCS band 520 nm immediately east of the oceanic thermal front, indicating the possible accumulation of floatables (oils, particulates) there, suggesting surface convergence. As to the existence of a poleward undercurrent, Holden (1984) has shown that perennial net southward flow occurred at depths below 80 m over the 180-m isobath west of Cape Columbine over periods longer than 10 days. His moored current meter data displayed a clear seasonal response in the poleward current with a maximum during winter (average 14 cm·s^{-1} and a minimum in summer (6 cm·s^{-1}). Occasional bursts of speed of up to 50 cm·s^{-1}, most frequent in winter, were noted. Further north the southward flow was weaker. His results have thus confirmed the existence of a deep compensation current postulated in the region. Low-oxygen water (De Decker, 1970) may at times be associated with a poleward coastal undercurrent. From data from the moored current meters and from a current profiling line west of Cape Columbine, Holden (1983) suggested that the interface between the surface and deeper currents sloped downwards away from the coast and that this interface tended to rise in winter, and occasionally during summer, to form a broad countercurrent at the surface, rather analogous with the California situation. This concept was to some extent supported by the intensive current profiling cruises undertaken within 20 km (180-m isobath) of the Columbine peninsula during September 1974 and June 1975 (NRIO, 1976). While the latter cruise showed southward or southwestward flow (10–60 cm·s^{-1}) throughout the water column under northerly wind conditions, the currents during September 1974 (southerly winds) were, however, equatorward and barotropic, around 50 cm·s^{-1}, showing no evidence of any poleward subsurface current. A drogue tracking experiment in November 1975 (NRIO, 1976) under strong southerly winds showed surface flow (typically $>$50 cm·s^{-1}) directed to the north with the strongest current or jet (up to 135 cm·s^{-1}) over the 100-m isobath *i.e.* in or immediately west of the Columbine upwelling tongue. A drogue released off the Cape Peninsula in recently upwelled water (Barlow, 1984) approximately followed the 150-m isobath northwards, and accelerated dramatically past the Columbine peninsula. The ship's drift measurements by Clowes (1954) east of 17°30′ E showed a similar equatorward set of about 50 cm·s^{-1} off the Columbine peninsula. Thus, there seems to be ample evidence of accelerated northward flow within 30 km of this peninsula which is probably strongest near the surface during summer, in particular during strong southerly wind events which favour the rapid development of the Columbine upwelling tongue. This is the Columbine jet. West of the 17°30′ E meridian the situation appears, however, to be somewhat different. The paths followed by three satellite tracked drifters during late summer to winter (Harris & Shannon, 1979; Lutjeharms & Valentine, 1981; Nelson & Hutchings, 1983) while also exhibiting accelerations west of Cape Columbine tended to diverge offshore in the region between 33° S (Cape Columbine) and 32° S, approximately following the run of the outer shelf break. Ship's drift observations (Clowes, 1954) showed similar westward or northwestward sets in these latitudes west of the 17°30′ meridian. Furthermore, the one current profiling line during June 1975 (NRIO, 1976) which extended 50 km offshore showed marked westward flow at the outer stations in sharp contrast to the currents within 20 km of the coast. Thus, there seem to exist two distinct branches of the current with a divide or divergence zone, the Columbine Divergence or Divide, near the 17°30′ E meridian. Whether the outer branch exists at subsurface depths is not certain, although the evident topographic steering of the satellite drifters and the shelf-edge jet inferred by Bang (1976) at 32°30′ S seems to suggest that it does. It is also not certain whether the two branches of the current coexist (Fig. 39, see p. 168, suggests that they do) or whether the current flips between the northward and westward configurations, with a preponderance for the former during spring and early summer and the latter from late summer to winter. If in fact the westward branch only forms in response to sea level adjustment during wind reversals then in fact the Columbine Divide could be associated with a confluence and retroflection of the currents rather than divergence. In any event the two branches do provide alternative mechanisms for the northward transport of material either into the St Helena Bay area or into the northern Benguela. The currents in the coastal area north of St Helena Bay are generally sluggish (Clowes, 1954; Holden, in press) and a cyclonic cell having mesoscale dimensions has been suggested (Duncan & Nell, 1969; Holden, in press). The residence time of water in this area is probably substantial. The northern boundary of this cell (30° S) is the perennial Namaqua upwelling area discussed earlier. The influence of the Namaqua region on the St Helena Bay area would appear to be largely restricted to the subsurface countercurrent, although a surface countercurrent may form in the absence of upwelling.

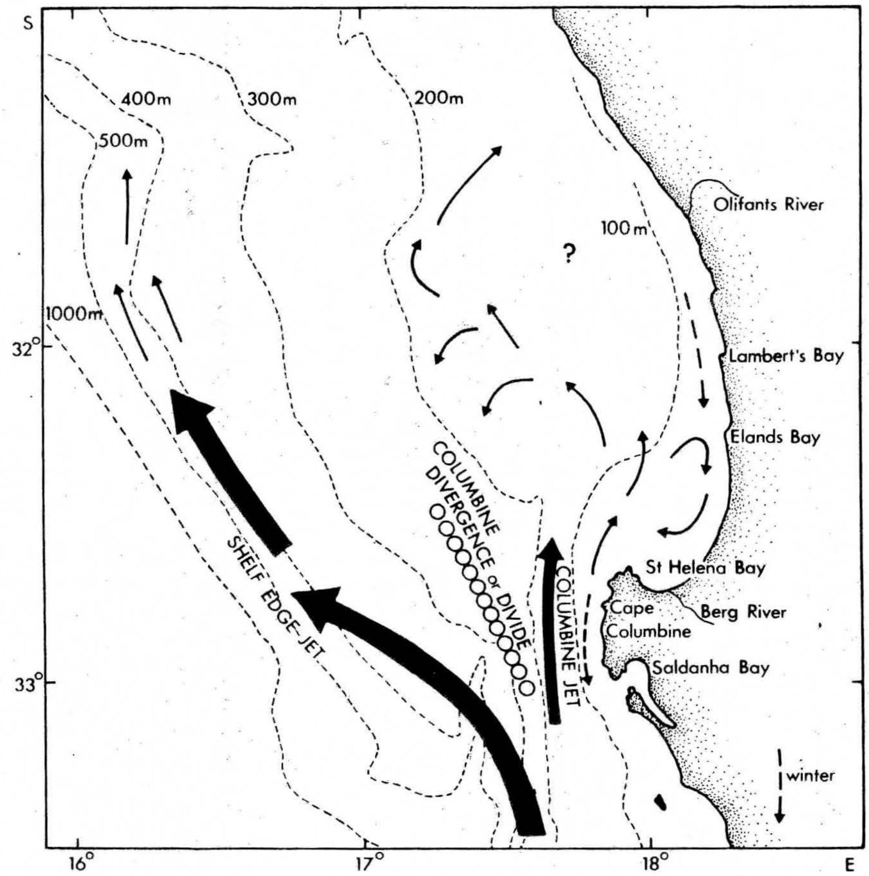


Fig. 41.—Schematic representation of currents in the Cape Columbine–St Helena Bay area.

The main currents in the region between 31° S and 33° S are shown schematically in Figure 41.

A CONCEPTUAL MODEL OF THE BENGUELA

The distinguishing features of the Benguela are highlighted in Figure 42 which is perhaps more of a conceptual image rather than a model. In it I have attempted to paint a cohesive picture of the system as a whole, and it is hoped that it may prove useful to workers in other disciplines and to those not familiar with the Benguela. While it does represent my interpretation of the system, it should, however, be viewed by readers in conjunction with the appropriate text and not in isolation.

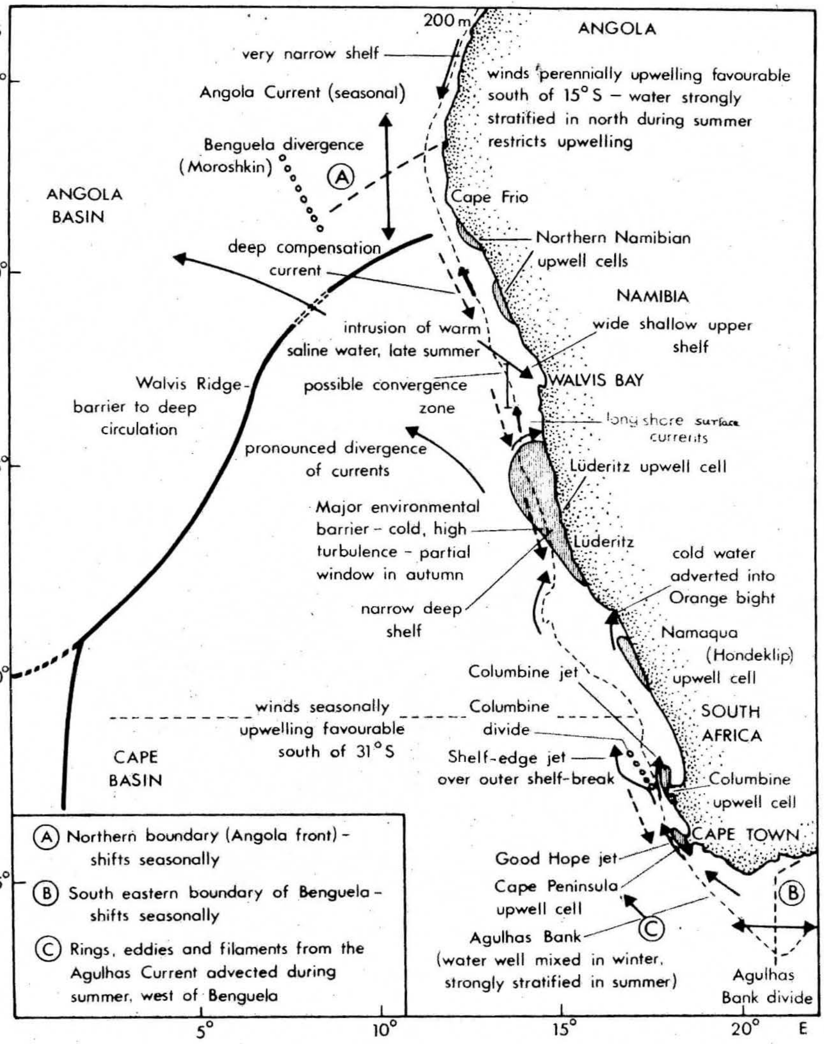


Fig. 42.—A conceptual model of the Benguela system.

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