3. Water circulation and sediment dynamics of large shallow lakes

Investigations to understand dynamics of lakes are all about seeking answers to basic questions such as (Bennett, 1974):

(i) What are the general patterns of large scale currents under different types of wind forcing?

(ii) What is the relative importance of physical factors such as geometry, friction, and waves in determining the current patterns?

Over more than a century attempts have been made to answer these and related questions by using both current measurements and theoretical analysis to explain observations. Theoretical studies (e.g. Rao, 1967; Csanady 1997; Koçyigit and Koçyigit, 2004) applying various simplifications have been supported with observations to reveal fundamental physics and significance of different factors. Water current measurements to derive resulting circulation patterns have been split in two distinct groups of either the drift bottle, drift card, or other type of surface water drift measurement, or measurement with fixed current meter on the other hand (Beletsky *et al.*, 1999). The former method, where the trajectories of moving tracer objects are used, is known as Lagrangian approach, while the later method, where time series of observations at fixed points are considered, is known as Eulerian approach.

Theoretical investigations to explain quantitatively the observational evidences employ assumptions either to simplify mathematical formulation or to isolate only a few of the physical factors and processes to identify their significance. Since the basic physics applies equally well to inland water bodies as well as to the continental shelf regions of the oceans, revision of theoretical studies and experimental results from different types will allow to draw some general conclusions about lake circulation patterns with varying environmental settings. It is therefore natural to begin with revising developed theories.

Much of the review materials in this chapter refer to numerous theoretical models and extensive works about circulation and mixing that have been published on the physical limnology of the North American Great Lakes. Spegel and Coulter (1996) note that significance differences in climate, latitude and basin shape make the direct application of some of the results to the East African Lakes questionable. For example, although the strength of density stratification in both tropical and temperate lakes is sufficient enough to influence water movements, the annual cycles of temperature stratification and prevailing wind over the tropical lakes follow important differences from temperate lakes. Nevertheless, the basic physics applies equally well to different types of lakes as well as to the coastal shelf regions of the oceans (e.g. Bennett, 1974). It is therefore felt that comparisons of developed models and observation results for lake basins with identical physical setup do provide both a framework for explanation of observations in the current study and develop working hypotheses to help organize further investigations.

3.1. Dynamics of winds in the coastal regions

The relative importance of the various relevant physical factors in determining the character of each type of wind classified according to dynamical principles is discussed by Jefferys (1922). He noted that not all of the forces due to the rotation of the Earth, friction and gravity are equally important under all circumstances. The large-scale upper layer air motion, corresponding to the geostrophic wind, is generated mainly by the balance between pressure gradient and Coriolis forces due to earth's rotation and flows nearly parallel to the isobars with velocities proportional to the pressure gradient force. The effect of friction, which is always opposite to the direction of the air motion and therefore tends to decrease the wind speed, cannot be neglected near the surface of the earth.

When the pressure difference between places at the same level is mainly occupied in overcoming friction, the resulting wind is called antitriptic wind. The Eulerian winds are characterized by accelerated motion of each particle of air due to the horizontal pressure gradient. The small-scale phenomena, which include all atmospheric disturbances whose horizontal dimensions in at least one direction are of the order of kilometres or tens of kilometres, embrace tropical cyclones, tornados, land and sea breezes and mountain and valley winds (Fleagle, 1950; Gleeson, 1953). Jefferys' analysis shows that tropical cyclones and tornados are Eulerian winds, whereas mountain and valley winds and sea and land breeze winds are mainly antitriptic.

The land and sea breeze circulation is one of the interesting mesoscale atmospheric phenomena occurring in coastal regions during periods of clear weather. A great deal of very detailed information from theoretical investigations and studies of observational evidences has come to light on land and sea breeze since the late 1940's (e.g., Haurwitz, 1947; Schmidt, 1947; Johnson and O'Brien, 1973; Arritt, 1993; Laird and Kristovich, 2001). In general, the occurrence of the land and sea breezes is due to the temperature difference between the air over land and that over water in the course of the day. This means that the phenomenon can be observed in its purest form when there is no general pressure gradient causing an offshore or an onshore wind, for in this case all air motions perpendicular to the coastline are due to the unequal heating mentioned above (Schmidt, 1947). On the other hand, if the winds connected to the large scale weather situation are strongly developed, their influence may overshadow that of the temperature difference (Haurwitz, 1947; Leopold, 1949; Estoque, 1962; Frizzola and Fisher, 1963; Johnson and O'Brien, 1973).

Diurnal variation of onshore wind in a costal area depends on the distance from the shoreline (Yu and Wagner, 1970; Crawford and Hudson, 1973; Johonson and O'Brien, 1973). A distinct wind maximum near shore follows the sea breeze front inland. The sea breeze can be classified as a gravity current (or density current) whose formation often intensifies during the early afternoon when the density difference between the land and sea air becomes large (Buckley and Kurzeja, 1997). The maximum sea breeze intensity occurs before the temperature difference between the air over land and over water has decreased to zero because of the effect of friction (Haurwitz, 1947; Weber 1979). Positive temperature difference is required to overcome the frictional force. As soon as the temperature difference has fallen below this critical value, the sea breeze starts to slow down. The stronger the friction, the larger is the critical value of the temperature difference necessary to maintain a sea breeze and, consequently, the shorter is the time between the maximum temperature difference and the maximum speed of the sea-breeze circulation. Stagnation point represents equilibrium due to balance between heating of the air column from below and advection of cold air from over the water (Staley, 1957). Another factor of possible significance for stagnation was the inertial oscillation associated with abrupt change at sunrise or sunset of the frictional force to a new relatively constant value.

The veering of the land and sea breeze in extratropical latitudes is due to the deflecting force of the earth's rotation (Schmidt, 1947; Haurwitz, 1947). In tropics, the sea and land breezes change direction when the general wind meets at an angle (Garbel, 1947). The resultant wind gradually changes direction during the day as the sea breeze component varies. Topographical constraints, such as irregular shoreline and the presence of mountains, result in non-uniform rate of turning of the direction of the sea and land breezes over the diurnal cycle (Frenzel, 1962; Alpert *et al.*, 1984; Kusuda and Alpert, 1983; Balling and Sutherland, 1988; Simpson, 1996).

Leopold (1949) described the interaction of trade wind and the sea breeze in Hawaii and the flow of air over the mountain masses of various shapes and sizes. He suggested that the interactions depend primarily on three factors:

1) the breadth of the area on which the heating and cooling will occur to cause the land and sea breeze, and which presumably governs the strength of the breezes so developed;

2) the height and shape of the mountain range which, in relation to the subsidence temperature inversion, will determine whether the trade wind will rise over the top or tend to split into two currents, passing one on either side; and

3) the aspect of the area on which the sea-land wind regime develops (exposure windward or leeward with respect to the trade wind).

With respect to the sea breeze over an area with irregular topography, its circulation is closely related in both theory and observation to other local winds resulting from differential heating, since the same forces are of importance (Fleagle, 1950; Gleeson, 1953; Stalely, 1957). The local topography is most important to air flow at individual locations. The land may contain valley or slope upward from the sea, causing the sea breeze to combine with other local winds such as air drainage and valley winds (Staley, 1957; Frenzel, 1962; Balling, 1988). Narrow valleys cause funnelling and greater variation in direction observed where the valley is relatively wide (Frenzel, 1962; Lemmin and Adamo, 1996).

The effects of various types of prevailing large-scale winds and thermal stratification on the development of the sea breeze circulation in straight costal regions are such that offshore prevailing geostrophic winds developed stronger sea breeze than onshore geostrophic winds (Estoque, 1962). Onshore large-scale flow produces weaker sea-breeze perturbations compared to those generated by an offshore flow (Zhong and Takle, 1992). This difference is due to development of strong pressure gradient as a result of the strong temperature gradient produced by advection of warm air over land towards the sea by offshore large-scale wind. On the other hand, the advection of cold air over sea towards over land inhibits the rise in temperature of the atmosphere over land. Consequently, the horizontal pressure gradient is weaker and a weaker circulation is developed.

A further effect of large-scale background winds on the characteristic of the sea-breeze circulations is that an onshore synoptic wind causes an earlier onset of the sea breeze, but delays the onset of land breeze, and a strong onshore flow of more than 5ms^{-1} does not allow the land breeze to develop at all (Zhong and Takle, 1993). The maximum offshore wind speed and vertical motion at night are less sensitive to the magnitude of surface cooling than to the large-scale flow and daytime surface heating, which together determine the initial flow at the beginning of the land-breeze phase. In addition, the magnitude, the sense of rotation, and the diurnal variation of the dominant forces governing the wind-vector rotation change as the orientation of the synoptic wind direction changes (Fisher, 1960; Neumann, 1977). The rate of rotation in the sea-breeze phase is dominated mainly by the balance between the mesoscale pressure gradient and friction; at night, the Coriolis effect also contributes significantly to the balance of forces in the land-breeze phase further away from tropics.

Examination of the characteristic features of sea breeze encompassing a broad range of onshore and offshore synoptic winds revealed that onshore synoptic flow of only a few meters per second was sufficient to suppress the thermally induced circulation, whereas for offshore synoptic flow as strong as 11 ms^{-1}, the circulation was still apparent (Arritt, 1993). Thus, a sea breeze circulation can exist entirely offshore, if the opposing flow is strong. The gradient wind is such that any land-breeze effect is only to reduce the night-time velocity but not to reverse the direction (Frenzel, 1962) Alongshore large-scale wind with low pressure over the sea produces the seaward frictional inflow in the lowest layers strengthens the horizontal pressure gradient in the same way as that produced by an offshore prevailing geostrophic wind (Estoque, 1962).

The most prominent feature of the general circulation occupying the lower levels in a vast area around the equator, broadly between 30^{o}S and 30^{o}N, is the trade winds that converge into the doldrums while being deflected westward (Petterssen, 1969; Nieuwolt, 1977; Riehl, 1979; Barry and Chorley, 2003). Trade winds, which originate at low latitudes on the margins of the subtropical high pressure cells, are strong and fairly steady (Petterssen, 1969; Riehl, 1979). Thus, the two general wind systems prevailing in the East Africa are the Northeast and Southeast Trade Winds (Bugenyi and Magumba, 1996; Barry and Chorley, 2003). The southeast and northeast trade winds prevail during the high and low sun seasons, respectively, and the associated oscillation of Inter Tropical Convergence Zone (ITCZ) north and south in an annual cycle (Nicholson, 1996; Barry and Chorley, 2003). For example, ITCZ is situated in January at about 15^{o}S and most of the East Africa is under the influence of northeasterly winds, whereas in July the ITCZ is about 15^{o}N and over East Africa southeasterly winds prevail (Nieuwolt, 1977).

Although the land and lake breeze regimes with a clear diurnal cycle in the tropics takes a course similar to that of higher latitude in summer, they generally show their most regular occurrence and strongest development in the tropics – as the heating of the tropical air over land can be up to five times that over adjacent water surfaces (Nieuwolt, 1977; Rierl, 1979; Nicholson, 1996; Barry and Chorley, 2003).

By way of summary, the dynamics of winds at coastal regions is a combined result of the passage of large scale pressure cells and small scale local winds. The land-sea breeze, the generation of motion by differential heating in coastal regions, is one of the most fundamental atmospheric processes. Interactions with other local wind systems and large-scale synoptic flow as well as physiographic characteristics may cause it to differ considerably from place to place. Thus, since the sea breeze is strongly affected by terrain, coast line shape, sea surface temperature, land temperature, and many other factors associated with the particular locale where the observations are made, it is felt that an many series of observations should be attempted at different locations in order to determine more accurately how much variation one might expect in the characteristics of the sea breeze from region to region.

3.2. Response of lakes to wind stress forcing

Work done by wind stress on a water surface generates turbulence that causes mixing in the surface layer, and horizontal currents and long waves that distort the thermocline, tilting it downward and bringing warmer surface water toward the leeward end of the lake (Spigel and Coulter, 1996). In a homogeneous lake with moderately strong, steady winds, steady currents are established whose directions are largely controlled by the wind direction and the presence of the shoreline (Murthy, 1971). However, steady and uniform winds rarely persist for more than a few hours, and in the normal course of events the winds often change slowly or vigorously or may even cease entirely for sometime. Since the hydraulic circulation (from inflow-outflow) is always small and thermally driven flow in nearly isothermal water must be small, lake currents should be wind driven (Pickett, 1977). Wind stress must play a key role in governing the transport and fate of riverine inputs (Balnton and Atkinson, 1983).

In general, the response of a lake to wind stress forcing depends both on the strength and the scale of the atmospheric disturbance (e.g. Spigel and Coulter, 1996). Since the stress of the wind over water is known to be a quadratic function (Charnock, 1955), the major prime movers of currents are short bursts of strong winds (Csanady, 1973). Strong winds usually have uniform spatial structure (Estoque, 1962). Because strong wind stress impulses are exerted on the water at random times, and because the strength and direction of these impulses vary in a random manner, water movements produced by them are always essentially in a transient state, there being never enough time to obtain any sort of a steady-state flow pattern. It is quite possible that some residual, long-term average circulation still exists, but the intensity of this is certain to be much less than that of the transient patterns (Csanady, 1973). This suggests that both the time history of the wind and the instantaneous value are important to understand wind-driven lake circulations.

Wind stress, which is caused by moving atmospheric disturbance, is known to have a major influence in lake water circulation and associated material transport. In the discussion of the dynamic response to a moving atmospheric disturbance, the scale of external disturbance is important (Rao, 1963). Because the scale of most natural lake bodies is small in comparison with the scale of cyclonic disturbances, the geostropic wind is usually uniform over one lake (Rao, 1967, and Csadady, 1968). Theoretical studies considered such cases as uniform wind. In addition, the rotary characteristics of the applied wind stress may be very important (Saylor and Miller, 1987).

Contrary to the uniform stress case which could be considered for very short duration, the spatial and temporal structure of the wind-stress field is variable. The manner in which the circulation in the lake adjusts depends, in addition to other lake's physical parameters, on the spatial and temporal structure of the wind field (Mohammed-Zaki, 1980). The rate at which energy is supplied by a time-dependent wind stress to the lake depends on the extent of the resonant coupling, which arises when the propagation speed of the atmosphere is nearly equal to that of free waves on the lake (Rao, 1967). Resonance, which corresponds to the maximum setup, is obtained when the speed of propagation of the stress band is the same as the free gravity wave for bandwidths less than the length of the lake. For bandwidths greater than the length of the lake, resonance is obtained for a speed of propagation of the stress band, for which the time taken by the band to cross a fixed point in the lake is the same as the time taken by the free gravity wave to cross the lake.

A good deal of discussion on forced motions of the model Great Lakes under three simplified wind systems: steady and uniform wind, uniform but periodic wind, and uniform potential vorticity built up by the curl of the wind stress over the past period presented by Csanady (1968). The steady and uniform wind corresponds to the geostropic wind, which is often uniform over one lake and of longer lifetime, whereas periodic wind is a relaxation of the steady and uniform wind to allow changes with the typical period of weather cycles. The horizontal stress variations to build the uniform potential vorticity may be caused by either the horizontal wind shear present in cyclonic disturbances or the relationship of the geostrophic wind to wind stress on the surface of the lake or both.

Csanady reported that the developed flow pattern, under no lateral and bottom friction assumption, when a uniform wind starts to blow over a circular lake at rest consists of two gyres, the maximum elevation (and velocities) occurring at the downwind end. When acted upon by periodic and uniform winds, the setup varies from maximum 90^{o} to the right of the wind at higher frequency to a maximum downwind, as in the steady uniform case, at very low frequencies. On the other hand, although wind stress of constant curl, applied to the lake surface for a certain period, setup a close circulation pattern of an amplitude proportional to the wind stress, Round-the-basin coastal jets setup by the wind stress curl are not likely to contribute significantly to current patterns observed (Rao, 1967).

In summary, the particular response of a lake depends on the characteristics of the forcing wind event. Wind energy is a most important physical, year-round control on circulation of tropical lakes in East Africa, where daily cycles in wind strength is one of the probable mechanisms to promote mixing of water column (Olago and Odada, 1996; Spigel and Coulter, 1996; Halfman, 1996). The annual wind cycle over the tropical lakes, with stronger southerly winds blowing along the axes of the lakes and persisting from May or June through August or September, results seasonal pattern of strong circulation and most mixing taking place during this period (Eccles, 1974; Coulter, 1968; Spigel and Coulter, 1996; Ochumba, 1996).

3.3. Large scale motion in lakes

The study of large scale motions in closed basins has received considerable attention over several decades. Bottom currents of a few centimetres per second on average may appear to be rather small, but steady flow of a 2 $cm \cdot s^{-1}$ equals about 50 $km \cdot month^{-1}$ (Saylor and Miller, 1987). Observations (e.g. Pickett, 1977) indicate that currents speed up, slow down, and change direction at different locations and depths. Even so, Pickett confirmed from inspection of the original data that most of the time, especially when speeds are high, the meters conform to the monthly mean pattern. Hence, one should be able to consider this mean pattern as the steady background flow driven by the monthly mean wind, but disturbed by wind variations.

The general circulation patterns identified from both numerical calculations (e.g. Gedney and Lick, 1972; Csanady, 1973; Bennett, 1974) and current meter records (e.g. Pickett, 1977; Saylor and Miller, 1987; Beletesky *et al.,* 1999) show that a top surface mass flux is being transported in the direction of the prevailing wind. A subsurface current driven by the surface gradients returns the surface mass flux in the opposite direction to conserve the volume of the lake water. The surface currents computed to 0.4 meters from the surface in the deepest and central basins are, in general, found to be smaller than near the shore for an actual shallow lake, Lake Erie (Gedney and Lick, 1972). It is noted that this effect is essentially due to the relatively large subsurface return current down the centre of the lake, which is opposite in direction to the surface current and subtracts from it. Furthermore, it was reported that the flow at all levels at many locations is essentially parallel to the shore.

The vertically integrated flow follows the same pattern; the water in the shallow region is accelerated in the direction of the longshore component of the wind and the flow is return in the deeper central region of the lake (Bennett, 1974). The transport distribution identified by Csanady (1973) in a long lake directly forced by suddenly imposed wind stress, for example, is such that the wind stress and pressure gradient are in balance at the average depth of a section and the computed transport is, hence, zero. Where the depth is less than the average depth of the section, wind stress overwhelms the pressure gradient and accelerates the water. Where the water is deeper, the reverse happens and the return flow develops.

The vector resultant currents computed on a monthly basis from current meter records yield a complicated set of charts on which concise and strong flow patterns did not persist for lengthy intervals of time stretching into many months, but they did sort into several frequently observed modes (Saylor and Miller, 1987). It is noted that the monthly or longer period averaged currents were composites of numerous wind-driven episodes as evidenced from the spectra of the wind stress which showed energy accumulations that represent some average values for the major weather system passages. A whole basin mixing episode of strong wind event that modifies the monthly averages plays a big part in determining the distribution of monthly resultant currents (Pickett, 1977; Saylor and Miller, 1987).

The overall circulation pattern could be either cyclonic (anticlockwise) or anticyclonic (clockwise) or consist of both gyres depending on the size of the lake: cyclonic pattern for large lakes and two gyre circulation pattern for small lakes (Beletesky *et al.,* 1999). The possible explanations suggested for such circulation patterns are that the meso-scale vorticity in the wind field due to large surface area and atmospheric temperature gradient would be the reason for cyclonic pattern in large lakes, and more uniform wind fields over small lakes to create two gyres pattern that resembles theories developed for circulation patterns under uniform wind. Csanady (1975) and Spigel and Coulter (1996) argue that a cyclonic circulation is required to achieve geostropic equilibrium, the balance between pressure gradient due to greater mixing and heating in the coastal regions than in pelagic regions and Cariolis forces that one would expect to hold in the long-term. Studies on tropical lakes (e.g. Eccles, 1974; Yuretich, 1979; Olago and Odada, 1996) show that wind-driven circulation, which appears to dominate any residual or long-term circulations, causes a closed-gyre circulation, with circulation centred on the main axis of the lake basins.

In summary fundamental principle of wind-driven flows in closed or partially closed basins states that in the shallow water, the dominant force balance is between surface wind stress and bottom friction, yielding a current in the direction of the wind. In the deeper water, the dominant force balance is between the horizontal pressure gradient (induced by surface slope) and bottom friction, yielding a current flowing in opposite to the wind (Hunter and Hearn, 1987). Major physical factors such as bottom friction, topography and presence of waves play interrelated roles that determine the hydrodynamics and associated transport process.

3.3.1. Effect of friction

The effect of bottom friction is to produce resistance opposing wind stress. Increasing effective bottom friction therefore reduces the strength of circulation. The controlling effect of the bottom friction to transport caused by wind stress depends on the depth distribution of the basin and the time scale of the wind stress forcing. In an initial period of depth-integrated transport directly forced by suddenly imposed wind stress is found to increase linearly in time; later, friction slows down this increase (Csanady, 1973). Thus, if the wind blew with constant force a very long time, the linear increase in transport would certainly be reduced by bottom friction. Furthermore, frictional effects are strong very close to the shores (within 7-9 km), but do not modify qualitatively the flow pattern which may be simply calculated from the frictionless, linearised equations (Csanady, 1973).

The straightforward damping effect of friction was revealed by the results of two models, where the one included friction and the other did not (Bennett, 1974). On the other hand, the relative importance of the lateral (depth-integrated) and overturning (vertically varying) circulations is dependent on the bottom roughness (Hunter and Hean, 1987). Increase in bottom roughness decreases lateral transport and hence flushing efficiency of the existing wind stress.

3.3.2. Effect of water depth

A uniform wind stress applied at the surface of a basin of variable depth sets up a circulation pattern characterized by relatively strong barotropic coastal currents in the direction of the wind, with return flow occurring over the deeper regions (Csanady, 1973; Bennett, 1974). Under quite general conditions the baroclinic response is confined in a narrow coastal region while the response in the main part of the basin is essentially barotropic. The barortopic large-scale low frequency motion follows contours of constant depth (Walin, 1972). The velocity distribution to suddenly applied wind stress shows the development of strong barotropic coastal currents (Csanady, 1973). The intense baroclinic motion in the coastal region will undoubtedly be a very important energy source for vertical mixing (Walin, 1972). For example water from deeper layers occasionally brought up to the surface will be more or less diluted with surface water before fining its way back to an equilibrium position in the deep layers.

The response of hydrographic systems to time dependent forcing caused by synoptic meteorological disturbances is such that unless the slope of the bottom is much smaller than the mean slope, the motion will be along the counters of constant depth (Walin, 1972). The time dependent response may be divided into two parts with distinctly different properties (Walin, 1972): (1) a barotropic part non zero in the whole basin, (ii) a baroclinic part non zero only in a relatively narrow coastal region. The width of the coastal region in which baroclinicity is a dominating feature is found to be of the order of 10 km.

A constant slope bottom causes a single gyre that intensifies the flow at portions of the boundary (Gedney and Lick, 1972). The variable slope effects of bottom topography cause the two gyre configuration, the asymmetries in the gyres being a direct result of the asymmetries in the bottom topography. Presence of narrow underwater ridges causes gyre and prevents subsurface return flow exchange (Saylor and Miller, 1987). The circulation pattern is complicated further by islands that limit basin interactions through the restricted passages.

Hunter and Hearn (1987) investigated the influence of basin bathymetry and bottom friction in determining the relative importance of depth-integrated transport at right angles to the motion vector (lateral) and vertically varying (overturning) circulations. They noted that these two elements of the total circulation can act to flush the natural system. However, under conditions of uniform wind stress and a flat-bottom basin the circulation consists only a vertical variation in the current (overturning). The interaction of variable topography leads to the development of lateral circulation in the basin. Their analysis showed that the ratio of the lateral to total circulation decreases with increasing bottom roughness or a flatter topography, due to reduction in lateral circulation. The ratio approaches a maximum value of unity for low bottom roughness (or small drag coefficient) and a broad distribution of depths.

There is a large difference in flushing rate (or exchange transport) between flat bottom and sloping bottom case (Singell *et al.,* 1990). In the flat bottom case, continuity requires the depth-averaged flow to vanish; thus the exchange transport limited by the amount of vertical shear that can develop. The bottom stress is small because the bottom velocity is small, and the primary balance is between wind stress and depth integrated pressure gradient. In the sloping bottom case, the depth averaged flow at any particular depth value need not vanish to satisfy continuity, and strong horizontal structure develops with downwind transport in the shallows and upwind transport at depth. In shallow water, the primary stress balance is between wind stress and bottom stress, while in deep water, wind stress balances the integrated pressure gradient. As a result there is a downwind transport driven by the wind stress, and upwind transport in the deep water by the pressure gradient. Departure of the depth from the mean, therefore, increases the efficiency of the wind-driven flushing. For homogeneous flow in realistic basins, most of the exchange transport occurs in the horizontal (Hunter and Hearn, 1987; Singell *et al.,* 1990).

3.3.3. Effect of surface waves

When waves and currents exist jointly in a coastal region, the shear stress identified with the wave and current are altered because of the nature of the turbulence generated by wave-current interaction at the bed and are different from the stresses expected in the case of pure waves or currents. The turbulent momentum transfer from the near-bottom mean flow to the bottom is increased by the additional turbulence generated at the bottom in the thin wave boundary by the oscillatory wave velocities (Signell *et al.,* 1990). The resultant is that the current in the region above the wave boundary layer feels a greater resistance that associated with the physical bottom roughness.

Analytical theory presented by Grant and Madson (1979) described the significance of waves in modifying the flow in the vicinity of the bed. They noted that waves are capable of entraining significant amount of sediment from the seabed when current of comparable magnitude may be too weak even to initiate sediment motion. On the other hand, waves are an inefficient transporting mechanism, and to the first order, no net transport is associated with the wave motion over a wave period. However, the simultaneous presence of even a weak current will cause a net transport. Their theory shows that when waves and currents exist jointly in a region, the shear stresses identified with wave and current are altered because of the nature of the turbulence generated by the wave-current interaction at the bed and are different from the stresses expected in the case of pure waves or currents. The net result is that the region above the wave boundary layer feels a greater resistance than that associated with the physical bottom roughness.

The magnitude of wind-driven currents in shallow bodies of water may be significantly influenced by wave-current interaction (Signell *et al.,* (1990). Surface wave in shallow water can increase the resistance felt by the bottom currents by an order of magnitude over the case of a current flowing over a rough bottom without waves present. Signell *et al.* (1990) compare results obtained using flat and sloping bathymetry cases. The results show that reduction of the exchange transport is larger for the sloping bathymetry case, whereas the reverse is found to be true for the increase of setup due to the presence of waves. If the wind stress is of sufficient strength and duration to generate surface waves which feel the bottom in at least the shallower part of the bay, the near-bottom current feels an enhanced drag which yields diminished velocities in the shallow water and reduces the overall exchange transport. If wave response varies with wind direction, then an oscillatory wind with a quasi-steady current response could generate a vertically uniform horizontal circulation that could increase the flushing. Overall, the magnitude of wind-driven currents in shallow bodies of water is significantly influenced by wave-current interaction.

3.4. Sediment dynamics

It is widely accepted that the major sources of sediments are inflow from rivers, shoreline erosion, dead organic material and dumping of dredged materials (e.g. Madsen *et al.,* 2001; Brigham *et al.,* 2001). The rate of infilling by sediment depends on the sediment supply from the catchments and aquatic sources, and the original volume of the lake basin (Ryan *et al.,* 2003). A thorough understanding of the process of sediment-water interactions and the dispersion of sediments in a lake will have significant implications for many important problems (Sheng and Lick, 1979). The way in which sediments accumulate in a lake is important because sediments provide a record of past events in the lakes and in their watersheds (Davis and Ford, 1982; Blais and Kalff, 1995). Lake deposits reflect environmental conditions within the whole drainage basin – climate, hydrology, sediment production, land use, water pollution, etc (Sundborg, 1992). Large amounts of contaminants such as nutrients and pesticides are transported into large lakes while being absorbed onto or associated with fine-grained sediment particles.

River mouth morphologies and depositional patterns comprise a broad spectrum of types (Wright, 1977). Wright (1977) discusses that when the wave power of the receiving lake basin is negligible or small relative to the strength of river outflow, effluent behaviour and consequent depositional patterns depend on the relative importance of three primary forces: (1) inertia and associated turbulent diffusion; (2) turbulent bed friction; or buoyancy. The role played by each force depends on factors such as the discharge rate and outflow velocity of the stream, water depths in and lake-ward of the river mouths, the amount and grain size of the sediment load, and the sharpness of density contrasts between the river and basin waters. High outflow velocities, small density contrasts, and deep water immediately lake-ward of the mouth permit inertial forces to dominate, causing the effluent to behave as a fully turbulent jet. When bed-load transport is large and water depths lake-ward of the mouth are shallow, turbulent diffusion becomes restricted to horizontal while bottom friction increases deceleration and expansion rates. Where the river mouth is deep relative to the riverine discharge, lake water enters the mouth as a salt wedge, and the buoyancy of the lighter river water becomes dominant; the effluent then spreads and thins as a relatively discrete layer.

River mouths located in relatively protected environments of fronted by flat offshore slopes may experience minimal wave effects. In a surface-trapped river plume, an external forcing (e.g. wind or ambient current) is required in order for entire fresh water volume discharged by a river to be transported downstream (Fong and Geyer, 2002; Fong and Stacey 2003; Warrick *et al.,* 2004; Piñones *et al.,* 2005). On the other hand, river mouths fronted by steep nearshore slopes in high wave-energy environments are profoundly influenced by waves. Outflow from river mouths reflects and steepens incident waves in such a way as to concentrate power on the effluent and to cause breaking in water depths greater than the normally breaking depth. The resultant wave-induced setup opposes the outflow, while wave-breaking enhances the mixing and momentum exchange between the effluent and ambient lake water. The effect is to cause very rapid deceleration and loss of sediment transporting ability within short distances from the outlet.

The initial direction of sediment transport at river mouth is determined by outflow pattern, but the subsequent dispersal of sediment is regulated by the mixing rates of the buoyant plume with ambient water (Van Maren and Hoekstra, 2005). Dynamically, advection is likely to dominate the transport in the primary flow direction, while the lateral and vertical structure and the temporal variability of the plume is determined by environmental dispersion (Fong and Stacey, 2003). Frequently, well-mixed conditions can be reached quickly in the vertical direction and the dispersion of sediment will occur primarily in the lateral horizontal direction, orthogonal to the dominant advective direction. When the mixing rate of surface waters with underlying water is low, then the removal rate of sediment from the plume is dominated by sediment settling velocities; otherwise mixing processes contribute to sediment removal and further downstream transport as well. River discharges into saltwater bodies also have been found to form sharp frontal boundaries, across which density changes abruptly, at times of high discharge (Garvine and Monk, 1974).

The resuspension and deposition of fine-grained cohesive sediments in lakes are dependent upon the shear stress applied at the sediment-water interface, on the bulk sediment water content, on the mineral and size composition of the sediment and the absence or presence and amount of benthic infauna and bacteria (Sheng and Lick, 1979; Fukuda and Lick, 1980). Waves breaking on the shore of a lake will create turbulence and resuspend material in the shore zone that will then be redeposited in deeper waters (Håkanson, 1977). In general, increase in water content, bottom shear stress, or clay content of the sediment results increase in resuspension rate. The energy parameter with the most direct impact on the bottom dynamic situation is the wave length, which in turn depends upon the effective fetch, the wind velocity and wind duration (Håkanson, 1977; Sheng and Lick, 1979; Douglas and Rippey, 2000). Johnson (1996) suggests, based on review of the evidence for various sedimentary processes that have been inferred from seismic reflection and side scan sonar profiles from the east African lakes, that sedimentation pattern in the large lakes of the East African Rift Valley that surface waves generated by strong winds can sort and redistribute sediments in water depths as great as 100m. The strong prevailing winds blowing over the rift valley lakes set up longshore currents in the nearshore zones that can generate significant lateral transport of sediment (Yuretich, 1979; Johnson, 1996).

Among 10 mechanisms identified to control distribution of sediments in lakes (Hilton *et al.,* 1986), the redistribution of settled material or occurrence of sediment focusing, resuspension of sediment in the shallower zones by waves and water currents with subsequent transport to and settling in the deeper zones of lakes, can be rationalized in terms of the dominance of four processes: peripheral wave attack, sliding/slumping on the slopes, intermittent complete mixing, and random redistribution (Hilton, 1985). These mechanisms, except random distribution, may cause sediment focusing (Hilton *et al*, 1986).

Based on their potential for resuspension, lake bottoms can be divided into erosion zone, transportation zone, and accumulation zone (Hakannson, 1977; Blais and Kalff, 1985). Erosion zone is marked by course-grained, non cohesive sediments and is found in areas of high water turbulence. Areas of sediment accumulation are characterized by soft deposits of fine material with relatively high water content, whereas areas of bottom transport may be characterized by discontinuous deposition of fine materials (Håkanson, 1985).

Overall, the basic textural trends and compositional variations of sediment distribution in lakes is known to conform to an inshore-to-offshore prograding associated with declining energy as water depth increases (e.g. Holmes, 1968; Thomas and Kemp, 1972). The rapid energy decrease at stream mouths appears to allow deposition of much of the sand and silt (Holmes, 1968). Transport dynamics of fine-grained sediments are functions of the energy imparted to the immediate surroundings (Petticrew and Kalff, 1991). Due to their small fall velocities, fine grained particles (i.e. those in the silt and clay size range) are transported easily by flow (Luettich *et al.*, 1990). The fine material is not deposited in high energy environment, e.g. in the beach zone, at lower water depths relative to large fetch, or when the water velocity for different reasons is high (e.g. in connection with topographical bottle-necks) (Håkanson, 1977). As the energy level further decreases out into the lake clay particles begin to flocculate and the floccules are deposited (Holmes, 1968).

Further, the mineralogical composition of modern lake sediments is intertwined with geological, climatic, biological, chemical and physical processes within the basin (Thomas, 1968; Yurtich, 1979; Hilton and Gibbs, 1984; Johnson, 1996; Halfman, 1996). For lakes in the East African Rift System, the timing and style of tectonic movements has influenced drainage pattern development, hydrological budget, and the amount and type of sediment supplied to the rift (Wescott *et al.*, 1996). The length of the rift segment has a significant impact on how effective the axial drainage is in affecting sediment distribution and sedimentation rates within a segment (Wescott, *et al*., 1996). The great length means that the effect of axial drainage is limited to the north most part of the rift segment, whereas the shorter lake segment in contrast is greatly affected by axial drainage in either direction. Localized compositional character of sediments in the vicinity of river deltas and near-shore localities is closely controlled by the detrial load and local source areas (Horowitz, 1974; Yuretich, 1979). Variations in distributional patterns of lake sediments offshore define specific sedimentological provinces (Yurtich, 1979; Hilton and Gibbs, 1984). Differences in mineral densities may also reflect mineralogical trend in lake sediments (Kennedy and Smith, 1977).

3.5. Study Objectives

In view of its large horizontal extent (more than 80 km), topography of surrounding land and existence of blocking islands distributed in the lake (about 2.5% of the lake area), the hydrometeorological condition of Lake Abaya may vary from place to place. This in turn could give rise to different zones characterized by spatiotemporal patterns of lake water circulation and associated transport processes. Consequently, the characterization of the spatial and temporal variability of physical parameters and sediment distribution is the main subject of this research. More specifically, this study addresses the following distinct, but related issues:

- Study the importance of meteorological forcing on lake water circulation, which influences many processes, including sediment transport and mixing for water quality considerations;
- Determine the depositional zones for sediment transported within the lake;
- Define the lake basin-scale magnitude and variability of fundamental water quality parameters; and
- Characterize the mean lake water circulation in terms of sediment distribution and lake geometry (local bathymetry, shape and islands) and their interactions with top-level forcing of the lake.