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RESEARCH ARTICLE

ENERGETIC POWER ESTIMATION OF SWELLS AND ORBITAL MARINE CURRENTS IN BENIN COASTAL ZONE (GULF OF GUINEA)

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Abstract

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The announced depletion of fossil energy resources is fueling some speculation. It is therefore urgent to explore the different renewable energy sources available to Benin. In the ocean, in addition to swells, there are also orbital marine currents generated by them. An advantageous solution for the countries located in the Gulf of Guinea is the recovery and transformation of the energy of the swell and its orbital marine currents into electrical energy. Swells are adult waves that propagate freely at the atmosphere-ocean interface. In this work, based on measurements made by the Millenium Challenge Account (MCA-Benin) as part of the extension of the Autonomous Port of Cotonou, we have shown that the swells at the Autonomous Port of Cotonou are Airy swells or of Stokes. We have identified the modified Boussinesq equations, proposed by Peregrine in 1967, as the theory for characterizing the evolution of these swells during their formation in the coastal zone. The variations of the different wave parameters (height, wavelength, period, phase and group velocities) are characterized from the shoaling zone to the breaking point. With this theory the results obtained are in agreement with the experimental measurements made, they perfectly describe the amplification of the height of these waves and show that these swells become very energetic when they approach the coast.

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Introduction:-

In the Gulf of Guinea, there is a degradation of the coastal strip due to the mechanical energy of the waves approaching the coast. On this coast, we observe the propagation of quasi-permanent swells whose height increases under the effect of the seabed and its slope in the coastal zone, which strongly induces coastal erosion. To ensure energy autonomy, the exploration of all available renewable energy sources is essential. The energy situation in

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Benin is characterized by a low rate of access to electricity and an almost total dependence vis-à-vis the exterior for electrical energy. Coastal country of the Gulf of Guinea, it has access to the ocean to the south over a length of approximately 125 km. To ensure energy autonomy, the exploration of all available renewable energy sources is essential. With the advent of climate change, reflection is increasingly focused on the optimal exploitation of the energy potential of each region [1]. Renewable energies are one of the keys to solving the energy crisis in the West African sub-region [2]. The ocean, the great forgotten in the assessment of the energy potential of the sub-region is however an almost inexhaustible source of energy. Marine dynamics is a source of energy that is still poorly documented and therefore not exploited in Benin[3]. The swells transport a significant amount of energy which they derive from the force of the wind on all the seas of the globe and which is dissipated by the bathymetric breaking on the coasts: these are renewable energy sources [4]. An advantageous solution for Benin and the countries of the Gulf of Guinea in general is the recovery and transformation of energy from swells into electrical energy.

Breaking is a dissipative process of energy which corresponds to the last stage of the life of a wave and which therefore most often takes place when approaching the shore. Wave breaking is an essential aspect of sea state dynamics, and in particular of coastal dynamics. Dissipative processes, including breaking, are one of the important terms of the energy balance for the wave field and require adequate parameterization in the calculation of sea states. However, breaking is still poorly understood and poorly parameterized. Senechal N. and al., 2004 [5], swells are the most important phenomenon to consider among the environmental conditions affecting maritime structures, because they exert the greatest influence. Farah Zemani, 2014 [4], the presence of waves makes the design process for maritime structures on land. Since waves are one of the most complex and variable phenomena in nature, it is not easy to achieve a full understanding of their fundamental character and behavior. Waves are oscillations of the surface of the sea, generated by the energy of the wind and maintained by gravity [6]; [7]. Under the effect of an external disturbance (wind, seabed, etc.), the profile of a swell can be modified. It can be subject to movements, such as: shoaling, breaking, refraction, diffraction, reflection… [8];[9]. Study the evolution of the main physical parameters and dimensionless numbers to characterize the dynamics of swells during their propagation in coastal areas is essential to understand the physics of the phenomenon. Understanding the mechanism of generation and the physics that accompanies the propagation of waves are of great interest to the scientific community. Several experimental, theoretical and numerical studies have been carried out for years with the aim of predicting such a phenomenon and subsequently understanding the risks that not only threaten coastal infrastructure but also human life [10]. In a given wave record, each free surface elevation can be considered as a random variable. The studies carried out in the laboratory form the most reliable method compared to numerical simulations which are less effective given the difficulty in perfectly describing the physics which accompanies the breaking **[**11**]**;[12] ; [13].Many studies on wave dynamics have certainly already been carried out, but mainly in deep or intermediate waters [14] ;[15]**.**In this study we are interested in events occurring near the coast, in this case adult waves (swells) and its orbital marine currents generated by them.

What is great in this work, using the breaking criteria of Sunamura 1980 or Kaminsky and Krauss 1993 which take into account the slope of the seabed ($tan\beta$), the evaluation of the local water height d_b and the height H_b of the swell at the point of breaking, the follow-up of the energy potentials of the swell and the orbital marine currents generated by its last, reveals that these swells break very close to the coast and the brutal dissipation of their energies constitutes the principal cause of the erosion of the littoral. The installation of a converter in the shoaling zone would have a double advantage for this purpose: the maximum recovery of energy and the attenuation of the brutal dissipation of energy which will strongly contribute to the smooth running of the preservation of the coastal strip.

Materials and Methods:-

Presentations of the study site

Benin is a Gulf of Guinea country situated between the parallels 6°15' and 12°30' of North latitude on the one hand and the meridians 1° and 3°40' East longitude on the other hand (**Fig. 1a**). It benefits from access to the ocean over a distance of approximately 125 km and goes from Hilla-Condji in the West to Kraké in the East. The coastline of Benin is more or less linear and interrupted in two places, namely the Bouche du Roy and the mouth of the Cotonou Canal. In this coastal zone, it receives an abundance of regular sea swell of low amplitude compared to their wavelength [1]. Their amplitude varies according to the period of the year on the one hand and on a daily basis on the other hand. As part of the extension of the Autonomous Port of Cotonou, the swell measurements carried out (**Fig. 1b**) over four consecutive years (from June 2011 to April 2014) in five-minute steps by the Millennium Challenge Account (MCA-Benin) and the bathymetric map of Benin obtained (**Fig. 2**) near Institute of Fishery and

Oceanological Research of Benin (IRHOB)/Benin Center for Scientific Research and Innovation (CBRSI), have made it possible to define the average stable daily, monthly and annual wave data for this site [2].

1. In deep water, the regular mean height of the swell is $H_o = 2a_o = 0{,}8m$, its mean period is $T = 12s$ and its wavelength $L_o = 200m$ at a position where the local water depth is $d_o = 130m$ about.
2. Experimental measurements have shown that these swells break at the point of local depth d_b such that $4m \leq d_b \leq 5m$ and that their period oscillates between $9s$ and $15s$.
3. In the coastal zone where the local water depth is $d = 50m$, the crest-to-trough wave height is $d = 2a = 0{,}95m$.
4. The gravity on the Beninese coast is approximately $g = 9.79 m.s^{-2}$.
5. The seabed in the coastal zone is almost flat and sloping. It is a low slope bottom $p = tan\,\beta$ such as $0{,}001 < tan\,\beta \leq 0{,}1 \Rightarrow tan\,\beta \approx \beta$. The average of this slope in the shoaling zone is $\beta_m \approx \frac{100}{2000} = 0{,}05$ (**Fig. 2**).
6. These values are almost confirmed by the Méduse-Benin satellite forecasting station, which indicates values between $8s$ and $16s$.

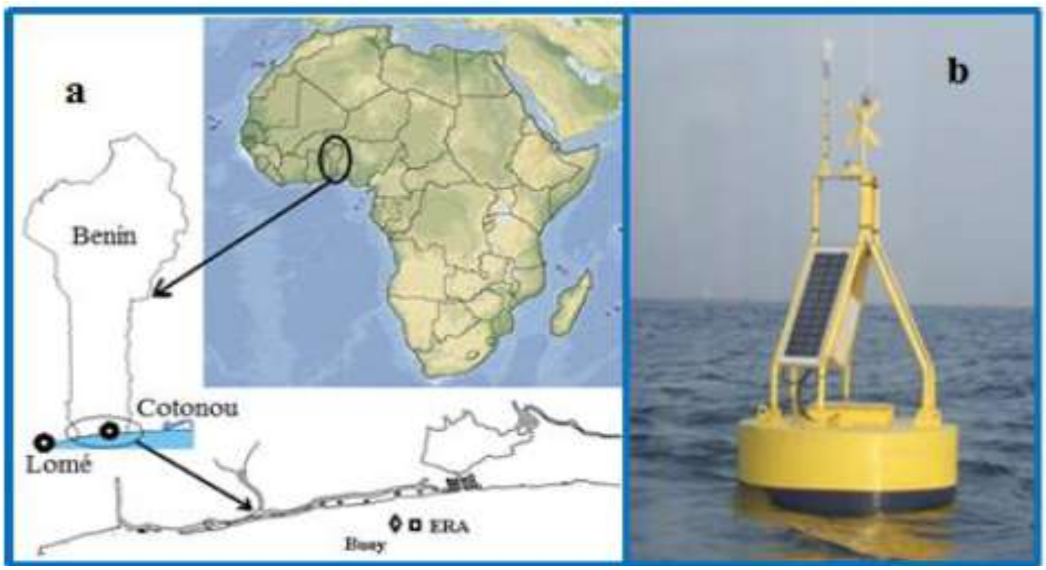

Figure 1:- Sketch of study area (Benin, Gulf of Guinea).
Location of ERA and (b) oceanographic buoy.

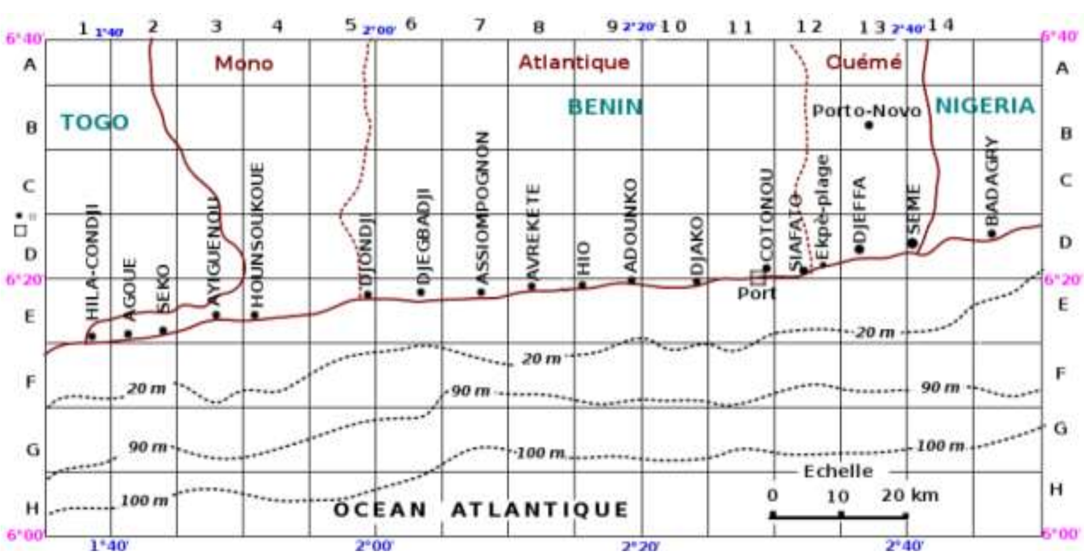

Figure 2:- Bathymetric map of Benin.

Choice of model and criteria of breaking

In the coastal zone, for an almost flat seabed, the propagation of a swell in dimension 1 is modeled by the system of equations below: [16] ;[6].

$$\begin{cases} \frac{\partial \eta}{\partial t} + \frac{\partial (hu)}{\partial x} = 0 \\ \frac{\partial u}{\partial t} + \varepsilon u \frac{\partial u}{\partial x} + g \frac{\partial \eta}{\partial x} = \alpha \mathcal{D} + \mathcal{O}(\alpha^2) \end{cases} \text{with} \begin{cases} h = \eta + d \\ u = u(x,t) \\ \eta = \eta(x,t) \\ d = d(x) \end{cases} \textbf{(1)}$$

- $\mathcal{D} = 0$, we have the equations of « shallow water », obtained from the theories of Airy or Stokes and which makes it possible to characterize the swells for a constant bathymetry.
- $\mathcal{D} = \frac{h}{2}\frac{\partial^2}{\partial x^2}\left(h\frac{\partial u}{\partial t}\right) - \frac{h^2}{6}\frac{\partial^3 u}{\partial x^2 \partial t}$; we are talking about Boussinesq's theory proposed by Peregrine in 1967. It takes into account the low non-linearity, the low dispersion of long swells of low amplitude which propagate over a variable bathymetry.
- $\mathcal{D} = \frac{1}{3h}\frac{\partial}{\partial x}\left[h^3\left(\frac{\partial^3 u}{\partial x \partial t} + \varepsilon u \frac{\partial^2 u}{\partial x^2} + \varepsilon\left(\frac{\partial u}{\partial x}\right)^2\right)\right]$; these are the equations of Serreor of Green-Naghdi[17]. Taking into account the dispersive and non-linear effects, they almost generalize the equations of « shallow water » and of Peregrine (Boussinesq) with a strong dominance of nonlinear effects. Non-linearity parameters ε , from the dispersive effect α , the camber S of the swell and the wave train μ are defined by:

$$\varepsilon = \frac{a}{d} \quad ; \quad \alpha = \mu^2 = \left(\frac{d}{L_o}\right)^2 \text{and} S = \frac{\varepsilon}{\alpha} = \frac{\varepsilon}{\mu^2} = \frac{aL_o^2}{d^3} \text{with} a = \frac{H}{2} \textbf{(2)}$$

When $\varepsilon \ll 1$ and $\alpha \ll 1$ with $S = \mathcal{O}(1)$, the swell is linear: it is called Stokes or Airy and we have:

$$\eta \ll d \Rightarrow h = \eta + d \approx d \quad \textbf{(3)}$$

For a swell, of wavelength L, three propagation zones are defined: deep waters when $d > \frac{L}{2}$; the zone of shoaling if $\frac{L}{25} \leq d \leq \frac{L}{2}$ and Surf and Swash zones for $0 \leq d \leq \frac{L}{25}$ [18].

Table 1:- Values of some parameters in the study zone [17].

Study sites	places of propagation	Non-linearity parameter	Dispersive effect parameter	Camber of the swell
Gulf of Guinea (Benin)	Deep Waters ($d_o = 130m$)	$\varepsilon_o = 0{,}003$	$\alpha_o = 0{,}423$	$S_o = 0{,}07 \approx \mathcal{O}(1)$
	coastal zone ($d_c = 50m$)	$\varepsilon_c = 0{,}0097$	$\alpha_c = 0{,}0625$	$S_c = 0{,}155 \neq \mathcal{O}(1)$

- $\varepsilon_o \ll 1$; $\alpha_o < 1$ and $S_o = \mathcal{O}(1)$: the swells in the Gulf of Guinea in Benin are Airy or Stokes swells.
- $\varepsilon_c \ll 1$ and $\varepsilon_c > \varepsilon_o$: the non-linear effects, even if they remain negligible, increase in the coastal zone.
- $\alpha_c \ll 1$ and $\alpha_c < \alpha_o$: the dispersive effect decreases in the coastal zone and $\alpha \to \mathcal{O}(1)$ when moving towards the beach.
- $S_c > S_o$ and $S_c \neq \mathcal{O}(1)$: the Airy or Stokes theory does not allow a better characterization of these swells in the coastal zone of this site.

All in all, the waves in the Gulf of Guinea at Cotonou are Stokes swells ($\eta \ll d$), weakly non-linear ($\varepsilon \ll 1$) and weakly dispersive ($\alpha < 1$). The use of the Boussinesq (Peregrine) equations, in the following, will make it possible to better characterize the variations in their height and their wavelength during their propagation in the coastal zone.

Kaminsky and Kraus, 1993 conducted a comparative study based on an analysis of previous work (409 cases) covering a wide range of wave steepness and beach slopes, propose to use for a slope $tan\,\beta \leq 0{,}1$, the index of breaking below which takes into account the number of Irraben[19].

$$\gamma = \frac{H_b}{d_b} = 1{,}2\left(tan\,\beta\sqrt{\frac{L_o}{H_o}}\right)^{0{,}27} \quad \textbf{(4)}$$

At the study site, $0{,}001 < tan\,\beta \leq 0{,}1$, this criteria is chosen to determine the place of breaking of these swells.

Parameter variations from offshore to breaking point

The Airy or Stokes swell dispersion relationship is [2]:

$$\omega^2 = \frac{4\pi^2}{T^2} = gk\,tanh(kd) \textbf{ (5)}$$

Expressions that translate the variations of the period T, phase velocity C_φ and group velocity C_g are :

$$\begin{cases} \boldsymbol{T} = \frac{2\pi}{\sqrt{gk\,tanh(kd)}} \\ \boldsymbol{C_\varphi} = \sqrt{\frac{g}{k}tanh(kd)} \\ \boldsymbol{C_g} = \frac{1}{2}\left(1 + \frac{2kd}{sinh(2kd)}\right)\sqrt{\frac{g}{k}tanh(kd)} \end{cases} \text{with} \begin{cases} L_o = \frac{g{T_o}^2}{2\pi} \\ C_{\varphi_o} = \frac{gT}{2\pi} \\ C_{g_o} = \frac{gT}{4\pi} \end{cases} \textbf{(6)}$$

In the deep waters $\left(d \geq \frac{L_o}{2} \Rightarrow \mu \geq \frac{1}{2}\right)$, $tanh(kd) \approx 1$, then all the parameters are constant in this propagation zone. Taking into account the previous conditions ($\eta \ll d$, $\varepsilon \ll 1$ $and\alpha < 1$), the Peregrine equations (Boussinesq) which correspond to the study sites are [17]:

$$\begin{cases} \frac{\partial\eta}{\partial t} + \frac{\partial(du)}{\partial x} = 0 \\ \frac{\partial u}{\partial t} + g\frac{\partial\eta}{\partial x} = \alpha\left[\frac{d}{2}\frac{\partial^2}{\partial x^2}\left(d\frac{\partial u}{\partial t}\right) - \frac{d^2}{6}\frac{\partial^3 u}{\partial x^2\partial t}\right] \end{cases} \Rightarrow \begin{cases} \frac{\partial\eta}{\partial t} + \frac{\partial(du)}{\partial x} = 0 & (i) \\ \frac{\partial u}{\partial t} + g\frac{\partial\eta}{\partial x} = \alpha\frac{d^2}{3}\frac{\partial^3 u}{\partial x^2\partial t} & (ii) \end{cases} \textbf{(7)}$$

By replacing (ii) in the derivative of (i) with respect to time t, we obtain:

$$\frac{\partial^2\eta}{\partial t^2} + \frac{\partial}{\partial x}\left[-g\mu L_o\frac{\partial\eta}{\partial x} + \mu^5\frac{{L_o}^3}{3}\frac{\partial^4 u}{\partial x^2\partial t^2}\right] = 0 \text{with} \begin{cases} \eta = \eta(x,t) = \eta_a e^{-i\omega t} \\ \eta_a = \frac{H(x)}{2}e^{ikx} \\ d = -\beta x = \mu L_o \\ \alpha = \mu^2 \end{cases} \textbf{(8)}$$

At the limit of deep waters and the shoaling zone where$\mu = \frac{1}{2}$, we have :

$$\boldsymbol{H\left(\mu = \frac{1}{2}\right) \approx H_o} \text{and} \boldsymbol{L\left(\mu = \frac{1}{2}\right) \approx L_o} \quad \textbf{(9)}$$

The equation of the tangent to the seabed is $d = -x\,tan\,\beta = \mu L_o$[20] and [21], so on the study site, we have:

$$\boldsymbol{tan\,\beta \approx \beta \Rightarrow d = -\beta x = \mu L_o} \quad \textbf{(10)}$$

By making the change of variable $\mu = -\frac{\beta}{L_o}x$and using the approximation $\mu^5 = \sqrt{\alpha^5} = \mathcal{O}(1)$, the previous equation becomes:

$$\boldsymbol{\frac{\partial^2\eta_a}{\partial\mu^2} + \frac{1}{\mu}\frac{\partial\eta_a}{\partial\mu} + \frac{2\pi}{\beta^2\mu}\eta_a = 0} \quad \textbf{(11)}$$

The general solution of this equation is$\eta_a = AY_0(z)$, where$z = \frac{2}{\beta}\sqrt{2\pi\mu}$ and $Y_0(z)$is the modified Bessel function of second species and whose complex asymptotic formula is $Y_0(z) = \sqrt{\frac{2}{\pi z}}e^{-i\left(z-\frac{\pi}{4}\right)}$. From the preceding boundary conditions, we obtain.

$$\begin{cases} \boldsymbol{H = H_o\left(\frac{1}{2\mu}\right)^{1/4} = H_o\left(\frac{2d}{L_o}\right)^{-1/4}} \\ \boldsymbol{L = L_o(2\mu)^{1/2} = L_o\left(\frac{2d}{L_o}\right)^{1/2}} \end{cases} \Rightarrow \quad \boldsymbol{kd = \pi\sqrt{2\mu}} \textbf{(12)}$$

According this theory, which takes into account the variability of the bathymetry, the different coefficients of variation of the parameters previously studied in the shoaling zone $\left(\mu_b \leq \mu \leq \frac{1}{2}\right)$, are :

$$\begin{cases} \boldsymbol{\delta = \frac{L}{L_O} = \sqrt{2\mu}} \\ \boldsymbol{K_s = \frac{H}{H_o} = \left(\frac{1}{2\mu}\right)^{\frac{1}{4}}} \\ \boldsymbol{\tau = \frac{T}{T_o} = \sqrt{\frac{\sqrt{2\mu}}{tanh(\pi\sqrt{2\mu})}}}\,(13) \\ \boldsymbol{\delta_1 = \frac{C_\varphi}{C_{\varphi_o}} = \sqrt{\sqrt{2\mu}tanh(\pi\sqrt{2\mu})}} \\ \boldsymbol{\delta_2 = \frac{C_g}{C_{g_o}} = \left(1 + \frac{2\pi\sqrt{2\mu}}{sinh(2\pi\sqrt{2\mu})}\right)\sqrt{\sqrt{2\mu}tanh(\pi\sqrt{2\mu})}} \end{cases} \text{with} \begin{cases} \boldsymbol{L_o = \frac{gT_o^2}{2\pi}} \\ \boldsymbol{C_{\varphi_o} = \frac{gT_o}{2\pi}} \\ \boldsymbol{C_{g_o} = \frac{gT_o}{4\pi}} \end{cases} \textbf{(13)}$$

Determining the breaking point of the swells:

Using the results obtained with the Boussinesq theory and the breaking criterion of Kaminsky and Krauss 1993, the local water depth $d_b$and the wave height$H_b$at the breaking point are:

$$\begin{cases} d_b = \frac{H_o^{0,908} L_o^{0,092}}{1,33(tan\beta)^{0,216}} \\ H_b = 0,903 H_o \left(\frac{L_o}{H_o}\right)^{0,227} (tan\beta)^{0,054} \end{cases} \Rightarrow \mu_b = \frac{d_b}{L_o} = \frac{H_o^{0,908} L_o^{-0,918}}{1,33(tan\beta)^{0,216}} \quad (14)$$

Potential energetic orbital marine currents generated by the swell

The movement of the waves has a maximum speed at the surface of the sea. This speed attenuates towards the bottom so that the waves do not possess no significant influence at depths $d = \frac{L}{2}$[22]. The equations that parameterize the position of a water particle struck by the swell in deep water is [23];[24]:

$$\overrightarrow{OM} \begin{cases} X = \frac{H}{2} \frac{cosh(kz+\pi)}{cosh\pi} sin(\theta) \\ Z = \frac{H}{2} \frac{sinh(kz+\pi)}{cosh\pi} cos(\theta) \end{cases} \text{with} \quad \theta = \frac{2\pi x}{L} - \frac{2\pi t}{T} (15)$$

The trajectory of the particles is so circular in infinite depth (offshore) and elliptical shape more and more crushed as the bottom rises. Waves induce orbital motion in the water and the amplitude of this motion decreases with depth. Equation (16) describes the behavior of the horizontal velocity V_x and the vertical velocity V_z induced by the swell [9].

$$\vec{V} \begin{cases} V_x = \frac{gHT}{2L} \frac{cosh(kz+\pi)}{cosh\pi} cos(\theta) \\ V_z = \frac{gHT}{2L} \frac{sinh(kz+\pi)}{cosh\pi} sin(\theta) \end{cases} \text{with } \theta = \frac{2\pi x}{L} - \frac{2\pi t}{T} (16)$$

And $V^2 = \frac{g^2H^2T^2}{4L^2cosh^2\pi}[cosh^2(kz+\pi)cos^2\theta + sinh^2(kz+\pi)sin^2\theta]$

The equations of system (17) generally parameterize the ellipses with horizontal major axis D' and minor vertical axis d' such that:

$$\begin{cases} D' = H \frac{cosh(kz+\pi)}{cosh(\pi)} \\ d' = H \frac{sinh(kz+\pi)}{cosh(\pi)} \end{cases} (17)$$

Thus, the amplitudes of the velocity v_g dof propagation of water particles put into motion by the swell (that of the orbital marine currents generated by the swell) are as follows:

$$\vec{v}_g = \begin{cases} \vec{v}_{gx} = \frac{D'}{2}\omega = \frac{H\omega}{2} \frac{cosh(kz+\pi)}{cosh(\pi)} \\ \vec{v}_{gz} = \frac{d'}{2}\omega = \frac{H\omega}{2} \frac{sinh(kz+\pi)}{cosh(\pi)} \end{cases}$$

$$\Rightarrow v_g = \sqrt{[cosh^2(kz+\pi) + sinh^2(kz+\pi)]} \quad (18)$$

On the seabed ($z = -d$), this velocity has the expression:

$$v_g = \frac{2\pi H}{2Tcosh(\pi)} \sqrt{[cosh^2(\pi - kd) + sinh^2(\pi - kd)]} (19)$$

With $H(d) = \begin{cases} H_0 \text{ if } d > \frac{L}{2} \\ H_0 \sqrt{\frac{cos\theta_0}{cos\theta}} \left(\frac{kd}{cosh^2(kd)} + tanh(kd)\right)^{\frac{-1}{2}} \text{ if } d_b \le d \le \frac{L}{2} \\ H_0 \left[\sigma\left(\frac{d}{d_b}\right)^{\frac{-1}{2}} + (1-\sigma)\left(\frac{d}{d_b}\right)^{\frac{1}{4}}\right]^{-1} \text{ if } 0 \le d \le d_b \end{cases}$ and $\sigma = \frac{2H_0}{Ttan\beta\sqrt{gd_b}}$

Energetic power of the swells and its orbital marine currents

- The average of the total energy E of one swell of height H per unit length is:

$$< E > = \frac{1}{8}\rho g H^2 \quad (20)$$

The energy power P per unit length of a wave is the product of the average of its total energy and its group velocity [1].

$$P = < E > \cdot C_g \quad (21)$$

In deep waters, we have:

$$P = < E > . C_{g_o} = \frac{1}{8}\rho g H_o^2 C_{g_o} = \frac{1}{32\pi}\rho g^2 H_o^2 T_o \quad (22)$$

In the shoaling zone, we have:

$$P =< E > \cdot C_g = \frac{1}{8}\rho g H_0^2 C_{g_o}\left(\frac{1}{2\mu}\right)^{1/2}\left(1+\frac{2\pi\sqrt{2\mu}}{sinh(2\pi\sqrt{2\mu})}\right)\sqrt{\sqrt{2\mu}tanh(\pi\sqrt{2\mu})} \quad (23)$$

$$\begin{cases} \psi = \frac{E}{E_o} = \left(\frac{1}{2\mu}\right)^{1/2} \\ \chi = \frac{P}{P_o} = \left(\frac{1}{2\mu}\right)^{1/2}\left(1+\frac{2\pi\sqrt{2\mu}}{sinh(2\pi\sqrt{2\mu})}\right)\sqrt{\sqrt{2\mu}tanh(\pi\sqrt{2\mu})} \end{cases} \quad (24)$$

- The average kinetic energy, at an altitude z, of the orbital currents generated by the offshore swell is [24]:

$$< E'_c >= \frac{1}{L}\int_x^{L+x} dx \int_{-L/2}^{z} \frac{1}{2}(\tilde{\rho}+\rho_0)V^2 dz = \frac{1}{16}\rho_0 g H^2\left(1+\frac{2z}{L}\right) \quad (25)$$

By taking the reference of the potential energy of gravity on the bottom, (at the depth $z = -d$) where the effect of the swell disappears, the average potential energy of the orbital marine currents is:

$$< E'_p >= \frac{1}{L}\int_x^{L+x} dx \int_{-d}^{z} \frac{1}{2}(\tilde{\rho}+\rho_0) g dz = \rho_0 g(z+d) \quad (26)$$

The average total energy of orbital sea currents generated by the swell is:

$$< E_t > =< E'_c > + < E'_p >= \frac{1}{16}\rho_0 g H^2\left(1+\frac{2z}{L}\right) + \rho_0 g(z+d) \quad (27)$$

- **Energetic potential of orbital marine currents:**

According to Airy's Theory [17] ; [25] the average energy power P of the internal waves is expressed as $P = v_g < E_t >$ where v_g is the group velocity following the propagation of internal waves.

- At an altitude z, this potential is given by:

$$P(z) = \frac{\pi g \rho_0}{16T cosh(\pi)}\left[\left(1+\frac{2z}{L}\right)H^3 + 16(z+d)H\right]\sqrt{[cosh^2(kz+\pi) + sinh^2(kz+\pi)\ ]} \quad (28)$$

- On the seabed where $z = -d$, this potential has the expression:

$$P(-d) = \frac{\rho_{0g}\pi}{16T cosh(\pi)}\left[\left(1+\frac{2z}{L}\right)H^3\right]\sqrt{[cosh^2(\pi-kd) + sinh^2(\pi-kd)\ ]} \quad (29)$$

$$H(d) = \begin{cases} H_0 if d > \frac{L}{2} \\ H_0\sqrt{\frac{cos\theta_0}{cos\theta}}\left(\frac{kd}{cos h^2(kd)} + tanh(kd)\right)^{\frac{-1}{2}} if d_b \le d \le \frac{L}{2} \text{and} \sigma = \frac{2H_0}{Ttan\beta\sqrt{gd_b}} \\ H_0\left[\sigma(\frac{d}{d_b})^{\frac{-1}{2}} + (1-\sigma)(\frac{d}{d_b})^{\frac{1}{4}}\right]^{-1} if \quad 0 \le d \le d_b \end{cases}$$

Results, Analysis And Discussion:-

Presentation of the results

The measurements were taken in five-minute steps over six months from June to December. We defined a typical day for each of the months when the measurements were complete. The curves in **fig. 3** below reflect the variation in height crest-to-trough swells according to time (GMT) during these typical days.

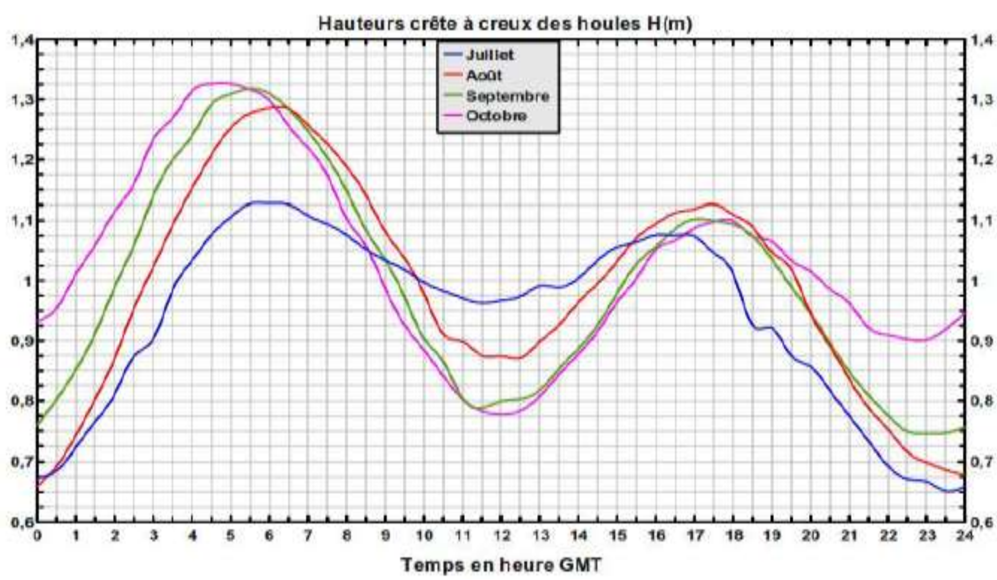

Fig 3:- Height of the swells during each typical day of the month.

➢ The curves in solid lines represent the results obtained with the Boussinesq (Peregrine) equations, while those in dotted lines correspond to the prediction of the Airy or Stokes theory. These curves reflect the variations of the various parameters studied, offshore to the breaking point. The curves in **fig. 4** present the variations coefficients in wavelength, height, period, phase velocity and group velocity of swells in Benin (Gulf of Guinea).

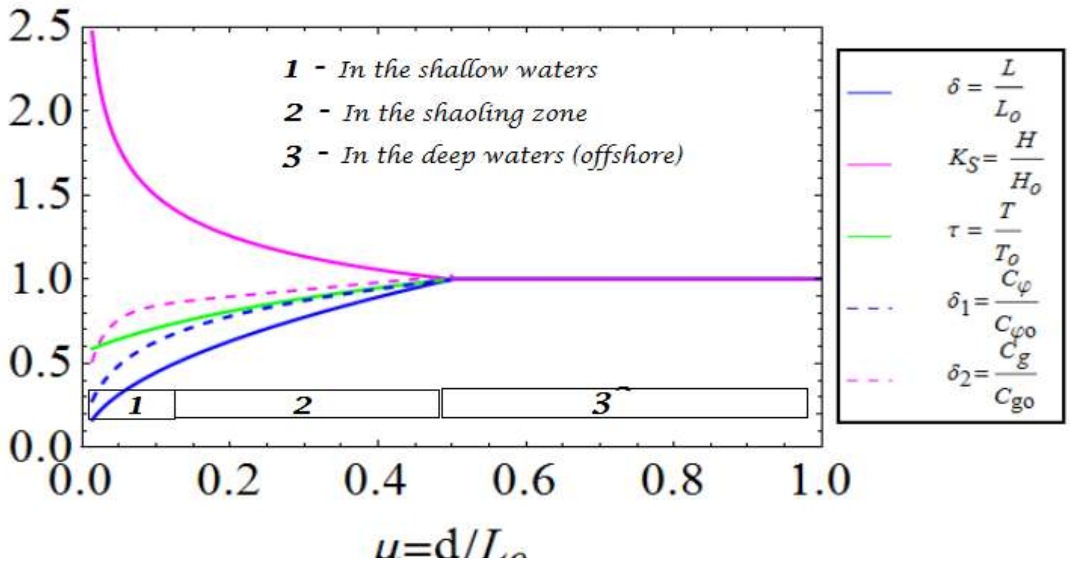

Figure 4:- Variation coefficients of characteristic parameters of the swell in Benin (Gulf of Guinea).

➢ The curves in **Fig. 5** below show the variations in the total energy ψ and the energy power χ of the swells per unit length during their propagation.

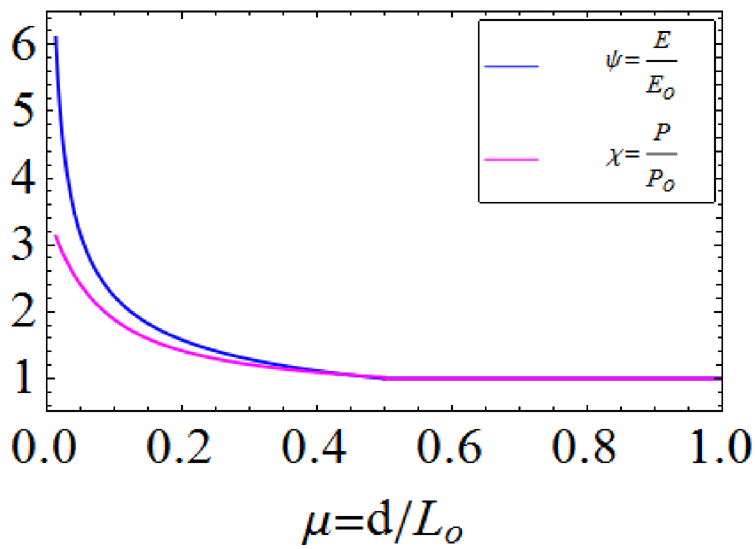

Figure 5:- Variations coefficients $\psi$of the total energy and $\chi$of the swell energy power.

➢ The **fig. 6**, **6a** and **6b**, show the evolution of the vertical elevation of the sea surface as a function of the local water depth μ and the slope of the seabed β. The curve in **fig. 6** is a representation in dimension 2 (2D) whereas those of **fig. 6a** and **6b** are its representations in dimension 3 (3D).

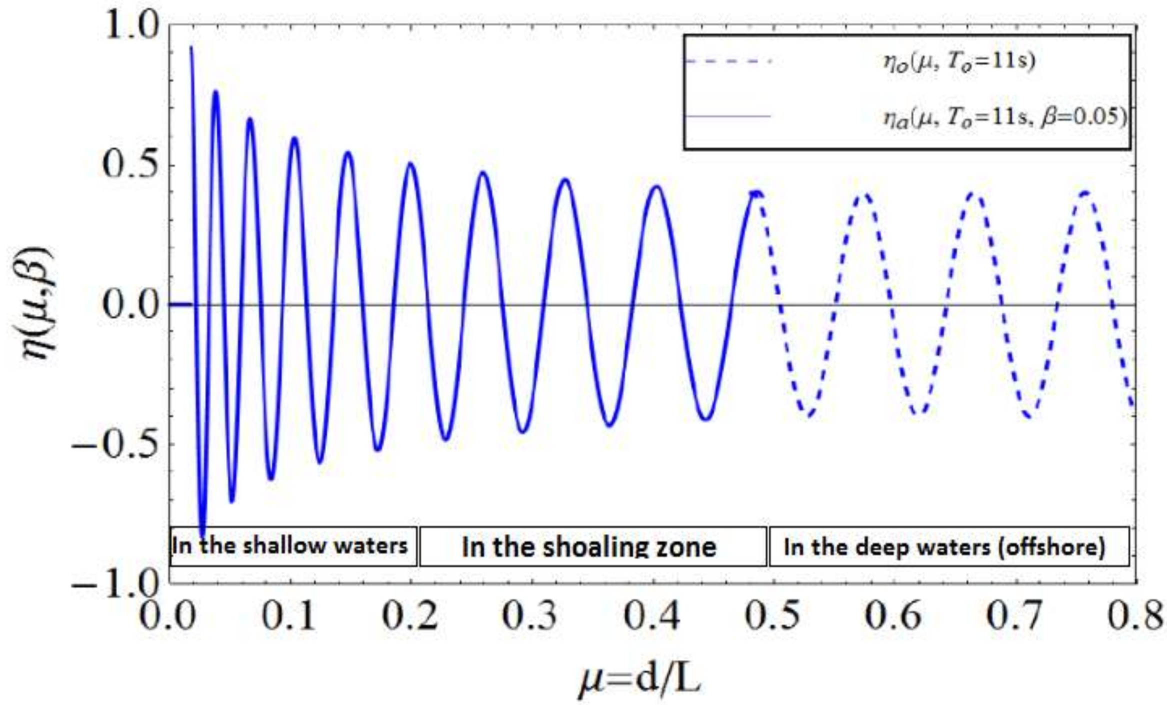

Figure 6:- Vertical elevation of the sea surface in deep water $\eta_0(\mu)$and in the shoaling zone$\eta_a(\mu,\beta) indimensiontwoofthelocalwaterdepth\mu$.

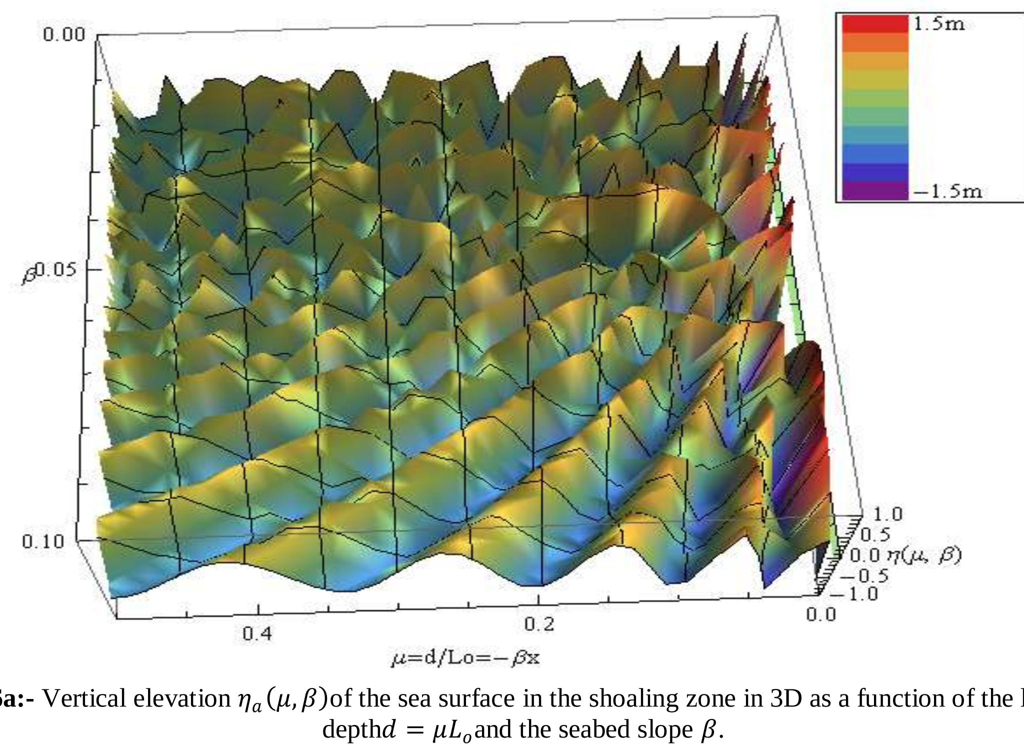

Figure 6a:- Vertical elevation $\eta_a(\mu,\beta)$of the sea surface in the shoaling zone in 3D as a function of the local water depth$d = \mu L_o$and the seabed slope β.

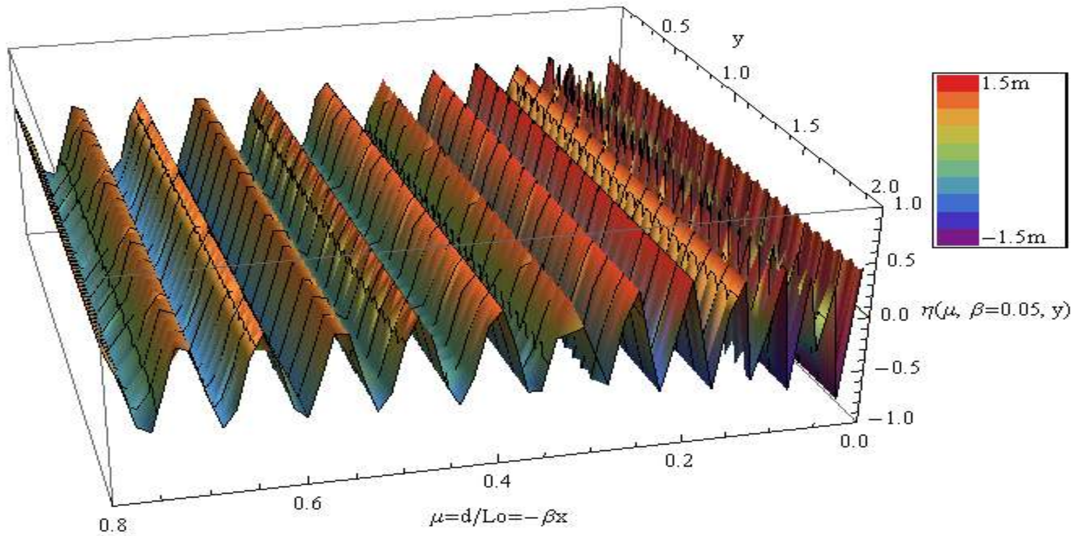

Figure 6b:- Vertical elevation of the marine surface offshore$\eta_o(\mu)$and in the shoaling zone $\eta_a(\mu,\ \beta = 0{,}05)$in 3D of the local water depthμ.

➢ The curves in **Fig. 7** and **8** below show the variations in the local water depth d_b at the breaking point as a function of the seabed slope β according to the criteria of Sunamura and Kaminsky, 1980 and Krauss, 1993 which take account of the slope of the seabed β on the one hand and that of the height of the swells H_b at this point on the other hand.

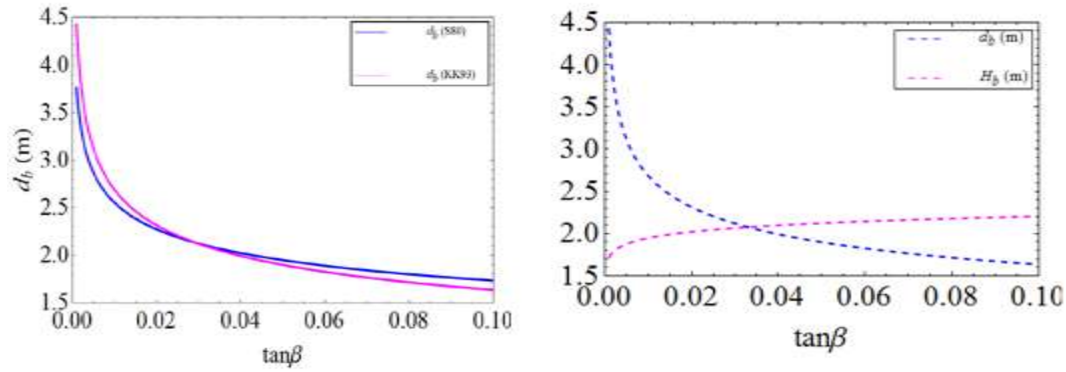

Figure 7:- Variations of the local water depth $d_b$at the breaking point as a function of the slope according to the criteria of Sunamura, 1980; Kaminsky and Krauss, 1993.

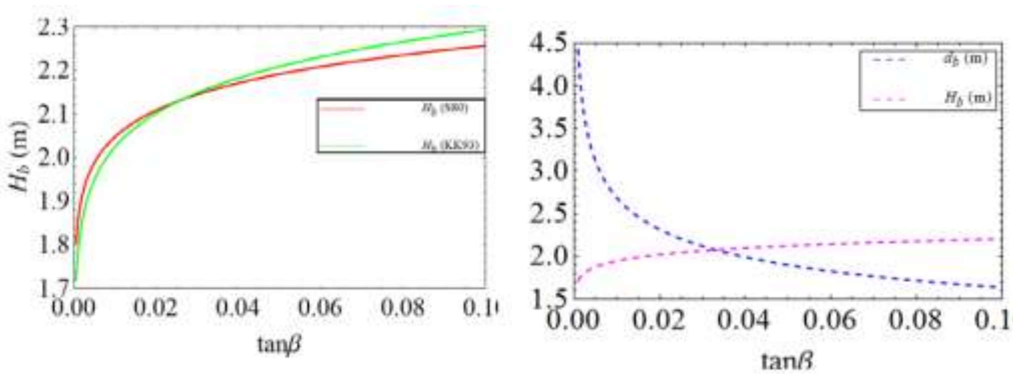

Figure 8:- Variations of the swell height$H_b$at the breaking point as a function of the seabed slope β according to the criteria of Sunamura, 1980; Kaminsky and Krauss, 1993.

➢ From the values of the heights measured by the Millennium Challenge Account in Benin (At the Autonomous Port of Cotonou), we defined the average values of these heights which made it possible to obtain the curves below. These curves in **fig. 9** translate the energetic disturbances of the orbital marine currents generated by the swell on the seabed in the coastal zones, before and after the breaking. **Fig. 10** is the zoom of figure 9 around the bathymetric breaking.

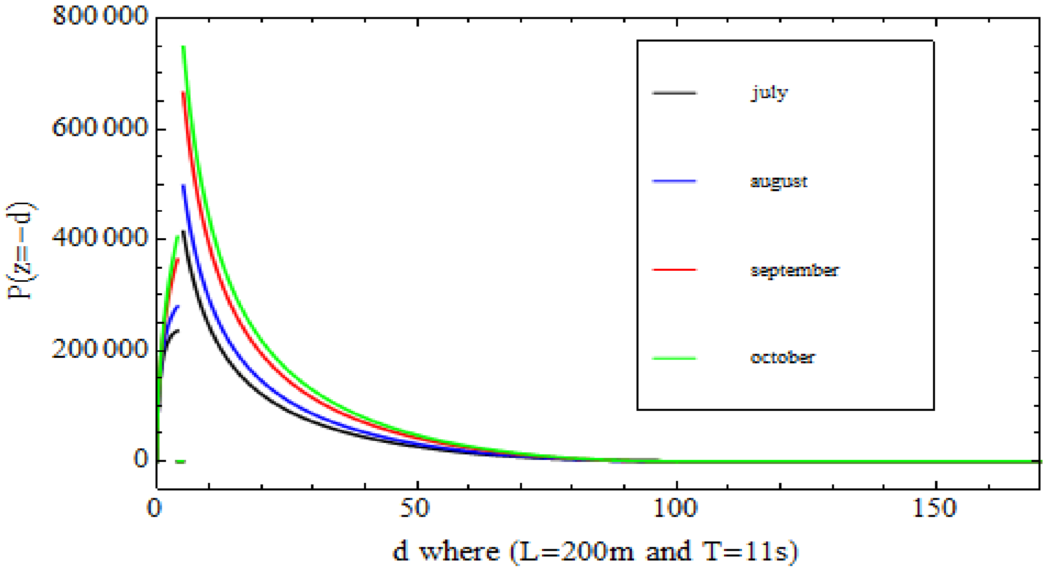

Figure 9:-Average energy potential of orbital marine currents on the seabed for each typical day of the months.

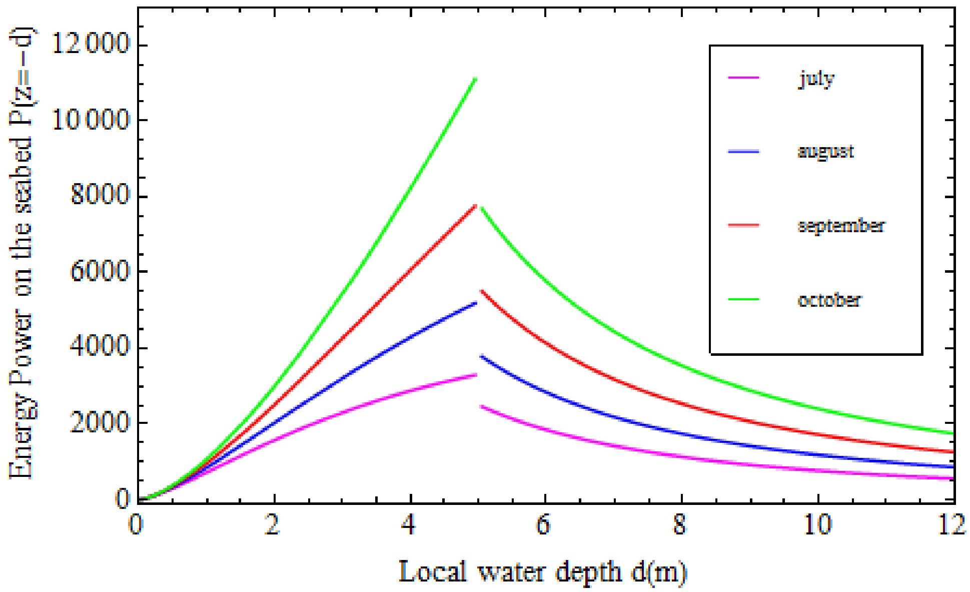

Figure 10:-Zoom of figure 9 around the breaking point.

➢ As for the curves in **Fig. 11** and **12**, they reflect the variations in the energy potential of these orbital marine currents at given local water depths.

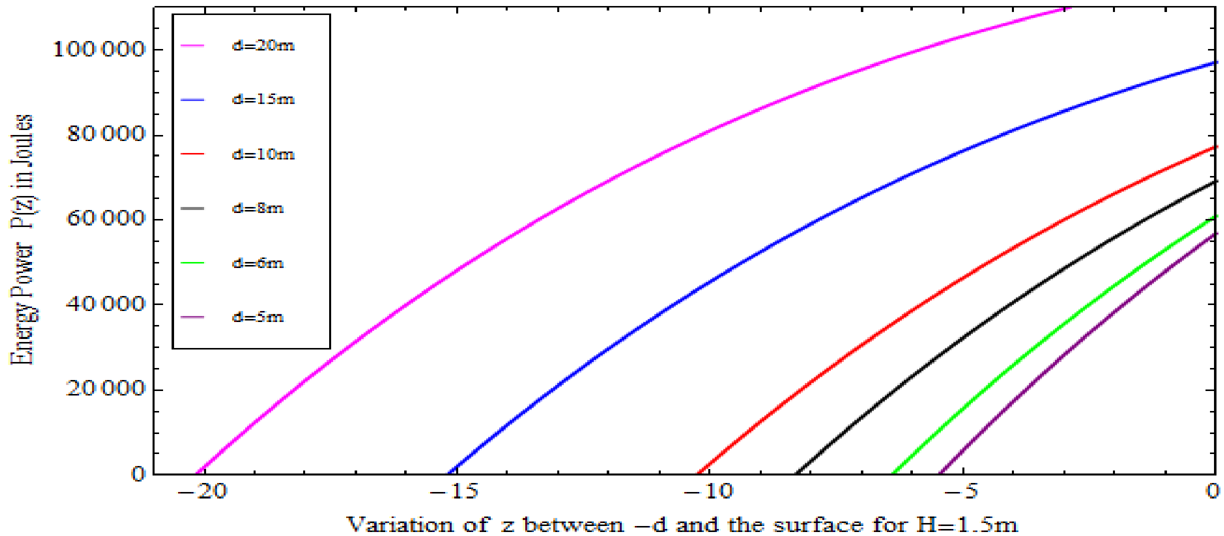

Figure 11:-Variations in the energy potential of orbital marine currents in the Shoaling zone for H=1,2m and T=12s.

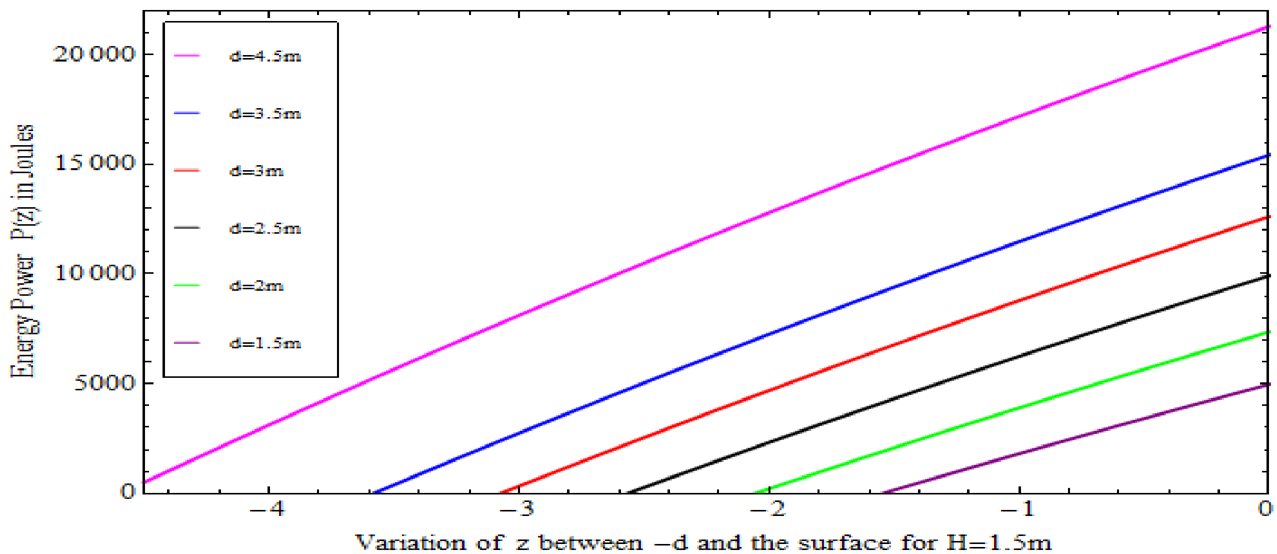

Figure 12:-Variations in the energy potential of orbital marine currents in the breaking zone: Surf and Swash zones for H=1,2m and T=12s.

Analysis and Discussion of the results:-

- The curves of **Fig. 3** show that the swells in Benin (during the four months of full campaign) are almost regular. Their heights vary almost sinusoidal and have two different maximum values and two different minimum values. These maximum values are observed around 05 hours and 17 hours GMT, those minimum around 00 hours and 12 hours. These heights vary between 0.6m and 1.4m.
- All the curves, in **Fig. 4**, show that all the parameters of the swells in the deep waters, apart from any obstacle, are constant. The curves$\kappa_s = \frac{H}{H_o}$ and $\tau = \frac{T}{T_0}$of **Fig. 4** represents the variations respectively of the height and the period in the deep waters and in the Shoaling zone. They are constant in deep waters ($\kappa_s \approx 1\ and \tau \approx 1$) and believe strictly in the shoaling zone ($1 \leq \kappa_s \leq 3$) and this amplification stops at the breaking point. The crest-to-trough height of the swell is therefore constant in deep waters (offshore) and strictly increasing in the shoaling zone. This height, which increases strictly when the local water depth decreases, remains proportional to$\left(d^{-1/4}\right)$; for $d_c = 50m$, $H_c = 0{,}97m$; $\mu_c = 0{,}25$ and $\kappa_{s_c} = \frac{H}{H_o} = 1{,}19$ or the curve of$\kappa_s = \frac{H}{H_o}$give for $\mu = 0{,}25$ and $\kappa_s = \frac{H}{H_o} = 1{,}21$.

From these results, we have $\kappa_{s_c} \approx \kappa_s$: this result shows that it is Peregrine's theory (Boussinesq) which makes it possible to better model the variations of the swell in the shoaling zone of the studied coastal zone.
The curve δ of the **fig. 4**, decreases in the shoaling zone. But the decrease is more accentuated in the case of the results obtained with the equations of Boussinseq. In the coastal area of the Gulf of Guinea at Cotonou, the wavelength of the swells decreases when the local water depth decreases and remains proportional to the square root of the latter. $(L \sim \sqrt{d})$.
The evolution of the curve δ_2 of the **fig. 4**, shows that the group velocity increases when μ varies from 0,5 at 0,125 before starting to decrease to the breaking point.
The variations of δ_1 show that the phase velocity decreases according to both theories. But this decrease is more accentuated in the case of Boussinesq's theory. The phase velocities C_φ and group C_g of the swells in the shoaling zone in Cotonou, decrease when the local water depth decreases. But in this zone, the velocity of group is always higher than that of phase.

- The evolution of the curves of the **fig. 5**, shows that the energy potential in deep waters varies. In the shoaling zone, this power varies exponentially with a very high average at a position where $d = 8m$. In the breaking zone where $d = 3m$, this potential also varies with a non-negligible average. The shoaling zone is therefore a zone of strong amplification of the energy power of the swell, while those of Surf and Swash are the zones of energy dissipation. These results show that the swells in Benin are more energetic in the shoaling zone. The curves of **fig. 6**, **6a** and **6b**, which translate the variations of the vertical elevation of the free surface of the ocean according to the local water depth. Note that the curve of **Fig. 6** is a representation in dimension 2 (2D) while those of **Fig. 6a** and **6b** are its representations in dimension 3 (3D). They confirm:

- The constancy of the various offshore parameters.
- Height amplification crest to trough of the swells in the shoaling zone under the disturbing effect of the seabed.

➢ The decrease or contraction of the wavelength caused by the slope of this seabed. The curve of **fig. 6b**, which represents the variations of the vertical elevation of the free surface of the ocean according to the local water depth $d = \mu L_o$ and the slope of the seabed β in the shoaling zone, it reveals that the randomness accompanied by small oscillations on the surface of the swell is due to the variability of the slope of the seabed. The curves in **fig. 7**, obtained from the results of Boussinesq's theory and the breaking criteria of Sunamura, 1980; Kaminsky and Krauss, 1993 which translate the evolution of the local water depth d_b show that the latter (d_b) varies according to the slope of the seabed β. The higher the seabed slope β the lower the local water depth d_b at this point. Both criteria give almost the same results. These results show that in the coastal zone of the Gulf of Guinea (Benin), the bathymetric breaking of the waves occurs at a position where the local water depth varies between 1,5m and 4,5m. As the slope increases, the local water depth at the breaking point decreases.

- The curves in **fig. 8** reflect the evolution of the height crest to trough H_b of the swell at the breaking point in the Gulf of Guinea (Benin). They are increasing functions of the seabed slope β, and show that the height of the swell H_b at the breaking point varies between approximately 1,6m and 2,3m. These swells therefore become more energetic when the slope of the seabed β increases.
- The curves of the **fig. 9** and **10** show the increasing variation of the energetic power of the orbiting marine currents of swells from the shoaling zone to near the coast, this explains the total energy dissipation generated by the swell in this zone. Swells induce an orbital motion in the water and the amplitude of this motion decreases with depth. Physically the water particles lose energy exponentially in the Surf and Swash zones. The discontinuity observed at the level of each curve explains the loss of energy in the breaking zone. Most of the energy power dissipation of the wave takes place in this so-called transition zone. This shows that wave breaking is a fascinating phenomenon that plays a major role in coastal processes. It is the source of very intense currents, responsible for most of the sediment transport in this area. Hence the degradation of the bottom observed near the coast.

As for the curves in **fig. 11** and **12,** they reflect the variations in the energy potential of these orbital marine currents at given local water depths. The evolution of these curves show that:
- For any given depth, the energy potential decreases from the surface to the bottom, this is due to the fact that the velocity orbital swell currents is higher when approaching the surface.
- The lower the depth, the lower this potential. That is to say, the energy potential of the orbital marine current decreases from the swell lifting zone towards the beach. The energy potential at a depth ($z \neq -d$), decreases from sea to coast. For the different months, each curve shows an upward trend. The energy power decreases with increasing depth and vanishes when the depth is equal to half the wavelength ($z = -d$). The Energy Potential increases from July to September due to the growth of the crest to trough height in the same period.

Conclusion:-

In the Gulf of Guinea at Cotonou, the swells are regular and have a constant average height$H_o = 0{,}8m$and an average period $T = 11s$in deep waters. In the coastal zone, the disturbing effect of the seabed causes them to rise to the breaking point and it is the modified Boussinesq theory, proposed by Peregrine in 1967, which makes it possible to model them. These swells become very energetic in this zone, their height is amplified and remains proportional to($d^{-1/4}$). At the breaking point, the maximum height reached by these swells varies between $1{,}7m$ and $2{,}5m$. Their bathymetric surge, occurs at a position where the local water depth$d_b$oscillates between 1.6m and 4.5m very close to the coastline depending on the value of the slope of the seabed. This Breaking is a sudden energy discharge that induces and accentuates the phenomenon of coastal erosion.

Swells induce an orbital motion in the water and the amplitude of this motion decreases with depth. It is important to note that the amplitude of the velocity decreases exponentially with a coefficient proportional to the number of waves k. So the orbital velocity of swells with a longer period (smaller wave number) will be more present at seafloor than that of waves with shorter periods. The propagation velocity of the swells depends on their wavelength, their amplitude and when approaching the coast, the water depth. The swells are therefore characterized by energy spectra, which show characteristic quantities, with for example a significant height H_s, a significant period T_s, etc….

The energy potential of the orbital marine current decreases from the wave lifting zone towards the beach. The energy potential at a depth ($z \neq -d$), decreases offshore towards the coast. For the different months, each curve shows an upward trend. The energy output decreases when the depth increases and cancels out when the depth is equal to half the wavelength ($z = {}^{L}/_{2}$). The Energy Potential increases from July to September due to the increase in height crest at trough in the same period. The swells therefore become more energetic when the slope β of the seabed increases. They are very energetic when approaching the coast. The breaking that occurs very close to the coastline and the sandy nature of the coastal soil in Cotonou (non-cohesive sediments), are the major causes of the coastal erosion that is observed along this site. A double benefit for all the countries of the Gulf of Guinea is the conversion of these energies into electrical energy in the shoaling zone in order to solve both the problem of the energy crisis and to attenuate and reverse the brutal energy discharge of these waves. away from the coastline, thus ensuring its protection.

Table of Scientific Notations

v: Ocean water flow velocity (m/s);
V_x : Horizontal component of the velocity of water particles struck by the swell (m/s) ;
V_z : Vertical component of the velocity of water particles struck by the swell (m/s) ;
V_g: Internal wave group velocity ($m.s^{-1}$) ;
C_g: Swell group velocity ($m.s^{-1}$) ;
β : Slope of the seabed;
β_m: Average slope of the seabed;
C_φ: Swell phase velocity ($m.s^{-1}$) ;
η : Vertical elevation of the water level relative to the reference (m)
z: Altitude (m) ;
H_0: Crest-to-trough height of the offshore swell (m) ;
H_b: Crest-to-trough height of the swell at the point of breaking (m) ;
H: Crest-to-trough height of swell in the coastal zone (m) ;
d_0: Local water depth of the offshore (m) ;
d_b: Local water depth at the breaking point (m) ;
d: Local water depth of offshorein the coastal zone (m) ;
L_0: Wavelength of swell of the offshore (m) ;
L: Wavelength of swellin the coastal zone (m) ;
T_0: Swell periodof the offshore (s) ;
T: Swell periodin the coastal zone (s) ;
T_m: Average swell period (s) ;
$\vec{r}$: Position vector of a point on the free surface;

$\vec{k}$: wave vector ;
P_0: Pressure at the reference (P_a)
ρ_0: Density at reference ($kg.m^{-3}$)
g: Gravitational acceleration in Cotonou / CRHOB/CBRST ($\approx 9{,}79\ m.s^{-2}$) ;
ω: Wave pulsation ($rad.s^{-1}$)
k: Wavenumber (rad/m)
μ: Dynamic viscosity ($P_a.s$)
$\vartheta = \frac{\mu}{\rho_0}$: Kinematic viscosity ($m^2.s^{-1}$)
$\tilde{\rho}$: Small fluctuation in density ($kg.m^{-3}$) ;
$\tilde{P}$: Small fluctuation in pressure (P_a) ;
$\gamma = \frac{H_b}{d_b}$: Breaking index proposed byKaminsky and Kraus (1993) ;
$Y_0(z)$: Modified Bessel function of the 2èmespecies ;
E_c: Kinetic energy of the swells (J) ;
E_p: Potential energy of the swells (J) ;
E_t: Total energy of the swells (J) ;
P: Energy potential (Power) of the internal waves at an altitude z (Watts) ;
χ : Variations coefficient of the swells energy power per unit length;
ψ : Variations coefficient of the total energy of the swells per unit length;
$\delta = \frac{L}{L_0}$: Variations coefficient of the wavelength;
$\delta_1 = \frac{C_\varphi}{C_{\varphi_0}}$: Variations coefficient of the phase velocity;
$\delta_2 = \frac{C_g}{C_{g_0}}$: Variations coefficient of the group velocity ;
$\tau = \frac{T}{T_0}$: Variation coefficient of the period;
$\kappa_s = \frac{H}{H_0}$: Variation coefficient of the crest-to-trough height of the swell ;
E'_c: Kinetic energy of orbital marine currents generated by swells (J) ;
E'_p: Potential energy of orbital marine currents generated by swells at an altitude z (J) ;
E_t: Total energy of the orbital marine currents generated by the swells (J) ;
P': Energy potential (Power) of orbital marine currents at an altitude z(Watts) ;

Authors Contributions

All authors have conceived the study. The main author led the drafting and coordination of the work (manuscript). All co-authors contributed to the draft and gave final approval for publication.

Conflict of Interest

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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Data availability

The datasets generated during and/or analyzed during the current study are available from the authors on reasonable request.

References:-

[1] Guy Hervé Hounguè, Basile B. Kounouhewa, Rafael Almar, Zacharie Sohou, Jean-Pierre Lefebvre, Mathias Houépkonhéhaand BernardTokpohozin(2018) ‘’ Waves Forcing Climate on Bénin Coast, and the Link with Climatic Index, Gulf of Guinea (West Africa)’’ Journal of Coastal Research; P 130–137; www.JCRonline.org

[2] Kounouhéwa B. B.;Houékpohéha A. M.; Tokpohozin N. B.; Awanou C. N. et Chabi-Orou B. J., 2014, Simulation du potentiel énergétique de la houle au Port Autonome de Cotonou, Journal de la Recherche Scientifique de l'Université de Lomé (JRSUL-Togo), JRSUL-0029_E Vol. 16 (2).
[3] BADAROU M. R., 2005/2006. Energie et écodéveloppement : République duBénin.
[4] Farah Zemani, Benameur Hamoudi, 2014. Effet de la profondeur de la mer sur la propagation des vagues' ; 2ème Congrès International en Aéro-Hydrodynamique (Faculté de Génie Mécanique USTMB BP 1505 El M'Naouer Oran AlgeriaCorrespondingauthor : zemanifarah@live.fr).
[5] Abbasov I. B., 2012, Numerical Simulation of Nonlinear Surface Gravity Waves Transformation under Shallow-Water Conditions , Applied Mathematics,3, 135-141, http://dx.doi.org/10.4236/am.2012.32021 , (http://www.SciRP.org/journal/am).
[6] Galan A., Simarro G., OrfilaA., Simarro J. and Liu P. L.-F., 2012, A Fully Nonlinear Model for Water Wave Propagation from Deep to Shallow Waters , Journal of Waterway, Port, Coastal, and Ocean Engineering,**Vol. 138, No. 5**, ©ASCE, ISSN 0733-950X/2012/5-1–10.
[7] Bougis J., 2004. ''Modèle d'approche de la houle par une méthode de réfraction - diffraction en coordonnées curvilignes'', Revue Paralia, VIIIème Journées Nationales Génie Civil – Génie Côtier, Compiègne, pp 55-64.
[8] ZAKI W., BASBOUS T., YOUNES R. et RAAD S., 2004. Modélisation de la houle au Port de Beyrouth. UniversitéLibanaise, Faculté de Génie, Beyrouth, Liban, 281-292.
[9] Hsu T.-W., Lin T.-Y., Wen C.-C. And Ou S._H., A complementary mild-slope equation derived using higher-order depth function for waves obliquely propagating on sloping bottom, Phys. Fluids**18(8),** 087106, 1-14, © 2006 American Institute of Physics.
[10] Webster W. C., Duan W. and Zhao B., 2011,Green-Naghdi Theory, Part A: Green-Naghdi (GN) Equations for Shallow Water Waves, J. Marine Sci. Appl., **10**, 253-258, v DOI: 10.1007/s11804-011-1066-1, Article ID: 1671-9433(2011)03-0253-06.
[11] ZOU Q., HAY A. E. and BOWEN J. A., 2003. Vertical structure of surface gravity waves propagating over a sloping seabed: Theory and field measurements. Journal of Geophysical Research 108 (C8, 3265), doi:10.1029/2002JC001432, 21.1-21.15.
[12] Fenton J. D., 1990, Nonlinear Wave Theories,The Sea, Vol.9: Ocean Engineering Science, Eds. B. Le Méhauté and D.M. Hanes, Wiley, New York.
[13] Didenkulova I. I. and Pelnovsky E. N., 2008, Run-up of long wave on a Beach: The influence of the incident wave form, Oceanologiya (Marine Physics), Vol. 48, 5-10.
[14] Gonzalez L. T., 2009, Calculs exactes des paramètres des principales théories non linéaires des vagues et proposition d'un outil simplifié de calculs, Thèse N°d'identification 0/1/A/I/X/3/0/1/0/3 de la faculté des études supérieures de l'Université de Laval Québec (Génie Civil), 1-251.
[15] Chazel F., 2007, Influence de la topographie sur les ondes de surfaces, Thèse N°d'ordre 3436 de l'Université de Bordeaux I, (Ecole doctorale de Mathématiques et Informatiques), 1-139.
[16] Lannes D. and Bonneton P., 2009, Derivation of asymptotic two-dimensional time-dependent equations for surface water wave propagation, Physics of Fluids DOI: 10.1063/1.3053183, 21, 016601, 1-9.
[17] Houékpohéha M. A., Kounouhéwa B. B., Hounsou J. T., Tokpohozin B. N, Hounguèvou J. V., and Awanou C. N. 2015. ''Variations of wave energy power in shoaling zone of Benin coastal zone.''Int. Journal of Renewable Energy Development, 4(1), 64-71. http://dx.doi.org/10.14710/ijred.4.1.64-71.
[18] Sabatier F., 2001, Fonctionnement et Dynamiques morpho-sédimentaires du littoral du Delta du Rhone, ThèseN°d'identification 0/1/A/I/X/3/0/1/0/3 de l'Université de AIX-Marseille III, (Ecole doctorale de la faculté des sciences et techniques de Saint Jerôme).
[19]TSAI C-P., CHENH-B. and HUANG M-J.,2002. Wave Shoaling on Steep Slopes and Breaking Criteria. Proceedings of The Twelfth (2002) International Offshore and Polar Engineering Conference, Kitakyushu, Japan, Copyright © 2002 by The International Society of Offshore and Polar EngineersISBN 1-880653-58-3 (Set); ISSN 1098-6189 (Set), 617-621.
[20] Senechal N. 2003. Etude de la propagation des vagues au-dessus d'une bathymétrie complexe en zone de surf, thesis of Bordeaux I University, N°: 2759, 1-288.
[21] Tissier M., 2011, Etude numérique de la transformation des vagues en zone littorale, de la zone de levée aux zones de surf et de jet de rive,Thèse N°d'ordre 4437 de l'Université de Bordeaux I, (Ecole doctorale Sciences et Environnements), 1-193.
[22] Umeyama M. 2010. Eulerian - Lagrangian analysis for particle velocities and trajectories in a pure wave motion using particle image velocimetry, Mathematical, Physical and Engineering sciences, doi: 10.1098/rsta.
[23] Catálan P. A. and Haller M. C., 2008, Remote sensing of breaking wave phase speeds with application to non-linear depth inversions , Coastal Engineering , 55, 93 – 111, www.elsevier.com/locate/coastaleng.

[24] TokpohozinN.B., Kounouhéwa B., Avossevou G. Y. H., Houékpohéha M A., Awanou C. N., .2015. Modelling of sediment movement in the surf and swash zones. ActaOceanologicaSinica, 34(2): 137–142, DOI **:** 10.1007/s13131-015-0610-2,http://www.hyxb.org.cn.
[25] Bernard Noukpo Tokpohozin, Basile BrunoKounouhewa, Zacharie Sohou, Mathias AdjimonHouékpohéhaetCossiNorbert Awanou, 2015. Modélisation de l'évolution du trait de côte sur le littoral du Bénin dans le Golfe de Guinée. AfriqueSCIENCE11(4) (2015) 1-15 1 ISSN 1813-548X, http://www.afriquescience.info.