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Coastal Processes under Hurricane Action: Numerical Simulation of a Free-boundary Shoreline

by

Peter William Sloss

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF

Doctor of Philosophy

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INTRODUCTION

Major storms are powerful agents of change--geologic, ecologic, hydrologic, and sociologic. Time and again coastal habitations have been obliterated by great storms. Undoubtedly, loss of life and destruction of property provided the initial motivation for studies of storm effects on sea level. The present study expands and refines some numerical simulation methods which have met with success in modelling storm surges.

The present study follows the state of the art in the use of net horizontal volume transport over a single layer to represent a vertically-uniform average horizontal velocity, but also includes the oscillatory currents from wind-waves in considering the maximum water currents under storm conditions. The principal addition made here to previous surge models is the inclusion of advective terms in the equations of motion and the use of a freely-moving boundary at the shoreline where other studies have used a stationary vertical-wall boundary. The storm high-water mark lies inland of the sea-level coastline; the model used in this study allows the surge to ride over low-lying coastal areas to carry erosive currents and waves inland. The effects of surge flooding may thus be examined quantitatively.
Longshore drift currents due to wave action are not considered here, although they can be effective in moving sediment. The drift currents generated by wind drag and surge movement are assumed to dominate the long-term motions. Oscillatory wave action should dominate the current regime in deeper water with large waves, while surge drift currents should represent most of the motion in shallow water where the large storm waves have broken. In both cases, it is assumed that the sum of wave and surge currents produces net motions stronger than either component alone with the result that larger particles may be eroded by storm effects that can be accounted for by either waves or surges alone. Large surge rises at the shoreline also can bring large waves into usually shallow water. The present one-layer model does not include any variation in current direction or speed with depth other than the application of linear wave theory to the bottom currents due to waves.

The end products of the present model are time-varying fields of volume transport and surge elevation which may be coupled with wave estimates to compute water depths and velocities for use in erosion studies, analysis of forces on offshore structures, tracing movements of water masses (and their contained pollutants), and flood-protection studies. The methods presented here are to be viewed as tools for studies on a variety of scales—a specific model of hurricane winds and pressures and
particular wave forecasting equations are used to examine a single case (hurricane Camille), but the general numerical method for surge computations should be valid for other storms, other bodies of water, and other wave-forecasting techniques better-suited to a different case. The numerical method and its associated computer program have been left as flexible as possible for broad potential applications in the future.
CHAPTER I

SURVEY OF PREVIOUS INVESTIGATIONS

Major storms of the past.

A review article by Welander (1961) summarizes some particularly devastating storms of recent centuries. In 1780 the islands of Santa Lucia and Martinique were raked by a great hurricane which took some 15,000 lives. This toll approaches that of another more recent disaster on Martinique--the 1903 eruption of Mt. Péléé, which killed some 28,000. The densely-populated coast of the Bay of Bengal has seen some 250,000 deaths from cyclones in 1864 and 1876; this toll was far exceeded by storms in November of 1970 and 1971, the total deaths from which may never be counted. In 1900 and 1915 hurricanes crashed into Galveston, Texas, killing some 6,000 and erasing the nearby resort town of Nottingham from the island. The destruction wrought on Galveston caused the shifting of this major seaport and population center inland to Houston. In 1953, an extra-tropical cyclonic storm breached the North Sea dikes of Holland, flooding over 25,000 square kilometers and causing some 2,000 deaths. Japan saw some 5,000 dead from typhoon Vera of 1959. Even inland waters may react disastrously to storms. A storm surge from a squall line passing over Lake Michigan drowned six fishermen
on a pier in Chicago's Montrose harbor on the morning of 26 June 1954 (Platzman, 1958). Another lake surge, on Lake Okeechobee, Florida, killed some 2,000 people in 1928. Hurricane Camille, 1969, was one of the most destructive storms in recent memory to strike the United States, leaving 262 dead and $1.0 billion in property damage in a swath across six states (U.S. Army Corps of Engineers, 1970). The primary source of destructive energy in all of the above cases was the storm surge, sometimes erroneously called a tide, produced by the action of wind and barometric pressure gradients on the ocean surface.

**Mathematical and numerical studies.**

An early mathematically-elegant treatment of sea waves was by Airy (1845). The effect of an atmospheric pressure disturbance was investigated first by Chrystal (1908), although most studies prior to World War II concentrated on wind-wave processes (Bretschneider, 1967). Hellstrom (1941) discussed wind set-up of water along lake and river shorelines where frictional drag from wind over water caused mass transport to the lee shore and a consequent rise in the free surface.

The combined effects of wind and pressure to produce a surge were subjected to numerical analysis by Freeman, et al. (1957) and many subsequent studies. The advent of high-speed computers has greatly accelerated the attack
on storm surges as a hydrodynamical problem. Platzman (1958) used numerical techniques to model a lake seiche. Charnock and Crease (1957) applied dimensional arguments in the linearization of the equations of motion to arrive at a set of equations similar to those used by Freeman, et al., Harris and Jelesnianski (1964), Harris (1966), Bretschneider (1967), and Sirkin (1970). Other equations utilizing nonlinear terms have been treated by Hansen (1956), Ito, et al. (1964), Isozaki (1970a, b), Jelesnianski (1965, 1966, 1967), and Galt (1971). All of these investigations have provided useful operational surge-forecasting techniques, but they were concerned mainly with prediction of sea-level rise in areas where coastal structures would be threatened. There has been little numerical analysis of the storm-surge currents. The present study is intended to pursue this topic.

Observations and predictions of storm-generated currents.

Currents may be generated in shallow water by processes other than an onwashing surge. Saylor (1966) calculated that a wind-driven current should reach 2 to 3% of the sustained wind speed. Murray's (1970a) measurements during hurricane Camille, 1969, of currents of 5% of the wind speed were somewhat higher than Saylor's range of values, although Saylor's values agreed with predictions by Reid (1957). Murray also noticed a strong offshore bottom flow after the storm center passed onshore, although
the wind was still onshore at the measurement point. This bottom countercurrent was cited in a flow model by Kajiura (1959) and also by Hellstrom (1941). Kajiura found that under conditions of a storm large in comparison to the continental shelf, the primary water circulation was vertical--onshore, parallel to the wind at the surface, with a bottom countercurrent running offshore. For storms smaller than the scale of the shelf, the circulation was primarily horizontal and followed the wind.

**Geologic effects of storms.**

There are many discussions of the erosive action of surf and normal wind-generated gravity waves. Among the best known of these studies is a monograph by Ingle (1966) on the movement of beach sand. Ingle's work provides insight into the mechanisms of wave erosion and littoral transport, but avoids consideration of storm conditions. Komar and Inman (1970) consider the energy balance in longshore transport, finding a longshore transport rate dependent on water depth, wave energy and approach angle. Johnson and Eagleson (1966) look at the momentum balance of moving sediment under the influence of hydraulic drag, gravity, and rolling friction, arriving at an energy balance expression dependent on the net water motion integrated over one wave cycle. A similar approach is taken by Harms (1969) in trying to relate the spacing of ripple marks to the distance over which sand is transported by
a single wave cycle. The work of Kalkanis (1964), however, shows that a particle may move back and forth over several ripple crests during a wave cycle. Mathematical treatments in the above studies assume either non-breaking waves which are small in height compared to water depth, or apply non-breaking wave theory even inshore of the breaker line.

These studies of wave-generated oscillatory motion have less applicability under storm conditions, in which the prevailing mean sea level is disturbed by the storm surge. Water depths may be raised temporarily far above normal, allowing much larger gravity waves to act on shallow-water sediments. Strong currents generated by the moving surge move large amounts of sediment under a flow regime which is more fluvial than wavelike. The similarity between storm-produced washover structures and tidal deltas has been noted in some investigations (Hayes, 1967; McGowen, et al., 1970).

Geologic investigations of hurricane effects have centered around observations of disruption of the biota and the redistribution of sediments. The effects on coral growths observed by Ball, et al. (1967) were seen to be a "large amount of boulder-sized rubble formed by hurricane surf on platform-edged reefs (which) far exceeded the amount produced by day-to-day processes." Perkins and Enos (1968) compared the effects of two storms on the same area, finding that the amount of erosion seemed to depend on the direction of the storm travel, while the
destruction of reef material occurred wherever there were large waves breaking. Stoddart (1964, 1969) also found deposition of mud above normal sea level, and regarded storm action as an accretionary process at places where vegetation was capable of trapping washover sediments. Behrens (1969) examined the effects of hurricane floodwaters, fresh rain as well as saline seawater, on the ecosystem of Baffin Bay, Texas. In his study, a single storm was able to change drastically the balance of life in the bay as rain floods killed the saltwater biota and salt flooding from the storm surge killed brackish swamp growths. Hayes (1967) and High (1969) also reported such effects in coastal lagoons in Texas and British Honduras.

Erosion due to hurricane action is the result of both rising and ebbing surge waters, although many traces of onshore-flow erosion would be wiped out by the ebb currents. Ball, et al. (1967) and Perkins and Enos (1968) noted that ebbing surge waters left large deposits of lime mud on supratidal flats, a possible explanation of the large amount of tidal-flat facies in ancient rocks. Ball, et al. also found large amounts of material removed from artificial fill, and Perkins and Enos noted erosion of "spillover lobes" created by earlier storms in the Cat Cay oolite belt which dug depressions down to the bottom of the pelletal layer and removed up to 140 feet of the avalanche slope of the lobes. All investigators noted that many of the hurricane-eroded features had
healed within a few months of their formation. Stoddart (1969) noted, however, that as much as three meters of new material could be accumulated above normal sea level by a single storm and that such accumulations lithified rapidly.

There may be only a small number of large-storm events represented in the geologic record. By assuming a Poisson probability distribution for "rare" or low-probability events, Greten (1967) stated a yearly probability of 0.001 for a hurricane to affect a given piece of coastline. Data taken over the geologically brief span of recorded human history show this figure to be reasonable or perhaps too low, since McGowen, et al. (1970) gave an annual frequency of 0.67 for hurricanes hitting the Texas coast. While the eye of a storm may cross a particular small portion of a coast with a frequency of once every few thousand years, a single storm may affect a very large region at one time—hurricane Carla of 1961 was felt on the entire Gulf of Mexico coastline (Hayes, 1967). The observations by Stoddart (1969) of up to several meters of new beach rock formed by a single storm contrast strongly with normal marine sedimentation rates of less than 10 cm in 1,000 years given by Rubey and Hubbert (1959). The lime mud flats mentioned by Perkins and Enos (1968) also represented an abnormally high accumulation rate. It may well be, therefore, that in coastal sedimentation only over the time scale of
millenia does a deposition rate have any meaning. Doehring and Vierbuchen (1971) noted evidence of catastrophic storms as a major agency in cave formation. Gretener used the term "catastrophic uniformitarianism" to characterize storm sedimentology.

Coleman (1966) noted that storm beaches have "an irregular but gently seaward shoreface; an exceptionally flat-topped crest; and a steep, landward-dipping, abrupt slope adjacent to the marsh." Internal structures showed steeply landward-dipping laminations of shell and sand which indicated strong onshore transport. The work of Ball, et al. (1967), coupled with these observations, would imply that onshore and offshore transport occur in different places. Krumbein and Aberdeen (1937) noted that storm-washed beach sediments were well-sorted and excess fines were found in the bays behind Grand Isle, Louisiana, after the passage of a storm of unspecified intensity. Thompson (1955) examined Atchafalaya Bay, Louisiana, noting that fine bay sediments have a low sediment-water interface roughness and erode only slightly even under storm conditions. He found bay erosion minor compared to the cutting of barrier islands exposed to large waves. The position of offshore shoals has a controlling effect on the location of maximum erosion. Erosion due to waves is greater behind gaps in the shoals off Cape Cod than where the shore is protected by the shoals (Zeigler, et al., 1959). Zenkovich (1967) attributed
almost all the erosion around Odessa (on the Black Sea) to wind-driven surges; a layer of coarse detritus protected the shoreline from normal waves. Thus, the erosional regime of a coastline may be closely coupled to the local sediment texture.

**Means of sediment transport.**

The mechanism by which a storm surge transports sediments may be characterized as a turbid underflow in the surge currents. Thompson (1955) proposed saltation along the bottom as a primary transport mode. Turbidity currents triggered by storm surges would produce characteristic structures such as were found by Fisher and Mattinson (1968). Such a process was suggested by Coleman (1968), with a tsunami as the triggering wave. There has been some discussion by Southard (1970, 1971) which questions the erosional competence of turbidity currents, but the depositional signature of such currents can be easily recognized (Middleton, 1967).
CHAPTER II

STORM SURGE MODEL

The equations of motion.

The equations of motion used in this study are based on the classic equations of Ekman (1905), which describe wind-induced water motions on a rotating earth. Water current velocity components in the $x$-(=east), $y$-(=north), and $z$-(=upward) directions are $u$, $v$, and $w$, respectively. Water motions are assumed quasi-horizontal, such that quantities directly dependent on the vertical velocity are neglected. The equations of horizontal motion are

$$\frac{du}{dt} + u\frac{du}{dx} + v\frac{du}{dy} = fv - \frac{1}{\rho_w} \frac{dp'}{dx} + \frac{1}{\rho_w} \frac{\partial \tau_x}{\partial z} \tag{1}$$

$$\frac{dv}{dt} + u\frac{dv}{dx} + v\frac{dv}{dy} = -fu - \frac{1}{\rho_w} \frac{dp'}{dy} + \frac{1}{\rho_w} \frac{\partial \tau_y}{\partial z} \tag{2}$$

where $p'$ is the total pressure, $f$ is the Coriolis parameter, $\rho_w$ is the density of sea water, and $\tau_x$ and $\tau_y$ are the tangential stresses in the $x$- and $y$-directions. Figure 1 illustrates the geometric relations among $u$, $v$, the perturbation from Mean Sea Level (MSL) of the free surface, $h$, the net water depth,

$$\zeta = d + h,$$

and prevailing water depth, $d$, on a portion of the finite-difference grid to be defined below (cf. Figure 15).
Fig. 1.--Isometric projection of a portion of data grid. Heavy lines represent free surface, \( h \); thin lines represent bottom and land topography, \( d \). Broken-line plane is Mean Sea Level. Grid squares are 15 km on a side. Vertical scale greatly exaggerated. Net water depth is \( h \). Volume-flux vectors \( u \) and \( v \) are shown for one grid location.
The continuity condition for an incompressible, isothermal fluid (water shall be treated as such) is given by
\[ \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0. \] (3)

Under the assumption of quasi-horizontal motions, vertical stratification of the horizontal flow is still allowed, but for numerical analysis a stratified-flow model requires definition of multiple layers within the fluid, each with its own set of velocities. It is common practice, in the present state of the art, to integrate the vertical degree of freedom out of the equations of motion and represent the flow in terms of a volume flux or transport per unit width. This flux in turn defines a uniform vertical-average velocity characteristic of the flow and equal to the flux divided by the local net water depth. The $x$- and $y$-fluxes and average velocities are called $M$, $N$, $\bar{u}$, and $\bar{v}$, given by

\[ M = \int_d^h u \, dz \]
\[ N = \int_d^h v \, dz \]
\[ \bar{u} = M/C \]
\[ \bar{v} = N/C \]

(1) and (2) become
\[ \frac{\partial M}{\partial t} + \int_d^h (u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y}) \, dz = fN + \frac{C}{\rho_w} \frac{\partial p^*}{\partial x} + \frac{\tau_{sx} - \tau_{x}}{\rho_w} \] (4)
\[ \frac{\partial N}{\partial t} + \int_{-d}^{h} (u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y}) \, dz = -\tau M + \frac{c}{\rho_w} \frac{\partial p}{\partial y} + \frac{\tau_Y - \tau_X}{\rho_w} \]  

(5)

where the stresses \( \tau_X \) and \( \tau_Y \) have been evaluated at 
\( z = -d \) (bottom, subscript \( b \)) and \( z = h \) (surface, subscript \( s \)). The integrals of the advective terms on the left of each of the above equations may be expressed as

\[ \int_{-d}^{h} (u \frac{\partial u}{\partial x} + v \frac{\partial v}{\partial y}) \, dz = \int_{-d}^{h} \left[ \frac{\partial (u^2)}{\partial x} + \frac{\partial (uv)}{\partial y} \right] \, dz - \left( \frac{1}{2} \frac{\partial (u^2)}{\partial x} + u \frac{\partial v}{\partial y} \right) \, dz \]  

(6)

and

\[ \int_{-d}^{h} (u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y}) \, dz = \int_{-d}^{h} \left[ \frac{\partial (uv)}{\partial x} + \frac{\partial (v^2)}{\partial y} \right] \, dz - (v \frac{\partial u}{\partial x} + \frac{1}{2} \frac{\partial (v^2)}{\partial y}) \, dz. \]  

(7)

Notice that the subtracted terms on the right of (6) and (7) are of the form

\[ \frac{1}{2} \frac{\partial (u^2)}{\partial x} + u \frac{\partial v}{\partial y} = u (\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}) \]  

(8)

and

\[ v \frac{\partial u}{\partial x} + \frac{1}{2} \frac{\partial (v^2)}{\partial y} = v (\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}) \]  

(9)

and that the results (8) and (9) contain the horizontal divergence of the velocity field, or \( -\partial w/\partial z \). It is to be assumed that the averaged motion in the vertical direction, \( w \), will be small compared to \( u \) and \( v \), with its effects seen only in displacements of the free surface. Thus, to a first approximation, the terms in (8) and (9) vanish.

Under the further assumption that integration over
z may be permuted with differentiation over x or y, the
advective terms in equations (4) and (5) may be rewritten
(with the restriction applied from [8] and [9]) as,

\[ \int_{-d}^{h} (u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y}) dz = \frac{\partial M}{\partial t} + \int_{-d}^{h} \frac{\partial M}{\partial x} dz + \frac{\partial N}{\partial y} + \int_{-d}^{h} \frac{\partial N}{\partial y} dz \]

(10)

\[ = \frac{\partial M}{\partial x}(\bar{u}^2) + \frac{\partial N}{\partial y}(\bar{u}\bar{v}) \]

(11)

and

\[ \int_{-d}^{h} (u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y}) dz = \frac{\partial N}{\partial t} \int_{-d}^{h} \frac{\partial N}{\partial x} dz + \frac{\partial M}{\partial y} \int_{-d}^{h} \frac{\partial M}{\partial y} dz \]

(12)

\[ = \frac{\partial N}{\partial x}(\bar{u}\bar{v}) + \frac{\partial M}{\partial y}(\bar{v}^2) \].

(13)

While the mean-squared velocities and velocity cross-
product averages in (10) to (13) would reflect the vertical
distribution of horizontal velocities, this degree of
freedom has been integrated out of the problem and the
mean velocity is all that can be recovered from the
volume transport—their assumed near-equivalents the
squared mean and product of means will be used as follows:

\[ \bar{u}^2 \approx (\bar{u})^2, \bar{v}^2 \approx (\bar{v})^2 \]

and

\[ \bar{u}\bar{v} \approx \bar{u}\bar{v}. \]

Thus, the left sides of equations (4) and (5) become

\[ \frac{\partial M}{\partial t} + \int_{-d}^{h} (u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y}) dz = \frac{\partial M}{\partial t} + \frac{1}{\zeta} (\frac{\partial M^2}{\partial x} + \frac{\partial MN}{\partial y}) \]

and

\[ \frac{\partial N}{\partial t} + \int_{-d}^{h} (u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y}) dz = \frac{\partial N}{\partial t} + \frac{1}{\zeta} (\frac{\partial MN}{\partial x} + \frac{\partial N^2}{\partial y}) \].

The equation of continuity (3), when integrated over
z from z = -d to z = h, becomes
\[
\frac{\partial M}{\partial x} + \frac{\partial N}{\partial y} + w_h - w_d = 0
\]

where

\[
w_h = \frac{\partial h}{\partial t}
\]

and

\[
w_d = 0.
\]

Thus, the vertically-integrated continuity condition is

\[
\frac{\partial M}{\partial x} + \frac{\partial N}{\partial y} = -\frac{\partial h}{\partial t}
\]

(14)

The pressure gradient really consists of two parts. The first is the hydrostatic balance of the free surface against barometric pressure,

\[
h_o = \Delta p_a / \rho_w g
\]

(15)

where \(\Delta p_a\) is the local deviation of the atmospheric pressure from its value outside the storm, \(g\) is the acceleration due to gravity, and \(\rho_w\) is the density of water. The second part is the slope of the free surface as it responds to the divergence of the volume transport as given by (14) above. Taking the specific gravity of sea water as 1.03, the "inverted barometer effect," \(h_o\), may be written as 0.9907 cm of water for every millibar of pressure change. Thus, equation (15) may be written

\[
h_o = 0.009907 \Delta p_a
\]

for \(h_o\) in meters and \(\Delta p_a\) in millibars. The pressure-gradient terms are then written

\[
\frac{1}{\rho_w} \frac{\partial \Delta p_a}{\partial x} = g \frac{\partial h}{\partial x} - \frac{1}{\rho_w} \frac{\partial p_a}{\partial x}
\]
and
\[
\frac{1}{\rho_w} \frac{\partial p^*}{\partial y} = q \frac{\partial h}{\partial y} - \frac{1}{\rho_w} \frac{\partial p_a}{\partial y}.
\]

The last terms on the right of equations (4) and (5) represent the \(x\)- and \(y\)-components of the tangential stress on the water surface and sea bottom, \(\tau^x_s\), \(\tau^y_s\), \(\tau^x_b\), and \(\tau^y_b\), respectively. The surface stresses may be written in the form
\[
\tau^x_s = \rho_a a^2 q (q^2 + r^2)^{1/2}
\]
and
\[
\tau^y_s = \rho_a a^2 r (q^2 + r^2)^{1/2}
\]
where
\[
a^2 = 0.0022 \text{ to } 0.0026
\]
(following Wilson [1960] and Miyazaki [1965]), and \(q\) and \(r\) are the \(x\)- and \(y\)-components of the wind field, and \(\rho_a\) is the density of air. The bottom stresses in shallow water (less than 100 m) have given by Reid (1957) as
\[
\tau^x_b = \rho_w a^2 \bar{u} \sqrt{(\bar{u}^2 + \bar{v}^2)^{1/2}} - 5 a \tau^x_s
\]
and
\[
\tau^y_b = \rho_w a^2 \bar{v} \sqrt{(\bar{u}^2 + \bar{v}^2)^{1/2}} - 5 a \tau^y_s
\]
where \(\bar{u}\) and \(\bar{v}\) have been assumed to be the currents at the bottom boundary layer.

There remains some variety among investigators in the drag coefficient to be used at the water surface in determining the current-producing stress there. Wilson
(1960) summarized a number of investigations, the consensus being that the coefficient for a quadratic drag function (stress proportional to wind-speed squared) was of the order 0.0015 to 0.0024, a value also assumed by Reid. Bottom-stress relations generally have been taken as quadratic functions of mean current, although Harris and Jelesnianski (1964) used a linear function in numerical prediction of surges. Jelesnianski (1965, 1966, 1967) subsequently used a quadratic friction law. Hansen (1956) used a value of 0.0026 which Wilson quoted for the wind stress coefficient (in stress calculations, the two coefficients also differ by the density ratio between air and water). For a wind of 50 m sec\(^{-1}\) and a current of 1 m sec\(^{-1}\), the correction terms dependent on the surface stress may reduce the bottom friction by about half, if the wind and current are parallel. Note that equations (16) and (17) may be combined to give the x-stresses

\[
(\tau_s^x - \tau_b^x)/\rho_w = \frac{[\tau_s^x(1 + 5y) - \tau_b^x]}{\rho_w}.
\]

Likewise, the y-stresses are

\[
(\tau_s^y - \tau_b^y)/\rho_w = \frac{[\tau_s^y(1 + 5y) - \tau_b^y]}{\rho_w},
\]

where the quantity \((1 + 5y)\) has the value 1.3 (Isozaki, 1970b).

Define

\[
M(M^o + N^o)^{1/2}/\epsilon^o = \bar{u}(\bar{u}^o + \bar{v}^o)^{1/2},
\]

and

\[
N(M^o + N^o)^{1/2}/\epsilon^o = \bar{v}(\bar{u}^o + \bar{v}^o)^{1/2};
\]
the full dynamical equations to be used here are thus,

\[
\frac{\partial M}{\partial t} + \frac{1}{\epsilon} \left( \frac{\partial M}{\partial x} + \frac{\partial MN}{\partial y} \right) = fN - g\epsilon \left( \frac{\partial h}{\partial x} - \frac{1}{\rho_w g} \frac{\partial p_a}{\partial x} \right) \\
+ \frac{\rho_a \gamma^2}{\rho_w} (1 + 5\gamma) q(q^2 + r^2)^{1/2} \\
- \gamma^2 M(M^2 + N^2)^{1/2} / c^2 \quad (18)
\]

and

\[
\frac{\partial N}{\partial t} + \frac{1}{\epsilon} \left( \frac{\partial MN}{\partial x} + \frac{\partial N^2}{\partial y} \right) = -fM - g\epsilon \left( \frac{\partial h}{\partial y} - \frac{1}{\rho_w g} \frac{\partial p_a}{\partial y} \right) \\
+ \frac{\rho_a \gamma^2}{\rho_w} (1 + 5\gamma) r(q^2 + r^2)^{1/2} \\
- \gamma^2 N(M^2 + N^2)^{1/2} / c^2 . \quad (19)
\]

**Finite-difference form.**

The partial-differential equations (18), (19), and (14) must be put into finite-difference form for computer analysis. In order to simplify the writing of the difference equations, a uniformly-spaced grid of points in \(x\) and \(y\) is used, with the point spacing

\[\Delta x = \Delta y = \Delta s.\]

All parameters are nondimensionalized by scaling with respect to \(\Delta s\) and the time step, \(\Delta t\). The space and time steps for the scaled difference equations become unity.

The nondimensionalized parameters for the computer are

\[M \times \Delta t/(\Delta s)^2 = U\]
\[N \times \Delta t/(\Delta s)^2 = V\]
\[h/\Delta s = H\]
\[d/\Delta s = D\]
\[ \zeta / \Delta s = Z \]
\[ \bar{u} \times \Delta t / \Delta s = U / Z \]
\[ \bar{v} \times \Delta t / \Delta s = V / Z \]
\[ (p_a / \rho_w) \times (\Delta t / \Delta s)^2 = P \]
\[ g \times (\Delta t)^2 / \Delta s = G \]
\[ f / \Delta t = F \]
\[ q \times \Delta t / \Delta s = Q \]
\[ r \times \Delta t / \Delta s = R \]
\[ x / \Delta s = X \]
\[ y / \Delta s = Y \]
\[ (Q^* + R^*)^{1/2} = W \]

Space derivatives are approximated by centered, first divided differences, e.g.,

\[
\frac{\partial U}{\partial X} \bigg|_{i,j} \approx \frac{(U_{i+1,j} - U_{i-1,j})}{2}
\]

where the index \( i \) denotes grid column (or position in \( X \)) and the index \( j \) denotes grid row (or position in \( Y \)). In some instances it becomes necessary to use a one-sided difference to approximate the derivative (see discussion of boundary conditions, below):

\[
\frac{\partial U}{\partial X} \bigg|_{i,j} \approx \frac{(U_{i+1,j} - U_{i,j})}{1}
\] (20)

where the quantity \( U_{i-1,j} \) is undefined and the differencing process spans two rather than three points.

For a typical point \((i,j)\), the difference equations to be used for storm surge prediction are
\[ U_{i,j}(t + \Delta t) - U_{i,j}(t) = [(U_{i+1,j} - U_{i,j}) \cdot U_{i,j+1} V_{i,j+1} \\
- U_{i,j-1} V_{i,j-1}) \\
\div 2(D_{i,j} + H_{i,j})] \\
\times \frac{G}{2}(D_{i,j} + H_{i,j}) \\
\times [H_{i+1,j} - H_{i,j+1} \cdot (P_{i+1,j} - P_{i,j})] \\
\div (D_{i,j} + H_{i,j})^a \] (21)

and

\[ V_{i,j}(t + \Delta t) - V_{i,j}(t) = [(U_{i+1,j} V_{i+1,j} \cdot U_{i,j-1} V_{i,j-1}) \\
\times U_{i,j} V_{i,j-1} - V_{i,j}^2 V_{i,j-1}) \\
\div 2(D_{i,j} + H_{i,j})] - F_{i,j} \\
\div \frac{G}{2}(D_{i,j} + H_{i,j}) \times [H_{i,j+1} \\
- (P_{i,j+1} - P_{i,j})] + \beta R_{i,j} W_{i,j} \\
\times [V_{i,j}^2 + V_{i,j} \cdot V_{i,j}^2] \\
\times (D_{i,j} + H_{i,j})^a \] (22)

where

\[ \beta = \gamma (1 + 5Y) \rho_a / \rho_w. \]
The difference expressions on the left of the equal signs in (21) and (22) above represent the derivative with respect to time, really at time \( t + \frac{1}{2} \Delta t \). Note that no time has yet been specified for quantities on the right of (21) and (22). These quantities will influence the time-difference function, making it forward, backward, or centered, depending on the "time" at which they are evaluated.

The equation of continuity (14) becomes

\[
H_{i,j}(t + \Delta t) - H_{i,j}(t) = \frac{1}{2} \left( U_{i+1,j} - U_{i-1,j} \right. + \left. V_{i,j+1} - V_{i,j-1} \right).
\]  

(23)

This completes the set of surge prediction equations for time \( t + \Delta t \).

Treatment of the boundaries.

At this point in the discussion it is time to examine the boundary conditions that control the solution of the prediction equations. Since it is impractical to model the entire world ocean at once, some assumptions must be made about conditions immediately outside the region being studied.

In the present study, the configuration of the shoreline is allowed to change in response to flooding or draining as surge waters rise or recede. The continuity equation (14 or 23) is used to control the amount of water at a grid point; if the volume flux divergence at a nearshore point is greater than the present water
depth divided by the time step, the point is defined as dry. Conversely, if the flux convergence surrounding a dry point is great enough, the free surface rises above the bottom there and floods a new point which is defined as wet. It is these newly-flooded areas which are of great interest from an engineering standpoint, since they represent sites of possible damaging coastal erosion and flooding of inhabited areas.

Previous investigations have made the model shoreline a stationary, rigid boundary with the properties of a no-slip vertical wall (Miyazaki, 1965), or a simple reflecting vertical barrier (Platzman, 1958). These investigations assume no flux of water into or out of the modelled region, making a closed control volume analogous to a wave tank or scale-model basin. If displacements are not too large, this method is acceptable, except very near the grid boundaries, where the model flow must conform to an unnatural shape. In order to keep these rigid boundaries far from the region of interest, a "patched grid" or varying mesh spacing is sometimes used (Jelesnianski, 1965).

An alternative method of treating the boundaries as "porous" according to known or predicted boundary fluxes has been used with some success by Miyazaki (1965). A small-area fine-mesh grid for high resolution is used in conjunction with large-scale predictions made in a previous computer run. In this way time-varying boundary
conditions may be used on a small-scale grid; in fact, a single large-scale prediction run can provide boundary conditions for several local studies.

In this model the outer grid boundaries hold the volume flux at its initial values (not necessarily zero) at each grid point while the free surface is in hydrostatic balance with atmospheric pressure on the grid boundaries.

**Moving the shoreline.**

Because of possible changes in the shoreline configuration, the gradients of velocity and pressure may have to be calculated according to a "one-sided" divided difference. In the case of a point where the point to its left is dry, the x-derivatives must be approximated by a form as given in (20). Similar expressions must be used if the north, east, or south neighboring grid point is dry.

Control of the shoreline position is supplied by the continuity equation, as described above, with a few modifications. Flooding a dry point requires that the volume flux convergence raise the free water surface above the land surface, but there is no water surface defined at a dry point. Therefore it is necessary to assign an arbitrary value to $H$ at every grid point on land and sea. Where the water level is below the land surface, the volume flux field is undefined. At those dry points
whose neighbors are wet, the water level is set equal to the average of its value at the point in question and those wet points among its eight nearest neighbors. If this average places the water above ground level, the point becomes wet with an initially zero volume flux. Volume flux values used for the continuity equations are averages of the newly-predicted flux (at time $t_o + [n + 1]\Delta t$) and the value from the previous time step (at time $t_o + n\Delta t$). A newly-flooded point then enters into all future flux calculations; a newly-dry point ceases to be considered. The calculated value of $H$ represents the free surface at time $t_o + (n + \frac{1}{2})\Delta t$, based on a centered time-difference of the volume flux divergence.

Surge forecasting equations for the computer.

The difference equations (21) and (22) must be rewritten in a form which is tractable for computer solution. As they stand, they represent forward time differences with the new values of $U$ and $V$ found strictly in terms of an explicit dependence on their values left over from the previous time step (cf. Fischer, 1959). At those points where new flooding occurs, explicit difference equations give a volume flux prediction that may not be in harmony with the general, established flow field. To balance the flow and free surface after the initial prediction, an implicit solution scheme is used (Richthmyer and Morton, 1967).
The presence of the viscous dissipation and nonlinear advective terms makes it difficult to recast the equations of motion in "conservation form," as used by Houghton, et al. (1966) and Lax and Wendroff (1964), according to Chorin (1968a). Likewise, the condition of movable boundaries at the shoreline makes it unclear that the decomposition of the flow field into separate nondivergent and irrotational fields (Chorin, 1968b) is applicable.

The implicit solution scheme.

The choice of an implicit scheme for this study was made after preliminary computations indicated that an explicit scheme based directly on (21) and (22) became unstable after only a few time steps for a model having complicated bottom topography. Stability was slightly increased by an implicit scheme which included the advective terms in the predictands, but long-term stability was not achieved until the bottom stress was included as well—see Equations (26) to (31) below.

The difference equations (21) and (22) are rewritten in two different forms for two different relaxation schemes. For the first estimate of the updated flow field, an entire grid row is solved by the tridiagonal algorithm (Carnahan, et al., 1969), with the equations in the following form:

\[ A U^*_{i-1,j} + B U^*_{i,j} + C U^*_{i+1,j} = D_x \] (24)
and
\[ AV_{i-1,j}^* + BV_{i,j}^* + CV_{i+1,j}^* = D_y \]  \hspace{1cm} (25)

where
\[ A = U_{i-1,j}/Z_{i-1,j} \]  \hspace{1cm} (26)
\[ B_x = 1 + U_{i,j}/Z_{i,j} + \gamma^d (U_{i,j}^2 + V_{i,j}^2)^{1/2}/Z_{i,j} \]  \hspace{1cm} (27)
\[ B_y = 1 + V_{i,j}/Z_{i,j} + \gamma^d (U_{i,j}^2 + V_{i,j}^2)^{1/2}/Z_{i,j} \]  \hspace{1cm} (28)
\[ C = U_{i+1,j}/Z_{i+1,j} \]  \hspace{1cm} (29)
\[ D_x = U_{i,j} + FV_{i,j} - Z_{i,j} (P_{i+1,j} - P_{i-1,j})/2 \]
\[ + \beta Q_{i,j} W_{i,j} - GZ_{i,j} (H_{i+1,j} - H_{i-1,j})/2 \]
\[ - [(U_{i,j+1}V_{i,j+1}/Z_{i,j+1}) \]
\[ - (U_{i,j-1}V_{i,j-1}/Z_{i,j-1})]/2 \]  \hspace{1cm} (30)
\[ D_y = V_{i,j} - FU_{i,j} - Z_{i,j} (P_{i,j+1} - P_{i,j-1})/2 \]
\[ + \beta R_{i,j} W_{i,j} - GZ_{i,j} (H_{i,j+1} - H_{i,j-1})/2 \]
\[ - [(V_{i,j+1}/Z_{i,j+1}) - (V_{i,j-1}/Z_{i,j-1})]/2. \]  \hspace{1cm} (31)

U and V are treated as constants from the previous time step, and \( U^* \) and \( V^* \) are predictands. The indices \( i \) and \( j \) range from the second to the next-to-last row and column.

Two concessions to simplicity have been made in the development above: \( U^* \) and \( V^* \) are solved for as first-order quantities, even where \( U \) or \( V \) appears squared or cubed, i.e., \( U^2/Z \) is written \( U^*x(U/Z) \), and so on. The unstarred values of \( U \) and \( V \) are constant through the relaxation iterations, keeping their values from the
previous time step until the solution for the updated flow has converged. To solve for higher orders of \( U^* \) and \( V^* \) would make (24) and (25) cubic equations and quite cumbersome. The second simplification for the tridiagonal algorithm is that the algorithm is applied along rows (constant \( j \), changing \( i \)). For this process the finite approximation to the \( y \)-derivatives of \( UV \) and \( V^* \) (differenced over \( j \)) produces no terms dependent on the values of \( U^* \) and \( V^* \) in the row being relaxed. Thus, the first estimate of the updated flow does not contain the effects of row-to-row variations in \( U^* \) and \( V^* \), but only variations between members of a row. An alternating-direction application of the tridiagonal algorithm could be used for further iterations—alternating the solution along rows and columns—but the usefulness of such a system in the present study is questionable due to the possible presence of interior dry points which may violate the mathematical assumptions under which the tridiagonal algorithm is used. An interior dry point or points breaks up the grid row into unconnected subrows which have no dynamical interaction. Hence, the first estimate of \( U^* \) and \( V^* \) must be refined by some scheme which will iron out imbalances on a scale smaller than an entire row.

The pointwise overrelaxation method is used here as a recursive correction process for matching local parameters and their gradients to their nearest neighbors, and hence, by a transitive relation, to all boundary values. In other
words, a grid point next to a boundary is brought into agreement with the boundary conditions; the next point interior to it is balanced against the first and by implication is therefore also in agreement with the boundary. Calling the four nearest neighbors to a typical point \((i,j)\) by their relative positions North \((i,j+1)\), South \((i,j-1)\), East \((i+1,j)\), and West \((i-1,j)\), the coefficients \(\alpha_N\), \(\alpha_S\), \(\alpha_E\), and \(\alpha_W\) for each parameter \(U\) and \(V\) may be determined from (21) and (22), the coefficient at the Point \((i,j)\) is called \(\alpha_p\). The coefficients are

\[
\alpha_p = 1 + \gamma^2 (U_{i,j} + V_{i,j})^{1/2}/Z_{i,j}
\]

\[
\alpha_N = V_{i,j+1}/Z_{i,j+1}
\]

\[
\alpha_S = V_{i,j-1}/Z_{i,j-1}
\]

\[
\alpha_E = U_{i+1,j}/Z_{i+1,j}
\]

\[
\alpha_W = U_{i-1,j}/Z_{i-1,j}
\]

\(U^*\) and \(V^*\) as predicted initially by the tridiagonal algorithm are updated point-by-point by the relations (for the \(k\)th iteration)

\[
U_{i,j}^{k*} = (\alpha_N U_{i,j+1}^{k*} + \alpha_S U_{i,j-1}^{k*} + \alpha_E U_{i+1,j}^{k*} + \alpha_W U_{i-1,j}^{k*} - T_{PU})/(-\alpha_p)
\]

\[
V_{i,j}^{k*} = (\alpha_N V_{i,j+1}^{k*} + \alpha_S V_{i,j-1}^{k*} + \alpha_E V_{i+1,j}^{k*} + \alpha_W V_{i-1,j}^{k*} - T_{PV})/(-\alpha_p)
\]
where \( T_{PU} \) and \( T_{PV} \) are terms not dependent on the changing values \( U^{k*} \) and \( V^{k*} \). The terms \( T_{PU} \) and \( T_{PV} \) are given by

\[
T_{PU} = U_{i,j} + FV_{i,j} - Z_{i,j}(P_{i+1,j} - P_{i-1,j})/2
- GZ_{i,j}(H_{i+1,j} - H_{i-1,j})/2 + \beta_{O_{i,j}}W_{i,j}
\]

\[
T_{PV} = V_{i,j} - FU_{i,j} - Z_{i,j}(P_{i,j+1} - P_{i,j-1})/2
- GZ_{i,j}(H_{i,j+1} - H_{i,j-1})/2 + \beta_{R_{i,j}}W_{i,j}
\]

(39) (40)

for the \( k+1 \)st iteration, the recursion relations for updating \( U^{k*} \) and \( V^{k*} \) are

\[
U_{i,j}^{k+1*} = U_{i,j}^{k*}(1 - \beta_R) + \beta_R U_{i,j}^{k**}
\]

\[
V_{i,j}^{k+1*} = V_{i,j}^{k*}(1 - \beta_R) + \beta_R V_{i,j}^{k**}
\]

(41) (42)

where \( \beta_R \) is the overrelaxation parameter (a value of 1.5 for \( \beta_R \) makes the procedure analogous to Newton's method). The relaxation scheme is carried across each row and row-by-row up the grid until (41) and (42) have been evaluated for all but the outermost columns and rows. The convergence test to be applied is that the r.m.s. value of the residuals \( U_{i,j}^{k**} - U_{i,j}^{k*} \) and \( V_{i,j}^{k**} - V_{i,j}^{k*} \) must fall below a specified limit.

The convergence criterion used in this study is that the r.m.s. \( U^* \) and \( V^* \) residuals given above must not exceed a flux magnitude equivalent to a current of 0.0001 m sec\(^{-1}\) in water of a depth equal to the r.m.s. value of the depth \( D \), taken over the whole grid. The convergence test used is thus

\[
\text{Residual}(U^*, V^*) \leq 0.01% \times D_{\text{rms}} \times \Delta t/(\Delta s)^2.
\]
To summarize the computing sequence, for the \( n+1 \)st time step, the fields of \( U \) and \( V \) are given a rough prediction by rows by the formulae (24) to (31). These preliminary values are refined by the pointwise overrelaxation scheme (41) and (42) using coefficients defined in (32) to (40) until the r.m.s. change in the new \( U \) and \( V \) fields from the \( k \)th to the \( k+1 \)st iteration is below a given tolerance. The new free surface for each wet point is then computed for time step \( n+1/2 \) by the relation

\[
H^{n+1}_{i,j} = H^{n-1}_{i,j} - [(U^{n+1}_{i+1,j} + U^n_{i+1,j}) - (U^{n+1}_{i-1,j} + U^n_{i-1,j})]
\]

\[
+ [(V^{n+1}_{i,j+1} + V^n_{i,j+1}) - (V^{n+1}_{i,j-1} + V^n_{i,j-1})]/4,
\]

where the superscripts refer to the time step. For dry points, the free surface is computed by

\[
H^{n+1/2}_{i,j} = \frac{1}{C} \sum_{k=-1}^{1} \sum_{\ell=-1}^{1} H^{n-1/2}_{k,\ell}
\]

(43)

where \( C \) is the number of wet points involved in the average. Points \((k,\ell)\) which are dry are not included in the average. If the \( H \) value found by (43) is less than the topographic height, \( D_{i,j} \), the point remains dry. In the foregoing discussion of equations (32) to (42) the superscript \( n \) is implied for all unstarred quantities.

Some investigators (Platzman, 1958, 1963; Miyazaki, 1965) use a completely centered-difference scheme, with \( H \) computed on odd-numbered steps on odd-numbered rows.
and columns and the flow field computed on even-numbered
time steps and grid positions. Such methods gain computing
speed but sacrifice resolution. The mixed system of
forward, centered, and one-sided differencing in the
present study is not as symmetrical as a pure centered-
difference scheme, but it is used under the assumption
that if the differential-equation model is a good simulation
of a well-behaved process, the finite approximation
thereof will be well-behaved as well. It is the intent
of the mixed-difference scheme to use the best and most
recent data possible for each step of the solution.

Conditions for stability.

In oceanographic problems, stability conditions
correspond to a relation between some characteristic
dimension of the modelled basin and the space and time
steps.

Stability of a computational scheme is tested by
harmonic analysis (von Neumann criterion) \( U, V, \) and \( H \)
assumed to be of the form

\[
(U, V, H) = (U_0, V_0, H_0) \exp[\sigma t + i(kx + \lambda y)].
\]

This is substituted into the equations of motion. In the
test functions the wave numbers \( k \) and \( \lambda \) represent wave-
lengths not shorter than \( \Delta s \) and the characteristic
frequency \( \sigma \) must be such that the product \( \sigma \Delta t \) is not
greater than unity. Attempts at actual calculation of the
stability criteria for the set of equations used in the
present study result in a very complicated cubic equation to relate $\sigma$ to $k$ and $\lambda$. The present investigation involves the von Neumann stability criterion which is well-known through the work of Lax and Richtmyer (1956), Platzman (1958, 1963), and Miyazaki (1965). Although their derivations did not include the advective terms, their results are applicable at least as guidelines for this study.

The common criterion for computational stability is that an error is not amplified from one time step to the next. From the studies cited above, the von Neumann criterion becomes

$$(\Delta s/\Delta t)^2 \geq 2gd_{\text{max}},$$

i.e., the speed of a free gravity wave in the deepest water in the model must be less than about 70% of a velocity scale $\Delta s/\Delta t$. This will restrict the maximal time step allowed for a particular grid spacing or the spatial resolution allowed for a particular time step. In order to increase resolution in space or decrease computing time by lengthening $\Delta t$, some sacrifice must be made in another parameter.

An assumption which may increase the value of $\Delta t$ allowed is that the basin depth $d_{\text{max}}$ is limited to the depth of the thermocline and that the flow under study is confined to the surface mixed layer above the thermocline. Phillips (1966) gives a thickness for this layer
of about 100 meters. This value will be adopted for the present study. Although 100 meters makes a rather shallow basin, the depths at which important storm-caused erosion occurs are generally much shallower, so no serious errors are likely to result from such an assumption. The author recognizes the existence of internal gravity waves on the thermocline or any such interface (Galt, 1971), but the choice for this study of a single-layer flow makes the assumption of a stationary thermocline necessary.
CHAPTER III

METEOROLOGIC CONDITIONS

The foregoing discussion of numerical forecasting of storm surges is generally applicable to any set of meteorological conditions and any body of water. In order to have some basis for testing the results of the surge forecasts, the area of the world to be modelled in this study is the continental shelf area of the Gulf of Mexico. Theoretical studies and sources of data include Miyazaki (1965), U.S. Army Corps of Engineers (1970), Murray (1970a, b), and Hayes (1967). Integration of actual storm data with a surge-forecasting model, however, has met with some difficulty. Miyazaki found it necessary to smooth the wind field measured for hurricane Carla of 1961 in order to keep solutions stable. For the present study, a numerical model for a hurricane wind field is used, based on modifications of the SPH (Standard Project Hurricane) (Graham and Nunn, 1959) by Collins and Viehman (1971).

The model hurricane.

Hurricane wind fields are divided into three regimes along any radius. Innermost, at the center of the quiescent eye, there is no wind. As the outer two-thirds
of the eye region are crossed, the wind speed builds approximately linearly from zero to a maximum in the towering eyewall clouds. From the maximum speed region outward, the winds fall off gradually, approximating an irrotational vortex with wind speed decreasing inversely with distance. The radially-symmetric wind field of a stationary hurricane may be described by (adapted in MKS units from Collins and Viehman, 1971)

\[
W_r = V_{\text{max}} 0.14213623 r^{-0.15128} \times \log_e(1.149 \times 10^3 R/r^{1.607})
\]  \hfill (44)

for \( r \geq R \),

\[
W_r = V_{\text{max}} (3r - R)/2R
\]  \hfill (45)

for \( R \geq r > R/3 \), and

\[
W_r = 0
\]  \hfill (46)

for \( R/3 \geq r > 0 \),

where \( R \) is the radial distance from the storm center to the maximum wind speed \( V_{\text{max}} \), and \( W_r \) is the wind speed at distance \( r \) from the storm center.

For a storm moving at speed \( V_H \), the speed relations (44) to (46) are modified to include the translation of the whole wind field through geometric relations depicted in Figure 2, as follows: The heading of the storm along its track is at the angle \( \delta \) relative to North. The SPH wind field as used by Collins and Viehman converges toward the storm center at an angle of 25° to tangency. Thus, the point where the forward motion of the storm and
Fig. 2.--Geometry of the hurricane model. Storm center is located as hurricane symbol, moving toward compass heading $\delta$. Maximum wind speed is distance $R$ from storm center. Representative point $P$ has wind $W$ as determined by equation (47).
the hurricane wind field add as parallel vectors lies
115° to the right (CW) of the storm heading, on a line
pointing toward
\[ Y = \delta + 115°. \]
A general point P lies on a heading \( \alpha \) from the storm center,
displaced from the heading from the storm through \( V_{\text{max}} \) by
the angle
\[ \theta = \alpha - Y. \]
At any heading away from \( Y \), the wind speed is reduced
as given by
\[ W_{r,\theta} = W_r - V_H (1 - \cos \theta)/2. \]  \hspace{1cm} (47)
The full wind speed of \( V_{\text{max}} \) is usually measured in a
moving storm by radar or aircraft and is assumed to
include the forward speed of the storm. The forward-
motion correction for a storm to be modelled obviously
would be simply \( V_H \cos \theta \) if forward speed were not included
in the estimate of \( V_{\text{max}} \). The heading of the winds at
P is 25° to the left of tangency or 115° CW of the
radius from the storm center, i.e., at heading
\[ \beta = \alpha - 115°. \]
The wind components at P are then
\[ W_x = W_{r,\theta} \sin \beta \]
and
\[ W_y = W_{r,\theta} \cos \beta. \]
The pressure field of the hurricane is assumed to
be in cyclostrophic balance (Saucier, 1955) with the wind
field, with a theoretical reduction of pressure in the eye given by

$$\Delta p_o = 0.02828428 (V_{max}^2 + fR V_{max})$$  \text{(48)}

where $\Delta p$ is in millibars, $V_{max}$ is in m sec$^{-1}$, and $R$ is in meters. Outside the eye the atmospheric pressure increases toward normal as

$$\Delta p_r = \Delta p_o /[1 + (r/R)^2]^{1/2}.$$  \text{(49)}

It is assumed that for the eye ($r \leq R$) the pressure remains constant at the value found by (49) for $r = R$. For very severe, compact hurricanes the central pressure drop is slightly overestimated by (48) and (49), but the resulting error in the "inverted barometer effect" is at worst only a few centimeters of water. In fact, the Coriolis term in (48) contributes to the pressure drop only minimally unless the storm has a very large eye. Note that the central eye pressure estimate is given by (49) with $r = R$ and not by (48).

**Numerical application of the model hurricane.**

The above hurricane model supplies the wind and pressure fields for the surge-forecasting procedure. In actual use with the computer program given in Appendix C the storm parameters are entered as initial storm location, forward speed of the storm, radius to maximum wind speed, magnitude of maximum wind speed, and storm heading. Each time the program calls for the storm to be moved, the wind
and pressure fields are recomputed for the new storm coordinates, according to (44) to (47) and (49). The storm is advanced along heading $\delta$ a distance of $V_H t$ times the model time elapsed since it last moved.
CHAPTER IV

COMPUTATION OF BOTTOM CURRENTS

The groundwork has now been laid for an examination of the actual agents responsible for shallow-water erosion during hurricanes. As a development of methodology for evaluating potential hurricane hazards to coastal activities, the present study would best examine the problem of estimating the maximum damage likely for a given area and set of storm parameters.

Erosive currents.

It is assumed here that erosive bottom currents consist of two components: the vertically-uniform surge currents derived from the volume transport due to pressure gradients and wind shear, and oscillatory currents due to waves superimposed on the surge (Bodine, 1971). The oscillatory currents are to be computed for the top of the boundary layer at the bottom so that maximum magnitudes of bottom-eroding currents can be identified. For engineering applications, the maximum current at the surface is more important than bottom currents in structural analysis. It is further assumed that maximum bottom currents occur where the drift current and wave current are parallel, although neither current is
necessarily parallel to the wind, and may, in fact; run opposite the real surface current (Murray, 1970b).

Wave forecasting methods.

There are many available choices among wave forecasting methods from which the erosive oscillatory currents may be calculated. There are graphical methods for use with meteorologic and hydrographic charts, e.g., Pierson, Neumann, and James (1955) and methods which lend themselves better to direct numerical computation of wave parameters, e.g., Bretschneider (1957, 1966). Most of these wave forecasting procedures are concerned with the so-called significant wave, the average of the highest one-third of the waves measured (Kinsman, 1965). It is not the province of this study to choose whether the significant height and its associated bottom current is a better predictor of erosive capability than the r.m.s. wave height, the average of the highest one-tenth of the waves, or the arithmetic mean wave height. The significant height is the figure most often quoted and is used here.

The wave forecasts in this study are made numerically from equations adapted from Bretschneider (1957) for selected points on the computer-grid representation of the coastal area. Bretschneider (1957, 1966) presents equations for significant waves for the SPH in deep water (there is a misprint in the form of two missing decimal
points on p. 147 of the latter reference) which give an exponential dependence of significant wave height on central pressure drop and radius to maximum wind. These formulae seem to lose generality in shallow water or where part of the storm wind field lies over land. Therefore, the fetch-dependent significant wave formulae from Bretschneider (1957) are used. For the present study it is assumed that the winds have been blowing for a sufficiently long time to develop a fully aroused sea. With MKS coefficients, they are

$$H_o = 7.63 \times 10^{-4} (\bar{w}^2 F)^{\frac{1}{4}}$$  \hspace{1cm} (50)

and

$$T_s = 0.1055 (\bar{w}^2 F)^{\frac{1}{4}}$$  \hspace{1cm} (51)

where $H_o$ is the significant wave height in meters, $T_s$ is the period of the significant wave in seconds, $F$ is the wind fetch in meters, and $\bar{w}^2$ is the mean-squared wind speed component parallel to the fetch in meters per second.

**Numerical wind-wave model.**

The geometric relation for finding the wind components and fetch are shown in Figure 3. For the point of interest located at $P(X_p, Y_p)$, the wind fetch vector $\underline{F}$ is extended tangent to the circle of maximum winds CCW of the storm center at $P'(X'_w, Y'_w)$. $P'$ lies along a compass bearing

*Vectors are denoted here by an underline.*
Fig. 3.—Geometry for wave-hindcast model. For hindcast point $P$, fetch line $PP''$ extends from $P$ tangent to maximum wind radius circle at $P'$ and beyond to edge of grid or low-wind cutoff point $P''$. Waves are computed relative to wind component along fetch toward $P$. 

$\bigcirc = \text{Point } (X_N, Y_N) \text{ for } N = 30$
of $B_p$, from P. The tangent bearing $B_p$, is found by first
determining the bearing $B_S$ of the storm center $O(X_H, Y_H)$
from P by

$$B_S = \tan^{-1}\left[\frac{(X_H - X_p)}{(Y_H - Y_p)}\right].$$

For a maximum wind radius of $R$ and a storm centered distance
D from P, the tangent $PP'$ points toward bearing

$$B_p = B_S - \tan^{-1}\left(\frac{R}{S}\right)$$

where the tangent length $PP'$ is

$$S = \left(\frac{D^2 - R^2}{2}\right)^{\frac{1}{2}}$$

and the storm distance

$$D = \left[\left(\frac{X_H - X_p}{2}\right)^2 + \left(\frac{Y_H - Y_p}{2}\right)^2\right]^{\frac{1}{2}}.$$ 

If $D \leq R$, the point P lies inside the storm eye and the
angle $B_p$, is obviously undefined, but it is also true that
the seas become calm in the eye of the storm, so no problems
arise here--the waves need not be computed in such a case.

The tick marks along $F$ divide the distance $PP'$ into
20 equal segments of length $\Delta F$. The line $PP'$ is extended
until it hits an edge of the computer grid; the last
increment of fetch inside the grid ends at $P''$, $N_o$ incre-
ments from P. For the $N^{th}$ step along $F$ the wind at point
$(X_N, Y_N)$ is $W_N$, and its component along $F$ toward P is
given by

$$\frac{W_N \cdot F}{|F|} = W_N \sin \theta_N$$

(52)

where

$$\theta_N = |\theta_N - B_p| - 90^\circ$$
and $\theta_N$ is the heading of $\mathbf{W}_N$. If $\theta_N$ is negative, the wind component along $\mathbf{F}$ at that point is away from $P$ and is ignored in the computation of the r.m.s. wind since most (but not all) wave energy would propagate away from $P$.

The value of $\overline{W^2}$ to be used in (50) and (51) is then

$$\overline{W^2} = \sum_{N=1}^{N_0} (W_N \sin \theta_N)^2 / N_0$$

(53)

where $N_0$ is the number of fetch increments from $P$ to $P''$.

The machine computations start at point $P$, which for simplicity is assumed to be located on a grid point, and step along $\mathbf{F}$ until the edge of the grid is reached at $P''$ or $W_N$ falls below 5% of $V_{\text{max}}$, in which case the fetch terminus $P''$ would be defined as the last point at which $W_N$ was above the stated minimum. This would include cases in which the winds at the far end of the fetch turn to point away from $P$. If the surge conditions have caused $P$ to become dry, the wave computations are skipped for that point.

The points 0 and $P'$ fall wherever the storm model dictates, so it is necessary to compute the wind components at $(X_N, Y_N)$ by averaging the four wind-component values at the corners of the grid cell containing $(X_N, Y_N)$.

Since all quantities used by the computer are nondimensionalized, the coordinates $(X_N, Y_N)$ may be rounded directly to the numbers of the nearest grid row and column, e.g., $[X_N]$ and $[X_N] + 1$ denote the grid columns on either side of position $X_N$. The heavy brackets denote the integer part [largest integer not exceeding] of $X_N$. Thus, a
simple linear interpolation of the wind components at $(X_N, Y_N)$ is done according to the following:

\[
Q_{X_N, Y_N} = (1 - X_N \times [X_N]) \times (1 - Y_N \times [Y_N])
\]

\[
\times [Q_{X_N + 1, Y_N}] \times [Q_{X_N, Y_N} + 1] \times (Y_N - [Y_N])
\]

\[
\times (X_N - [X_N]) \times (1 - Y_N \times [Y_N]) \times [Q_{X_N + 1, Y_N} + 1]
\]

\[
+ (Y_N - [Y_N]) \times [Q_{X_N + 1, Y_N + 1}]
\]

(54)

An identical interpolation formula is used for $R_{X_N, Y_N}$.

The local wind heading, $\beta_N$, is given by

\[
\beta_N = \tan^{-1}(Q_{X_N, Y_N} / R_{X_N, Y_N})
\]

The mean-squared wind to be used in (50) and (51) is computed by (52) to (54); the fetch length $|\mathcal{E}|$ is given by

\[
|\mathcal{E}| = [(X_{NO} - X_P)^2 + (Y_{NO} - Y_P)^2]^{1/2}
\]

where $(X_{NO}, Y_{NO})$ are coordinates of $P$.

Motions at the sea bottom.

The model is now fully defined for computing the water motions at the sea bottom. As stated above, the assumption is made that the maximum bottom current arises where the surge drift and oscillatory wave motions add as parallel vectors. From linear wave theory, the current at the bottom of shallow water under a wave reaches
a periodic peak amplitude of

\[ u_b = \frac{A \zeta}{c} \]  \hspace{1cm} (55)

where \( A \) is the wave amplitude, \( c \) is the phase velocity corresponding to depth \( \zeta \), and \( u_b \) in MKS units is in m sec\(^{-1}\). In shallow water,

\[ c = (g\zeta)^{\frac{1}{2}}. \]  \hspace{1cm} (56)

The wave amplitude to be used is either the maximum wave impinging on \( P \) or the maximum non-breaking wave allowed for the water depth \( \zeta_{X_P, Y_P} \), whichever is the lesser. The limiting wave height is generally assumed to be about 78% of the water depth (Munk, 1949). Street and Camfield (1967) give a slightly lower ratio of height to depth of 73% for a horizontal bottom. Putting Munk's depth limitation into (55) and substituting \( c \) from (56) gives

\[ u_b = \frac{1}{2} \tilde{H}_o (g/\zeta)^{\frac{1}{2}} \]  \hspace{1cm} (57)

where

\[ \tilde{H}_o \leq 0.78\zeta, \]

or

\[ u_b = 0.396 (g/\zeta)^{\frac{1}{2}} \]  \hspace{1cm} (58)

where waves of

\[ \tilde{H}_o > 0.78\zeta \]

would break offshore of the point of interest. The incoming wave spectrum obviously will be distorted by breaking waves, but a strong component of period \( T_s \) will remain (Bretschneider, 1961a, b). Combining the nondimensional parameters derived in Chapter II with (57) and (58) gives the
dimensional current maximum of surge drift plus wave motion as

\[ u_{b_{\text{max}}} = \frac{1}{2} \min(0.78Z_p \Delta s, \bar{H}_o) \times (G/Z_p)^{\frac{1}{2}}/\Delta t \]

\[ + \left[ (U_p^a + V_p^a)^{\frac{1}{2}}/Z_p \right] \times \Delta s/\Delta t \]

where the subscript \( P \) implies quantities evaluated at \( (x_p, y_p) \).

Problems arising in using the forecasted waves to estimate sediment movement.

The foregoing development of the surge drift currents gives no vertical profile for the distribution of horizontal motions, hence any discussion of bottom sediment movement must be limited to at most a model based on the assumed form of the bottom stress. To include sediment transport in the overall surge model would involve changing the contours of the bathymetry in response to sediment displacement at each time step. While this procedure obviously would simulate more closely the natural system, it also would involve a detailed description of the set of pertinent sedimentological parameters.

Only the most cataclysmic storm would produce erosion on a scale visible on a coarse mesh of data points. The mesh needed to study the effects on a scale of meters is much too fine to use for predicting the surge drift—the number of time steps required would be immense due to the stability limitations noted in Chapter II. For
small-scale analysis, it would be appropriate to employ either a patched grid or second computer run on a fine grid using larger-scale predicted surge conditions as initial and boundary values. Both methods carry a penalty of computer run-time increases.

The wave computation method presented in the previous section is a numerical adaptation of graphical methods and does not include the effects of refraction and convergence of wave orthogonals, angle of wave approach, or the divergence of the drift currents. Graphical methods of computing wave refraction effects are discussed in Pierson, Neumann, and James (1955). Waves passing into progressively shallower water turn toward a normal approach to the beach, hence any angle-dependent effect is reduced for most of the Gulf of Mexico coastline. Phillips (1966) notes that divergent currents cause a significant reduction in wave height. All of these effects are highly dependent on the integrals of several quantities along the wave ray passing through the point of interest.

The processes by which material is removed, transported, and redeposited on the sea bottom depend so strongly on local conditions that they, too, are difficult to include in a general model. Proper treatment of erosion and deposition requires a detailed description of the materials making up the bottom sediments, including such parameters as the distribution of sizes and densities of the sediment grains and their bed form (ripple marks,
grading, etc.). Inferences such as were drawn by Harms (1969) could be used to test the validity of a chosen erosion model, but the greatest applicability of the numerical analysis would seem to be in predicting the location of washover facies and damage to coastal structures.

**Engineering applications.**

There is much engineering concern with respect to hurricane waves and surges impinging on oil drilling and production platforms and other offshore structures. As a result there has been an intensive data-gathering effort in the oil industry, the results of which unfortunately remain largely proprietary. Data currently available to the public are listed in Thrasher and Aagaard (1969).

The greatest proportion of wave-force studies in the literature treats forces on cylindrical pilings by computing the sum of inertial and drag forces for a given wave. Aagaard and Dean (1969) compute the necessary velocity profiles by fitting stream functions to measured wave profiles. Evans (1969) and Wheeler (1969) both examine the contribution to total force of a spectrum of waves. Any of these methods could be applied to analysis of a given structure given the waves predicted by the numerical method of the present study, or any of the more detailed methods appropriate to a special
case. Surge and wave forces for a hypothetical test case are evaluated in the next chapter.
CHAPTER V

APPLICATION OF THE MODEL

The numerical model was applied to a hindcast of the response of the northern Gulf of Mexico to hurricane Camille for the six-hour period 1700 to 2300 CDT 17 August 1969. During this period the storm moved from near the tip of the Mississippi delta to the shore at Gulfport, Mississippi. The numerical grid is shown in Figure 4, superimposed on a map of the Gulf coast from central Louisiana to the Florida panhandle. The grid spacing is 15 km and extends E-W over 30 points and N-S over 24 points. The course of the storm is indicated by the circles with the period modelled indicated by the broken-line arrows. The track of the storm was determined from U.S. Weather Bureau advisories (U.S. Army Corps of Engineers, 1970), using the stated latitude and longitude of the eye.

The storm surge, waves, and drift currents are computed for an approximate representation of part of a real body of water, using a mathematical approximation to a highly variable meteorological phenomenon. For treatment of a special case in which the hurricane parameters and local sediment properties are well defined, the computer program developed here is capable of accepting actual
Fig. 4.—Track of hurricane Camille as given by U.S. Weather Bureau advisories. Numerical-model period is indicated by broken arrows. Dot grid for computation points has 15 km spacing. Shoreline follows Hydrographic Office maps 1115-1116.
measured data without major programming changes. Each special case would likely require a particular hurricane model or raw measured data, a bottom-stress drag coefficient which best fits the region of interest, perhaps a refracted-wave path, and many, many hours of card punching.

**Initial and boundary conditions for the model—comparison with actual conditions.**

At the start of the model period (1700 CDT), the entire model water mass was at rest \((U = V = 0)\) everywhere and in hydrostatic equilibrium with the barometric pressure field of the storm. The storm wind field and pressure field were as determined by the model in Chapter III.

The assumed conditions must be contrasted with the actual weather at the time. At 1700 CDT the storm center had been within 300 km of the delta area for about six hours, gale-force winds had been blowing over the southern part of the grid area for half a day, and hurricane-force winds were felt in the delta since 1500 CDT.

In addition to barometric response, the approaching storm had already piled up as much as 1.5 m of water in the coastal bays north and east of the delta. Current speeds inferred following Murray (1970a) were on the order of 2 to 3 m sec\(^{-1}\) near the windward shores. The winds at the shoreline west of the Mississippi passes were from the east to north, or offshore at all times during the model period. The Corps of Engineers report does not contain
any tide data for the southern Louisiana shores, but the
given high water mark of 1.22 meters at Grand Isle pro-
bably was reached while the winds were onshore early in
the afternoon of 17 August 1969.

The model hurricane is based on the SPH model, and
as it is therefore constructed according to averaged
storm geometry, does not exactly duplicate the storm
data in the U.S. Weather Bureau advisories. The advisories
give the radius of hurricane-strength winds (33.5 m sec\(^{-1}\))
as 97 km and gales (20 m sec\(^{-1}\)) to a radius of over 300 km.
The numerical model sets the hurricane wind radius at 70 km
and the gale radius at 170 km. Thus, the model storm
was somewhat more compact than the real one. Peak winds
for the storm were estimated as high as 100 m sec\(^{-1}\), but
no anemometer survived the strongest part of the storm.
Sustained winds were more likely of the order of the
model peak, \(V_{\text{max}}\), taken as 75 m sec\(^{-1}\). This peak wind
speed taken with the cyclostrophic pressure formula (48)
gives a central pressure of 898 mb for \(R = 15\) km; the
lowest recorded pressure in the real storm was 904 mb.
The difference in hydrostatic surge rise between the model
and actual central pressures is only 0.006 m. Wind and
pressure fields for the model storm at 1700 CDT are
shown in Figures 5 and 6.

Boundary conditions for the model require that the
free surface remain in hydrostatic equilibrium with the
storm pressure field at all wet points on the grid
Fig. 5—Isotachs (m/sec) for model hurricane at 1700 CDT. Arrows indicate wind direction at four points on 30 m/sec contour.
Fig. 6.--Isobars (millibars) for model hurricane at 1700 CDT. Minimum central pressure in model was 898.5 mb.
borders; the volume flux at the grid borders is zero everywhere. Because of the computer run-time consumed by the storm model, the storm wind field is advanced along the storm track only every third time increment. With a maximum depth of 100 m and a grid spacing of 15,000 m, the von Neumann criterion allows a time step of 180 sec; the storm therefore advances along its track in steps of 9 minutes of model time. From the reported storm track, the model storm advances at 8.33 m sec\(^{-1}\) toward a heading of 341° for the first two hours of the model period (1700 to 2300 CDT). From 1900 to 2300 CDT the storm moves due north at 6.25 m sec\(^{-1}\). The model storm moves a maximum of one-third of a grid step each time it moves, so little harm is done by not moving it every step. From Figure 4 it is evident that the real storm turned abruptly to a more westerly course after landfall, but this period was not modelled.

The model storm starts abruptly at model time \(t_0 = 1700\) CDT with the model Gulf of Mexico at rest. The highest initial barometric surge rise is under the storm eye and amounts to 1.63 m. Along the southern Louisiana coast computed barometric surface rises are not more than 0.35 m. Since the wind blows offshore at the start of the model, the water level is expected to drop steadily as the computations progress and the wind sweeps water away from the coast. In regions where the winds are onshore, the surface is expected to rise rapidly toward
dynamic equilibrium with the storm.

**Results of the model: surge heights.**

It is assumed here that due to the limited specification of the initial surge field, results from the leeward southern Louisiana coast must be ignored—the initial water level there was higher in reality than its starting barometric rise in the model. The discussion will concentrate on the windward shores from the Mississippi delta eastward. Figure 7 presents a comparison of the surge hydrograph measured by the tide gage at Shell Beach, Louisiana (U.S. Army Corps of Engineers, 1970) and the computed surge hydrograph for the nearest grid point. The model hydrograph nearly parallels the measurement, with the difference being largely made up of the difference in initial values. It is unfortunate that the other tide gages in the area were located on bayous and river channels not represented in the coarse 15 km grid. Correlations and tests of the model at other locations depended on observed high-water marks given by the Corps of Engineers compared with computed surge maxima. Figure 8 presents comparisons of computed and measured high water at 11 points. Note that points on the southern Mississippi delta (East Jetty, Venice, North Pass) showed little surge rise, a reflection of the initially understated surge and short time span of highest model winds there.

The development of the surge over the entire grid
Fig. 7.--Comparison of measured and computed surge hydrographs for Shell Beach, La. (grid column 11, row 18). Line marked "DIFFERENCE" indicates that model surge differed from tide-gage report by an amount near the difference in initial value.
Fig. 8.—Comparison of observed highest surges vs. computed maximum surges for nearest grid points. Broken line represents 1:1 correspondence. Points mentioned in text are East Jetty = 165,150; Venice = 180,180; North Pass = 195,180. Best fits are obtained at Dauphin Island, Ala. (300,300); Quarantine Bay, La. (180,195); Pascagoula, Miss. (270,330); and off Biloxi, Miss. at Ship Island (225,300).
is depicted in Figures 9 to 11, for times 2, 4, and 6 hours after starting the model. The computed surge rose ahead of the storm as expected. Large rises were also computed for the windward side of the Mississippi delta behind the storm, where the storm winds returned to parallel or onshore after storm passage. Position of the storm center is marked on each map by the hurricane symbol.

On the scale of the present model, the grid spans the width of many of the coastal barrier lagoons so that the surges seen of up to 7 m in the Gulfport, Mississippi area are not present in the model. The highest water at Gulfport was, however, reported after midnight, at least one hour after the end of the modelled interval. Peak surges may not have developed on the inner Mississippi coast before the end of the model.

**Currents and waves.**

Mean current velocities \((u^2 + v^2)^{1/2}\) are displayed for several points along coastal areas in Figures 12 to 14, at times corresponding to the surges shown in Figures 9 to 11. In general, the currents run less than 6% of the wind speed, although isolated points have extremely high--and probably spurious--velocities when the model water layer is extremely thin, e.g., a point in Caillou Bay, Louisiana, where the current ran 95% of the wind speed of 17.8 m sec\(^{-1}\) in water 0.20 m deep at 2000 CDT
Fig. 9.—Computed surge-height contours (m relative to MSL) for 1900 CDT, or 2 model hours. The storm has built up a surge of 3 m at the north side of the Mississippi delta. Apparent surface fall at near 90°W, 29°N reflects low initial model surge height relative to actual level.
Fig. 10.—Computed surge for 2100 CDT, or 4 model hours. Surge is still high on north side of delta, building to above 3 m in Chandeleur Sound. Low water at Gulfport, Miss. (coast near 89°W) may be artifact of low initial surge value in model.
Fig. 11.--Computed surge for 2300 CDT, or 6 model hours. Surge has decayed on delta and approaches maximum in Gulfport-Pascagoula area. Model surge peak trails storm center onto shore, as did real surge.
Fig. 12.—Current vectors for selected points, 1900 CDT. Length is proportional to speed according to scale shown. Currents generally follow wind direction in open water, but show strong offshore flow on Mississippi coast. Note onshore flow on north delta where surge is rising.
Fig. 13.—Current vectors for same points as Fig. 12, 2100 CDT. Offshore currents near Gulfport are turning shoreward as storm approaches and surge builds.
Fig. 14.--Current vectors, 2300 CDT. Currents near Mississippi shore are now strongly onshore as maximum surge builds.
in the model. This is probably an artifact of the inadequacy of the grid resolution there. Currents in the deep-water areas well offshore run 0.1 to 0.2 m sec\(^{-1}\), but still represent a considerable flux in a water layer 100 m thick, as modelled. It can also be seen from the current vectors that the direction of flow is nearly parallel to the wind and is largely controlled by surface stresses. In very shallow water, such as at Edgewater and Biloxi, Mississippi, the slope of the free surface indicates a pressure gradient that should be driving water inshore at 1900 CDT (Figure 12), but the currents point offshore, perhaps due to the wind stresses pulling water westward toward Lake Borgne, and, certainly, south into the approaching surge. At 1900 CDT, the current at Ship Island, 15 km offshore of Biloxi, is nearly parallel to the wind and follows the general deepwater pattern. As time advances, all the coastal currents turn from south through west to directly onshore as the surge maximum nears the coast at Gulfport (Figures 13 and 14).

Significant-wave hindcasts were made at 15 grid points in order to examine conditions given by the model which could be compared to high water marks and damage reports. Figure 15 shows the locations of the trial points. Wave and drift currents are given below for point #13, which is near Transworld Petroleum Rig 50. This rig measured winds of 172 mph (77 m sec\(^{-1}\)) and was one of a very few
Fig. 15.--Locations for trial wave hindcasts (numbered squares). Grid coordinates given as x and y distances from SW grid corner and as column and row numbers i, j. Points for hydrographs Figs. 7 and 16 are denoted by stars: Shell Beach (row 18, column 11) and Transworld Rig 50 (point 13, row 16, column 16).
offshore structures which survived passage of the storm center. Hindcast wave data and surge hydrograph are shown for this point in Figure 16. Maximum significant waves reached the breaking point at 2100 to 2200 CDT at just under 10 m height. No corroborating data were available for this study. A sample of wave hindcasts for all 15 points is presented in Appendix D along with all other pertinent parameters displayed as they came from the computer. Model time is 2100 CDT for the sample. In general, the hindcast waves are in good agreement with average hurricane conditions for period and amplitude, particularly with regard to a strong component of the wave spectrum near a period of 10 sec (Bretschneider, 1961b).

Resonance phenomena and stability of the model.

There are two types of resonance effects in the model—one is a physical process, the other is an artifact of the grid and numerical methods. As the storm surge approaches the coast and passes into shoaling water, it reaches a depth at which the free wave speed \((g\epsilon)^{\frac{1}{2}}\) equals the forward speed of the surge. In a real system, a hydraulic jump or bore should occur at this point, although no such process is inherent in the model. The forward speed of the storm was 6.25 m sec\(^{-1}\) at landfall. That speed corresponds to water of depth slightly under 5 m, such as might be found in the coastal lagoons.
Fig. 16.---Surge hydrograph and wave parameters for Transworld Rig 50. Significant wave height and period shown are for deep water waves, do not include shoaling or refraction effects. Breaker limit height is 78% of sum of MSL depth plus surge. Wave current is peak speed of wave oscillatory current at bottom due to indicated significant wave. Wave + drift current is magnitude of wave oscillatory peak added to surge drift speed, giving maximum expected current speed at bottom.
The highest computed surge occurs at a point with a model depth near this value, so the amplification of the moving surge as it enters shoaling water is correctly present in the model.

The second resonance operates over a longer time scale than the first but also concerns gravity waves. The model gives the arbitrary maximum depth of the open Gulf of Mexico as 100 m. The free wave speed for this depth is 31.3 m sec\(^{-1}\) or 113 km hr\(^{-1}\). At that speed, a gravity wave is able to traverse the north-south span of the model grid twice, and complete two round trips from the southern grid border to the Mississippi delta. As can be seen from Figure 11, there are places near the boundary where the free surface has been displaced over 8 m from sea level. This effect is most likely due to wavelike disturbances propagating from the border to the shore and back to interfere with other waves. The grid borders were not devised to absorb wave energy of this type, so continuance of the solution past the 120 steps used to model 6 hours would not be feasible without some means to damp this instability. While the von Neumann criterion is not violated there remains a limit to the duration of a model.

Geologic applications of the model results.

A detailed application of the model to a barrier island-lagoon system can forecast or hindcast erosion
and transport of bottom sediments given a textural description of the bottom or determine the limiting grain size that could be affected by a particular storm. For example, the maximum bottom current given in Figure 16 would be able to erode and transport any size particle up to at least 0.5 m diameter, while the maximum surge drift would be able to keep in motion particles up to about 0.05 m in diameter (Hjulström, 1935). From such figures it is apparent why Rodolfo, et al. (1971) found a marked increase in average grain sizes off North Carolina after hurricane Gerda, 1969. They suggested such storms as the principal mechanism for offshore transport of inner-shelf sediment.

From the wave and surge drift current maxima it appears that the drift current alone is not capable of eroding large particles and the waves, with their back-and-forth motion, are not capable of transporting large particles over any great distance. However, their combined effect can move material as described by Perkins and Enos (1968) and Ball, et al. (1967). In certain shallow-water conditions, such as the Caillou Bay case mentioned earlier, it may be possible that the very high wind-driven water current velocities computed actually exist and would act as powerful cutting agents, although it is much more likely that any fast-moving water would be in the form of wind-driven rain or spray.

The influx and efflux of the locally-large volume
of water represented by a storm surge passing into a barrier lagoon induce strong currents through tidal channels. The surge along the Mississippi-Alabama coast shown in Figure 10 covered some 900 km$^2$ with 3 m of extra water. This volume of water ($2.7 \times 10^9$ m$^3$) surely left some trace of its passage. Hayes (1967) reported that storm-surge cuts in barrier islands were soon healed by normal wave action, but the washover facies produced by hurricanes remain in the geologic record. Such a large input of sea water into coastal lagoons has had serious biological consequences, such as killing of fresh-water grasses and disruption of the hypersaline conditions in Baffin Bay, Texas (Behrens, 1969). Behrens also noted effects of fresh rainwater, but the present model excludes precipitation.

**Application to pollution control problems.**

The model can be adapted to tracer studies by following the trajectory of a marker particle. At each time step the particle position is advanced along a vector equal to the local mean velocity times the time step. Currents and trajectories computed from the model would indicate the movement of river discharge along shore, or of spilled oil from damaged offshore rigs and temporarily flooded coastal refineries. The environmental effects of storm flooding and erosion have been discussed in the preceding section.
Engineering applications: forces on structures.

Many offshore oil platforms were lost during hurricane Camille. One which survived was Transworld #50, off Chandeleur Island, Louisiana. Figure 16 gives the surge hydrograph, significant wave parameters, and currents calculated for the model storm; these give an indication of the forces this platform withstood. The crests of the hindcast significant waves plus the surge elevation reached a maximum height of about 7 m above MSL around the time 1930 CDT. The maximum current at that time was 5.53 m sec\(^{-1}\), of which 1.40 m sec\(^{-1}\) was due to surge drift. The period of significant waves was 12.06 sec. Taking as an example, a cylindrical piling on the platform substructure, the force on the piling due to water motions may be calculated by the approximate formula (Aagaard and Dean, 1969; Evans, 1969)

\[
\text{Force} \text{Unit length} = \rho_w C_D \frac{1}{2} u^2 |u| D + \rho_w C_M \frac{d u}{dt}
\]

where \(D\) is the cylinder diameter, \(u\) is the water velocity, \(C_D\) is the drag coefficient for a cylinder (= 0.5), and \(C_M\) is the virtual mass coefficient (= 1).

For the conditions at 1930 CDT, the force on a 1 m piling amounted to almost \(9.4 \times 10^3\) kg per meter of length at the Transworld platform. At that same time, a rig (reported destroyed) off North Pass, east of the tip of the delta, experienced a significant wave of 9.58 m, a period of 11.82 sec, a maximum current of
5.79 m sec\(^{-1}\), and a wave-oscillatory current peak of 3.58 m sec\(^{-1}\) near the bottom. These conditions would produce a force of about \(9.7 \times 10^3\) kg per meter of length.

It must be remembered that the wave-oscillatory currents computed here are for near the bottom of the water—still higher values would be reached at the surface. The water in the crest of a near-breaking wave moves at the local wave phase velocity, which was about 11.8 m sec\(^{-1}\) in water of the depth encountered at the Transworld rig. The force on the upper part of a vertical 1 m diameter piling would then be about \(40 \times 10^3\) kg per meter. For the second example platform the force at the wave crests would be some \(48 \times 10^3\) kg per meter.

The hindcast bottom currents coupled with an erosion model may also be of some value in analysis of scour around pipelines and foundations of offshore structures. Pipelines laid near the bottom, as in offshore tanker terminals, would be subjected to forces resulting from bottom currents of the type modelled. From the results of this study it appears that considerable forces may be exerted on pipes and pilings on the inner continental shelf during storms.
SUMMARY AND CONCLUSIONS

This study presents a general numerical model of wind- and pressure-driven storm surges and ocean currents for use in analysis of hurricane effects on coastal areas. The model is not restricted to any specific geographic scale or locality, nor to any specific scale or type of storm.

The method of computation employs well-known equations of motion cast in finite-difference form but uses a novel boundary condition at the model shoreline—a moveable rather than fixed boundary that may change the shape of the fluid region as the flow develops. This boundary treatment allows modelling of storm-surge flooding, including the development of currents in the flood waters. Solution of the equations is by a modified pointwise overrelaxation procedure.

Hurricane Camille, 1969, is examined as a test case, using numerical approximations for the storm wind and pressure fields. The modelled time period covers the six hours during which the storm center passed from the tip of the Mississippi delta to landfall. The numerical grid used covers 175,500 km², spanning the Gulf of Mexico coast from near Pensacola, Florida, westward to near Morgan City, Louisiana. Grid spacing is 15 km, making
a total of 720 points on the grid. A limiting depth of 100 m is assumed for the Gulf of Mexico, allowing a time step of 3 minutes for the model. Coupled with the surge and storm models is a wind-wave estimation procedure which supplies the magnitude of the maximum current produced by the combined effects of surge and waves.

Results of the test computations show agreement with some of the conditions reported from the real storm. Maximum computed surge heights compare favorably with reported debris strand lines, with many of the discrepancies between measurement and hindcast attributable to the stationary initial state of the model. Current speeds estimated from the surge model are of the same order of magnitude as the sole reported measurements of bottom currents during Camille. The lack of available wave, current, and surge data makes it difficult to check details of the model results, but it seems clear from the few published figures that the model is successful in simulating the overall response of the eastern Mississippi delta region to hurricane forces.

It is not the intent of this study to analyze in detail a particular event from either a geologic or engineering standpoint. No sedimentological parameters are included in the model, and only a sample structural-analysis computation is made to indicate how the results of a more detailed (smaller scale) model might be applied.
APPENDIX A

SYMBOLS USED

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APPENDIX B

DESCRIPTION AND USE OF PROGRAM STORM/SURGE

Program STORM/SURGE is written in the B-5700 COMPATIBLE ALGOL (XALGOL) language as implemented on the Rice University Burroughs B-5500 computer. A knowledge of this language is assumed for readers of the listing in Appendix C.

The program operates entirely in core while running, with long-term disk storage used for data fields which must be read point-by-point at the start of a run. I/O functions are designed to operate either on-line with card input and line-printer output or from a remote teletype terminal with teletype input and output or line-printer output. The teletype mode is useful mainly for debugging and is far too slow and costly for large runs. The largest data grid allowed by array dimensions in the program is 30 x 30 points. Data arrays on disk are stored in the whole-word "*" format by rows starting at the lower left (SW) corner of the grid.

Run-time divide checks are eliminated for dry points by the dummy array $S_{i,j}$ which takes the value 1 for a wet point ($Z_{i,j}$ positive) and 0 for a dry point ($Z_{i,j}$ zero). Divisions by $Z_{i,j}$ are done as divisions by
\[ Z_{i,j+1} - S_{i,j} \text{ so that the divisor is} \]
\[ Z_{i,j+1} - 1 = Z_{i,j} \]
for wet points and
\[ Z_{i,j}(=0) + 1 - 0 = 1 \]
for dry points.

The bathymetry array, \( D \), is entered as positive downward--land above sea level carries a minus sign. All data fields entered onto the disk files are in MKS units, except pressure, \( P \), which is in millibars.

Run-time control parameters set I/O mode (default value = on-line); indicate whether the run begins from zero or stored initial conditions (MODE); set starting time and number of steps in run (TO, NDT); grid dimensions (IX, JY), grid spacing and time increment (DS, DT); drag coefficient, Coriolis parameter for average latitude of grid, and acceleration of gravity (CD, F, G); storm parameters (XH, YH, VH, BIGH, VMAX, DELT); overrelaxation parameter (BETAR); convergence tolerance in \( \% \) and maximum iteration count (CRIT, LIMIT). After these parameters have been entered, the program is ready to receive map control parameters for initial condition maps and run steps (KODE). Consult the program listing, Appendix C, for input formats and see also the sample input "DATA FOR RUN..." following the listing for formats used.

The program listing contains explanatory notes about each PROCEDURE subprogram. The rows of asterisks (*)
are not included in the program deck, they merely separate listings of subprograms in the printout. The %-sign indicates a comment in XALGOL.
APPENDIX C

LISTING OF WORKING FORM OF PROGRAM STORM/SURGE BY P.W. SLOSS, 1972

PAGE 1

BEGIN

FILE VARS DISK SERIAL (2,30,600)
PARAMETERS P, Q, R
FILE PARS DISK SERIAL (2,30,600)
PARAMETERS N, P, Q, R
FILE INPT DISK SERIAL (2,30,600)
PARAMETERS PTR, N
FILE OUT DISK SERIAL (2,30,600)
PARAMETERS N
FILE CODE DISK SERIAL (2,30,600)
PARAMETERS N

ARRAY

U, V, H, S, D, P, Q, R, UNEW, VNEW, MINDRIFT, HNEW (1:30,1:30)
ARRAY

U, V, U, U, V (1:30)
ARRAY MAPCODE (1:9)

DEFINE

VA, UB, UC, UO (1:30)

WRITE LOCATION OF STORM CENTER IN X, Y VALUES

READ DATA FROM DISK FILE BY THIS PROCEDURE

PROCEDURE TRIU(L, A, X, Y);
PROCEDURE PUNK(X,RESVMS,CRT); REAL RESVMS,CRT; INTEGER K,FORWARDJ

PROCEDURE RELAX(SCRIPT, CRT,LIMIT);

REAL BETAR, CRT;

BEGIN
FOR J=2 TO N DO
  IF S(I,J) = 0 THEN GOTO L90;
  UNEM(I,J) = UNEM(I,J) + 1;
  GO TO L90;

L90: IF S(I,J) = 0 THEN GOTO L17;
  OME(I,J) = OME(I,J) + 1;
  GO TO L90;

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* * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * *

INTEGER L;
ARRAY A,B,C,D [111];

BEGIN
ARRAY GAMMA, BETA [111];
BETA(I) = B(I);
GAMMA(I) = GAMMA(I); I = 2100 BEGIN
BETA(I) = B(I) - A(I-1) * B(I) / BETA(I-1);
GAMMA(I) = GAMMA(I) - D(I) * A(I-1) / BETA(I)
END UNTIL I = 0 GTR L;
D(L) = GAMMA(L);
K = 100 BEGIN
I = L - I;
D(I) = GAMMA(I) - C(I) * D(I+1) / BETA(I)
END UNTIL K = 10 GTR L = 1;
END;

* * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * *

PROCEDURE RELAX(SCRIPT, CRT,LIMIT);

% THIS PROCEDURE RELAXES THE U AND V ARAYS POINTWISE BY ROWS, ALLOWING UP TO LIMIT NUMBER OF ITERATIONS TO CONVERGE TO RMS RESIDUALS CRIT. FAILURE TO CONVERGE IN LIMIT # ITERATIONS CALLS PROCEDURE "PUNT" WHICH DUMPS PRESENT AND PAST VALUES OF U, V AND THE RUN.

REAL BETAR, CRT;
INTEGER LIMIT;

BEGIN
REAL SX,SY,AP, AN, AS, AL, T, TP, DV, DHY, VPSTAK, UPSTAR, RESU, RESV, RESUI, RESIDV, RESRMSJ;

LABEL L9, L5, L10, L17, L90, QUIT;

K = 11;
L10: RESIDUI = RESIDV = 0;

J = 2300 BEGIN
  IF S(I,J) = 0 THEN GOTO L17;
  OME(I,J) = OME(I,J) + 1;
  GO TO L90;

L17: SX = (2 * SL1 + SLJ) / S(I-1,J) / 2;
  SY = (2 * SJ1 + SJ2) / S(I,J-1) / 2;
  AP = ((UI) * SX * SC1 + SJ1 * SX + V1) * (SJ1 = SL1 + SJ1)
  ASY = ((UI) * SX * SC1 + SJ1 * SX + V1) * (SJ1 = SL1 + SJ1)
  AN = (SJ1 * SX * SC1 + SJ1 * SX + V1) * (SJ1 = SL1 + SJ1)
  AE = (SJ1 * SX * SC1 + SJ1 * SX + V1) * (SJ1 = SL1 + SJ1)
  AN = (SJ1 * SX * SC1 + SJ1 * SX + V1) * (SJ1 = SL1 + SJ1)
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TPU1=U(I,J)*FXV(I,J)-7(I,J)*X(I-1,J)+P(I,J)*P(I-1,J)*P(N+1,J)*Q(I,J)*WINN(I,J);
TPV1=V(I,J)*FXU(I,J)-2(I,J)*P(I,J)+P(I+1,J)*P(I,J)-P(I,J)*WINI(I,J);
DHDX=0.5*Z(I,J)*(H(I+1,J)-H(I,J))*S(I,J)+H(I,J)*H(I,J)*ST(I,J)+S(I,J);
DHY=0.5*Z(I,J)*(H(I,J)-H(I-1,J))*S(I,J)+H(I,J)*H(I,J)*ST(I,J)+S(I,J);
UPSTAI=(ANVU+UL(I,J)*AE+UVNI+IL(I,J)*AS*VNI)/UPAI;
VPSRAI=(ANVU+UL(I,J)*AE+UVNI+IL(I,J)*AS*VNI)/VPSRAI;
RESU1=UPSTAI-VNEM(I,J);
RESV1=VPSRAI-VNEM(I,J);
RES1=RESU1+RESV1;
RESD1=RESU1+RESV1;
VNEM(I,J)=UVNI(I,J)*(1+ETAR)+ETAR*UPSTAI;
VNEM(I,J)=UVNI(I,J)*(1+ETAR)+ETAR*VPSRAI;
L001 END UNTIL I=1+GTH INX2;
END UNTIL J=J+1 GTH JMAX;
% WORKINGPOINTS = NUMBER OF POINTS IN ARRAYS, EXCLUDING BORDERS.
% RESMNS=RECIPE OF RES1=RESD1/WORKINGPOINTS;
IF RESMNS LEQ CRT ON LSS LIMIT THEN GO TO L41;
PUNI(K,RESMNS,CRTI)= GO TO QUIT1;
L41 IF RESMNS LEQ CRT THEN GO TO L51;
K=K+11;
GO TO L101;
L51 WRITE(DUPTL(F4),"","CONVERGENCE IN "+" ITERATIONS"),K);
QUIT1 END;

PROCEDURE MAPS(KODE);
% THIS PROCEDURE PRODUCES LINE-PRINTED MAPS OF THE SELECTED ARRAYS, AS
% CONTROLLED BY THE RUN-TIME PARAGRAM "KODE";
% KODE = FIELD PRINTED =
% 0 RETURN TO MAIN PROGRAM FOR NEXT TIME STEP
% 1 U-FLUX IN SQ METERS PER SEC
% 2 V-FLUX IN SQ METERS PER SEC
% 3 H IN METERS
% 4 J IN METERS
% 5 PRESSURE IN MILLIGARS
% 6 WIND IN METERS PER SEC
% 7 WIND SPEED IN METERS PER SEC
% 8 WIND DEPTH IN METERS
% 9 LATEST VALUES OF U-FLUX IN SQ M PER SEC AT STEP
% 10 WHEN SOLUTION FAILED TO CONVERGE
% 11 LATEST VALUES OF V-FLUX IN SQ M PER SEC AT STEP
% 12 WHEN SOLUTION FAILED TO CONVERGE
% 13 VBAR IN M PER SEC
% 14 VBAR IN M PER SEC
INTEGER CODE1
BEGIN
  ALPHA ARRAY TITLE(1:2,1)
  INTEGER I,J
  ARRAY ARRAY(1:30,1:30)X(1:1)
  REAL CT,TH,TM,TS,YJ
  INTEGER IPAGE(1:12)
  INTEGER ITH,ITM
  LABEL L299,L300,L305,L320,L125,ENDALL,L330,L200
  IF CODE LEQ 0 THEN G0 TO ENDALL
  TITLE(1:1)="TITLE(2:1)="
  % SELECT ARRAY TO BE PRINTED
  CASE CODE=1 OF BEGIN
    BEGIN % CODE = 1
      CI=DS*2/DTJ; J=1JD0 BEGIN I=1JD0 BEGIN
      ARRAY(I,J)=U(I,J)*CI
      END UNTIL I=I*1 GTR IXJ
      END UNTIL J=J*1 GTR IYJ
      TITLE(I,J)="U-FLUX"
      GO TO L200
    END;
    BEGIN % CODE = 2
      CI=DS*2/DTJ; J=1JD0 BEGIN I=1JD0 BEGIN
      ARRAY(I,J)=V(I,J)*FJ
      END UNTIL I=I*1 GTR IXJ
      END UNTIL J=J*1 GTR IYJ
      TITLE(I,J)="V-FLUX"
      GO TO L200
    END;
    BEGIN % CODE = 3
      J=1JD0 BEGIN I=1JD0 BEGIN
      ARRAY(I,J)=IF S(I,J) NEQ 0 THEN H(I,J)*DS ELSE 100;
      END UNTIL I=I*1 GTR IXJ
      END UNTIL J=J*1 GTR IYJ
      TITLE(I,J)="FREE SM"
      TITLE(2,J)="FC (H)"
      GO TO L200
    END;
    BEGIN % CODE = 4
      J=1JD0 BEGIN I=1JD0 BEGIN
      ARRAY(I,J)=M(I,J)*S
      END UNTIL I=I*1 GTR IXJ
      END UNTIL J=J*1 GTR IYJ
      END;
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TITLE(1)="MATHYM"
TITLE(2)="ENTRY"
GO TO L2001
END
BEGIN % CODE = 5
J1=1;UN BEGIN
I1=1,J00 BEGIN
ARRAY(I1,J1)=1013.25+P(I1,J1)
END UNTIL I1=I+1 GTR IX
END UNTIL J1=J+1 GTR JY
TITLE(1)="PRESSU"
TITLE(2)="R";
GO TO L2001
END
BEGIN % CODE = 6
C1=D5/D1
J1=1,J00 BEGIN
I1=1,J00 BEGIN
ARRAY(I1,J1)=RA(I1,J1)X PI
END UNTIL I1=I+1 GTR IX
END UNTIL J1=J+1 GTR JY
TITLE(1)="W=EST"
TITLE(2)="WIND"
GO TO L2001
END
BEGIN % CODE = 7
C1=D5/D1
J1=1,J00 BEGIN
I1=1,J00 BEGIN
ARRAY(I1,J1)=RA(I1,J1)X PI
END UNTIL I1=I+1 GTR IX
END UNTIL J1=J+1 GTR JY
TITLE(1)="H=OUT"
TITLE(2)="H Wind"
GO TO L2001
END
BEGIN % CODE = 8
C1=D5/D1
FOR J1=1 STEP 1 UNTIL JY DO BEGIN
FOR I1=1 STEP 1 UNTIL IX DO BEGIN
ARRAY(I1,J1)=IND(I1,J1)X PI END
TITLE(1)="WINN S";
TITLE(2)="PEE">
GO TO L2001
END
BEGIN % CODE = 9
J1=1,J00 BEGIN
I1=1,J00 BEGIN
ARRAY(I1,J1)=Z(I1,J1)X SN
END UNTIL I1=I+1 GTR IX
END UNTIL J1=J+1 GTR JY
TITLE(1)="WATER"
TITLE(2)="DEPTH"
GO TO L2001
END
BEGIN  % KODE = 10
C=0.5*2/DTJ
J=1:JDO BEGIN
I=1:IDO BEGIN
ARRAY(I,J)=V*N(W(I,J)*C)
END UNTIL I=I+1 GTR IX
END UNTIL J=J+1 GTR JY
TITLE(1)="U-FLUX"
TITLE(2)="(NEW)"
GO TO L2001
ENDJ
BEGIN  % KODE = 11
C=0.5*2/DTI
J=1:IDO BEGIN
I=1:IDO BEGIN
ARRAY(I,J)=V*N(K(I,J)*C)
END UNTIL I=I+1 GTR IX
END UNTIL J=J+1 GTR JY
TITLE(1)="V-FLUX"
TITLE(2)="(NEW)"
GO TO L2001
ENDJ
BEGIN  % KODE = 12
C=0.5/DTJ
J=1:JDO BEGIN
I=1:IDO BEGIN
ARRAY(I,J)=U(K(I,J)*S(I,J)/(Z[I,J]+1-S[I,J]))
END UNTIL I=I+1 GTR IX
END UNTIL J=J+1 GTR JY
TITLE(1)="URAN "
GO TO L2001
ENDJ
BEGIN  % KODE = 13
C=0.5/DTI
J=1:JDO BEGIN
I=1:IDO BEGIN
ARRAY(I,J)=V(E(I+J)*C*S(I,J)/(Z[I,J]+1-S[I,J]))
END UNTIL I=I+1 GTR IX
END UNTIL J=J+1 GTR JY
TITLE(1)="VRAR "
GO TO L2001
ENDJ
END
% BEGIN PRINTING
L2001
IPAGE=0
I=1
% FOUT = 01 PRINT 6 COLUMNS PER PAGE ON TWX
% FOUT = 11 PRINT 12 COLUMNS PER PAGE ON LINE PRINTER (PRN)
I2=5+6*FOUT
L2901
IF FOUT EQ 1 THEN WRITE(OUTPUT(FOUT)(PAGE))
WRITE(OUTPUT(FOUT),"\"AT OF \"A6, TITLE(1), TITLE(2)\"
TH=T MOD 3600)
TMS=T MOD 3600
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ITHI=INTEGER(THI)
ITHI=ITHI MOD 60
ITM=INTEGER(TMS/T5)/60)
WRITE(OUTPUT(OUT1),"TIME="I3,H","I3,H","F5.1","SECS."")
ITHI=ITHI+TS)
IPAGE=IPAGE+1)
WRITE(OUTPUT(OUT1),"PAGE",I4,"//",IPAGE)
IF (PAGE EQL 1 THEN WRITE(OUTPUT(OUT1),"Y(KM)"/")
J1=1JD0 BEGIN
K1=J1+1
IF PAGE EQL 1 THEN GO TO L3051
IF FOUT EQL 0 THEN BEGIN
WRITE(OUTPUT(OUT1)<6F10.3X6",""FOR I1=I1 STEP
1 UNTIL 12 ON ARRAY(I1,K))
END)
IF FOUT EQL 1 THEN WRITE(OUTPUT(OUT1)<12F10.3X6",""FOR I1=I1 STEP
1 UNTIL 12 ON ARRAY(I1,K))
IF FOUT EQL 1 THEN WRITE(OUTPUT(OUT1))
GO TO L3001
L3051
Y=0.5X(K-1)x.001
IF FOUT EQL 0 THEN WRITE(OUTPUT(OUT1)<6F10.3X6",""Y FOR I1=I1 STEP
1 UNTIL 12 ON ARRAY(I1,K))
IF FOUT EQL 1 THEN BEGIN
WRITE(OUTPUT(OUT1)<12F10.3X6",""Y FOR I1=I1 STEP 1 UNTIL
12 ON ARRAY(I1,K))
WRITE(OUTPUT(OUT1))
END)
L3001
END UNTIL J1=J+1 GTR JY
I1=I1JD0 BEGIN
K1=I1+1
X(I1)=YX(K-1)x.001
END UNTIL I1=I1 GTR I12)
IF PAGE EQL 1 THEN GO TO L3201
WRITE(OUTPUT(OUT1)<12F10.3X6",""FOR I1=I1 STEP 1 UNTIL
12 ON XXX(I1))
GO TO L3251
L3201
WRITE(OUTPUT(OUT1)<"X(KM)=I1F10.3","FOR I1=I1 STEP 1 UNTIL
12 ON XXX(I1))
L3251
IF PAGE NEQL 1 THEN GO TO L3301
I1=01
L3301
I1=I1+6X(FOUT+1))
I2=I2+6X(FOUT+1))
IF 12 LEQ 1 THEN GO TO L2991
II=I1X1)
IF II GEQ 12 THEN GO TO ENDA1)
GO TO L2991
ENDALL WRITE(OUTPUT(OUT1)(PAGE)) ; END)

* * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * *
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BEGIN

PROCEDURE PUNT(KRESMS,CRT)

REAL RESMS,CRT;
INTEGER KJ;
BEGIN
WRITE(OUTPUT,14)<"SOLUTION FAILED TO CONVERGE IN"I4>
"ITERATIONS"I4,KJ);
WRITE(OUTPUT,14)<"RESIDUAL RMS = "E11,4", CRITICAL"=
E11,4,RESMS,CRT);)
MAPS(1);)
MAPS(10);)
MAPS(2);)
MAPS(11);)
MAPS(3);)
WRITE(OUTPUT,14)<"PROCESS TIME = "F8,2" SECS","TIME(2)/60);)
PUNTFLAG := TRUE;
END;

PROCEDURE WINUYJ

BEGIN
DEFINE V1 = VMAX**(0.1221367)WLN(1198(SMALLR+(-0.6077))*
/RGR/SMALLR)/SMALLR+0.15128 #,
V2 = VMAX**(3*SMALLR - RGR)*0.5/GR #;
REAL PROCEDURE ATAN2(X,Y);
REAL X,Y;
BEGIN
REAL PHI,PSI,PI;
PI := 3.1415927;
IF Y EQ 0 THEN IF X GE 0 THEN PHI := 1.5707963 ELSE PHI := 1.5707963*
ARCTAN(Y/X) ));
ELSE PSI := PHI+PI ELSE PSI := PHI+PI
ATAN2 := PSI;
END;
REAL SMALLR,WFR,ALPH, BETA, GAMMA, THETA,X,Y;
INTEGER I,J;
GAMMA := DECT + 2.00713;
XYMAX := XH + 3RGR*STN(GAMMA)I;
YYMAX := YH + 3RGR*COS(GAMMA)I;
FOR J=1 STEP 1 UNTIL JY DO BEGIN;
FOR I=1 STEP 1 UNTIL IX DO BEGIN;
X := I-I;
Y := J-J;
SMALLR := SQRT((X-XH)**2+(Y-YH)**2);
ALPH := ATAN2((X-XH),(Y-YH));
THETA := ALPH - GAMMA;
END;
END;
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BETA = ALPH = 2.00713
IF SMALLR LSS HIGH THEN IF SMALLR GEO BIGR/3
THEN VOFR1 V2 ELSE VOFR1 = 0
ELSE VOFR1 = V1
WIND[1,J] = WOFH + 0.5*(1-COS(THETA))
IF SMALLR LSS HIGH THEN SMALLR = BIGR
P[1,J] = -NP/SQR(T1 + (SMALLR/BIGR)+2)
Q[1,J] = WIND[1,J]*SIN(BETA)
R[1,J] = WIND[1,J]*COS(BETA)
WIND[1,J] = ABS(WIND[1,J])
END
ENDJ ENDJ

REAL PROCEDURE MNSPH(I,J)
% FIND RMS WATER DEPTH OVE WHOLE ARRAY FOR DETERMINATION OF VON %
% NEUANN CONVERGENCE CRITERION.
ARRAY D(I+1)
BEGIN
INTEGER I,J, REAL SUM
SUM = 0
FOR J=1 STEP 1 UNTIL JV DO BEGIN
FOR I=1 STEP 1 UNTIL IX DO BEGIN
SUM = SUM + D(I+J+2)
END
END
DRMS = SQRT(SUM/(IX*JY))
END

PROCEDURE HOMDER(I)
BEGIN
% SET H ON GRID HOMDER TO HYDROSTATIC BALANCE WITH PRESENT PressURES.
INTEGER T,J,FREAL CMH
CMH = 0.009907/0.75
FOR I=1 STEP 1 UNTIL IY DO BEGIN
H[I,J] = IF U[I,J]+(HN+WI)=CHXP[I,J] GTR 0 THEN HNOW ELSE H[I,J]
H[I,J] = IF U[I,J]-(HN+WI)=CHXP[I,J] GTR 0 THEN HNOW ELSE H[I,J]
END
FOR J=1 STEP 1 UNTIL JMAX DO BEGIN
H[I,J] = IF U[I+1,J]+(HN+WI)=CHXP[I,J] GTR 0 THEN HNOW ELSE H[I+1,J]
H[I+1,J] = IF U[I+1,J]-(HN+WI)=CHXP[I,J] GTR 0 THEN HNOW ELSE H[I+1,J]
END
ENDJ ENDJ

REAL PROCEDURE NINEPOINTAVG(ARG[I,J])
% COMPUTE AVERAGE VALUE OF PARAMETER "ARG" FOR A NEAREST WET POINTS %
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********** PROBLEM 1 **********
********** SURROUNDING POSITION I,J **********
********** Prog

ARRAY ARG(I, J); INTEGER I, J

BEGIN
  INTEGER M, N; REAL SUMOF, SUMARG;
  SUMARG = SUMOF; = 0;
  FOR M = 1 STEP 1 UNTIL 1 DO BEGIN
    FOR N = 1 STEP 1 UNTIL 1 DO BEGIN
      SUMOF := SUMOF + S(I + M, J + N);
      SUMARG := SUMARG + ARG(I + M, J + N) * S(I + M, J + N);
    END;
  END;
  IF SUMOF = 0 THEN SUMOF = 1;
  V_INITAVG := SUMARG / SUMOF;
END;

********** PROCEDURE DUMP_DISK **********

PROCEDURE DUMP_DISK BEGIN
  % AT END OF RUN TRANSFER FINAL VALUES IN DIMENSIONAL FORM TO DISK
  % FILES FOR USE AS INITIAL CONDITIONS IN NEXT RUN.
  INTEGER I, J; REAL NSVT, DSOT;
  DEFINE DISKOUT(WHERE, WHAT) = BEGIN
    FOR J = 1 STEP 1 UNTIL JY DO BEGIN
      WRITE(WHERE, FOR I = 1 STEP 1 UNTIL IX DO WHAT[I, J]);
    END;
  END;
  REAL PROCEDURE REDIMENSIONALIZE(x, const);
  ARRAY X(I, J) REAL CONST;
  BEGIN INTEGER I, J;
    FOR J = 1 STEP 1 UNTIL JY DO BEGIN
      FOR I = 1 STEP 1 UNTIL IX DO BEGIN
        X[I, J] = X[I, J] * CONST;
      END;
    END;
  END;
  DSOT := (NSVT = DS/DT) * 5;
  REWIND(PARS); REWIND(VARS);
  REDIMENSIONALIZE(U, US(VT)); DISKOUT(VARS, U);
  REDIMENSIONALIZE(V, DSOT); DISKOUT(VARS, V);
  REDIMENSIONALIZE(US, US(VT)); DISKOUT(US, H);
  REDIMENSIONALIZE(L, DX); DISKOUT(PARS, D);
  REDIMENSIONALIZE(W, DSOT); DISKOUT(PARS, W);
  REDIMENSIONALIZE(WSO, D); DISKOUT(PARS, R);
  WRITE(@PRINTOUT@, "FINAL VALUES HAVE BEEN TRANSFERRED TO DISK"
    " FILES AT STEPS " @TIME@); WRITE(@PRINTOUT@(@DNL) + PT1 @TIME@));
END;

********** PROCEDURE FIRSTITERATION **********

BEGIN

********** PROBLEM 1**********
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% APPLY TRIDIAGONAL ALGORITHM (PROCEDURE TRIDI) TO GET FIRST ESTIMATES
% OF NEW U AND V FIELDS.
%
LABEL L200, L300;

J1 = 2 J00 BEGIN
  I = 2 J00 BEGIN
    K = 1
    UA(K) = UB(K) = UC(K) = UO(K) = 0
    URA(K) = 1
    IF S(I,J) EQ 0 THEN GO TO L200
    SX = (2*S(I+1, J) + S(I-1, J))/2
    UB(K) = H(I+1, J) * SX(I, J) * S(I+1, J)
    UC(K) = H(I, J) * SX(I, J) / (2*S(I, J) + S(I+1, J) + S(I-1, J))
    DUAX(V) = (S(I, J) + V(I-1, J) * S(I-1, J) + S(I-1, J) * M(I-1, J)) / M(I-1, J)
    DUX(V) = S(I, J) * V(I-1, J) * S(I-1, J) * M(I-1, J)
    UDV(K) = UDV(V) + CD * 0.5 * S(I-1, J) * M(I-1, J)
    VDV(V) = V(I-1, J) * S(I-1, J) * M(I-1, J)
    VDV(K) = VDV(V) + CD * 0.5 * S(I-1, J) * M(I-1, J)
  END UNTIL I = J01 GR TMAX
  TRIDI(NEWX=UA, UB, UC, UD)
  END
  UNEW(V) = UDV(K)
  UNEW(U) = UDV(K)
  END UNTIL I = J01 GR TMAX
  J1 = 2 J00 BEGIN
  K = 1
  VA(K) = VV(K) = VC(K) = VD(K) = 0
  VBA(K) = 1
  IF S(I, J) EQ 0 THEN GO TO L300
  SX = (2*S(I+1, J) + S(I-1, J))/2
  VA(K) = U(I-1, J) * SX(I-1, J) * S(I+1, J) + S(I, J)
  VBA(K) = H(I, J) * SX(I, J) * S(I, J) * M(I, J)
  VDV(K) = VDV(V) + CD * 0.5 * S(I-1, J) * M(I-1, J)
  VDV(V) = V(I-1, J) * S(I-1, J) * M(I-1, J)
  VDV(K) = VDV(V) + CD * 0.5 * S(I-1, J) * M(I-1, J)
  END UNTIL I = J01 GR TMAX
  TRIDI(NEWX=VA, VB, VC, UD)
  END
  END
PROCEDURE SURFSUP;
%*****************************************************************************
% COMPUTE SIGNIFICANT WAVE HEIGHT AND PERIOD BY BRETSCHneider WINDs
% FETCH METHOD, COMPUTE MAXIMUM BOTTOM CURRENT SPEED BY ADOING WAVE
% CURRENT IN SURGE DRIFT.
%*****************************************************************************
BEGIN
READ FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY,
           FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY,
           FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY,
           FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY,
           FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY,
           FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY, FETCHY,
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MAXSURF)
WRITE(OPTFOUT),</"WAVES COMPUTED FOR FETCH OF","RA,3"," KM","/
" WIND R.M.S.=""R8,2"," M/SEC CALCULATED OVER","I5,
" POINTS""=FFIGH,4001,50R1(HMEANSQ),41
WRITE(OPTFOUT),</"SIGNIFICANT DEEPWATER WAVE"»,</
"X5,"SIGNIFICANT HEIGHT =","R8,2"," METERS PERIOD =","R8,2,
" SECONDS""=REALSURF1(50,131)
WRITE(OPTFOUT),</"NEAR SCILLATORY CURRENT =","R8,2"," M/SEC","SWASH")
WRITE(OPTFOUT),</"SWAVE DRIFT CURRENT + WAVE SWASH","R8,2,
" M/SEC","SWASH+DRIFT(X)*YP1)/(ZXP1*YP1*DOS)
WRITE(OPTFOUT,Fuhl,J"="**"*.*"*.*"*.*"*.*")
GO TO START1
DRYPOINT) WRITE(OPTFOUT),</"POINT (XP,YP) =","R8,3","R8,3,
" (X) IS DRY","(XP1)*XP1*DOS,001,(YP1)*YP1*DOS,001)
GO TO START1
INSIDEYE) WRITE(OPTFOUT),</"POINT (XP,YP) =","R8,3","R8,3,
" (X) IS INSIDE EYE OF STORM SIZE WAVES ISSUE CALM","(XP1)*XP1*DOS,001,(YP1)*YP1*DOS,001)
GO TO START1
RETURN: END1

START1 READ(OPTFIN)/*MOD1C131*/
%$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$
% $$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$
% ** MAIN PROGRAM **
% $$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$
% $%
% READ "MODE" FIRST TO SET PROGRAM FUNCTION
% $%
% IF MODE = 1 THEN COMPUTE WINDS AND PRESSURE
% $%
% IF MODE = 2 THEN USE PRESET INITIAL W,V,H,P,W,R,FIELDS
% $%
% MODE = 2 INDICATES RESTART USING SAVED U,V,W,R,FIELDS FROM DISK FILE
% $$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$$
% RESTARTING IS IF MODE 0 THEN TRUE ELSE FALSE
% IF RESTARTING THEN MODE = MODE - 4
% IF MODE EQL 1 OR MORE EQL 3 THEN FIN=01
% $% INPUT ON TMX
% GO TO INDX
% C11: FIN=FOUT=111
% $% I=0 ON CDR PTR
% GO TO START1
% L201: INDX IF MODE FWL 1 OR MORE EQL 3 THEN BEGIN
% WRITE(TMX,"ENTER OUTPUT OPTION 0 FOR TMX, 1 FOR PTR")
% IF FIN EQL 0 THEN READ(TMX,"FOUT")
% WRITE(OPTFOUT,FOUT,PTATME=20,40)
% $% PRINT ELAPSED PROCESS TIME WHEN THIS STATEMENT APPEARS
% WRITE(TMX,"ENTER TO AND NOT") END1
% $% READ MODEL START TIME (SEC) AND NUMBER OF TIME STEPS IN RUN
% READ(OPTFIN),/10,NOT11)
% IF NOT LESS 0 THEN OMPUTIONS)
% $% NOT = 0 FLAGS PROGRAM TO SAVE U,V,H,P,W,R ON DISK
% IF NOT = 0 THEN GO TO END1
% NOT = 0 MEANS RUN IS OVER--STOP
% IF FIN EQL 0 THEN WRITE(TMX,"ENTER IX AND JY")
LISTING OF WORKING FORM OF PROGRAM STORM/SURGE BY P.W. SLOSS 1972

% ENTER # COLUMNS AND ROWS IN ARRAYS
READ(INPFIN1,J1,J2,J3)
% IF FIN EQL 0 THEN WRITE(TM, "ENTER DS AND DT")
% READ(GRID STEP (METERS), TIME STEP (SECONDS))
READ(INPFIN2,L1,L2,L3)
% IF FIN EQL 0 THEN WRITE(TM, "ENTER CD, F, G")
% ENTER DRAG COEFF, CORIOLIS PARAMETER (UNITS PER SEC), GRAVITY (MKS)
READ(INPFIN3,L4,L5,L6)
IMAX=10+1
JMAX=10+1
NEQX=(JY-2)*NEQX
PTUNFLAG = FALSE
% GET DATA FROM DISK FILES
LREAD(VARS, UJ)
LREAD(VARS, VJ)
IF RESTARTING OR MODE GEC 3 THEN LREAD(VARS, HJ)
LREAD(PARS, DJ)
LREAD(PARS, PJ)
IF MODE GEO 3 THEN BEGIN
LREAD(PARS, QJ)
LREAD(PARS, RJ)
END;
CTS=(DTS-10)/DS
IF MODE LEQ 2 THEN BEGIN
% READ STORM PARAMETERS XM, YM, RIGR (KM), VMAX (M PER SEC), DELT (DEG)
READ(INPFIN4,L7,L8,L9)
RIGR = BIGR+1000
DP = 0.02828*S*(VMAX^2 + R*VMAX*RIGR)
WRITE(INPFOUT, "<CENTRAL PRESSURE = "VMAX, "M", ">1013.25-0.7071")
WRITE(INPFOUT, "<MP MAX = "MAX, "M/SEC AT"R, "DEGREES AT")
WRITE(INPFOUT, "<STORM MOVEMENT TOWARD"R, "DEGREES AT")
% NONDIMENSIONALIZE PARAMETERS FOR STORM MODEL
XM = XM/1000/DJ
YM = YM/1000/DJ
STORMCTR;
BIGR = RIGR/DSJ
VM = VM/KMDSJ
VMAX = VMAX/DSJ
DELT=DELT/0.1745329
VM = VM*SIN(DELT/2);
VY = VM*COS(DELT/2)
WINDY END;
% NONDIMENSIONALIZE ALL ARRAYS EXCEPT P
J1=100 BEGIN
J1=100 BEGIN
UI(J1)=UI(J1)*CTS
VI(J1)=VI(J1)*CTS
HI(J1)=IF NOT RESTARTING AND MODE LEQ 2 THEN
-0.009997*P(I,J)/OS ELSE HI(I,J)/DSJ
N(I,J)=N(I,J)/DSJ
IF MODE EQL 3 OR MODE EQL 4 THEN BEGIN
Q(I,J)=Q(I,J)/DSJ
RI(J1)=RI(J1)/DSJ
WIND(I,J)=SORT(Q(I,J)+2+HI(I,J)+2) END
LISTING OF WORKING FORM OF PROGRAM STORM/SURGE BY P.W. SLOSS 1972

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Z[I,J]=D[I,J]+H[I,J];
IF Z[I,J] GTR 0 THEN S[I,J]=1 ELSE S[I,J]=0;

$ S IS A FLAG ARRAY TO INDICATE #ET (S[I,J]=1) OR DRY (S[I,J]=0) POINTS
$ DRIFT[I,J]=59.3X(I[I,J]+2)*V[I,J]*2);
END UNTIL I=I+1 GTR 1X1;
END UNTIL J=J+1 GTR JY;

L100: G=6X(I*I+2)/551;
FI=FKT(J);

$ FACTOR TO CONVERT MTT TO NONDIMENSIONALIZED METERS OF WATER
$ PNO = 0.0049535X JT05+2.1
$ SET OUTER MORTAL POINTS OF UNEW ARRAY TO INITIAL U VALUES
$ SET OUTER MORTAL POINTS OF VNEW ARRAY TO INITIAL V VALUES
$ J=1 DO 200 BEGIN
UNE[I,J]=U[I,J];
UNE[I+1,J]=U[I+1,J];
VNE[I,J]=V[I,J];
VNE[I+1,J]=V[I+1,J];
END UNTIL J=J+1 GTR JY;
I=1 DO 200 BEGIN
UNE[I,J]=U[I,J];
UNE[I,J+1]=U[I,J+1];
VNE[I,J]=V[I,J];
VNE[I,J+1]=V[I,J+1];
END UNTIL I=I+1 GTR INAY;
200: NTIME=NTIME+1;
IF FIN EQL 0 THEN WRITE("*X","ENTER OVERRELAXATION PARAMETER")
READ(INPTFNUM); IF FIN EQL 0 THEN WRITE("*N","ENTER CRIT. VALUE AND ITEM. LIMIT")
$ ENTER CONVERGENCE TOLERANCE IN % OF VON NEUMANN LIMIT, AND ENTER
$ ITERATION MAXIMUM COUNT LIMIT
READ(INPTFNUM);
$ COMPUTE ACTUAL CONVERGENCE TOLERANCE RELATIVE TO DRMS
$ CRITERION = CRIT/DRMS(3)X DT05X0.01;
WRITE(9TFFINO1,J); "OVERRELAXATION PARAMETER =", R7.2,
"CONVERGENCE TOLERANCE =", P9.4,"%", OR ACTUAL VALUE =", R11.4,
"(NONDIMENSIONALIZED) =", BETA, CRIT, CRITERIOV)
GO TO L900; % INITIAL CONDITION MAPS
% RELAX U,V FIRST TIME BY WHOLE ROWS
L9000: FIRST ITERATION;
WRITE(9TFFINO1,J); "TIME(2)/600;
% RELAX U,V POINWISE UNTIL SOLUTION CONVERGES
RELAX(BETA, CRIT, INAY, LIMIT);
IF PUNTFLG THEN GO TO ENDFRIVJ;
% IF NO CONVERGENCE IN LIMIT ITERATIONS, STOP RUN
WRITE(9TFFINO1,J); "TIME(2)/600);
% COMPUTE NEW FIELD
% FOR JJ=2 STEP 1 UNTIL JMAX NO BEGIN
FOR JJ=2 STEP 1 UNTIL JMAX DO BEGIN
$ IF DRY POINT, CHECK NEIGHBORING POINTS FOR POSSIBLE NEW FLOODING
$ IF S[I,J] EQL 0 THEN NNEW[I,J]=0.5X NNEWPOINTAVG(N[I,J]) ELSE BEGIN
 FOR JJ=2 STEP 1 UNTIL JMAX NO BEGIN
FOR JJ=2 STEP 1 UNTIL JMAX DO BEGIN

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DIVY = ((V(I,J+1)+VNEW(I,J+1))*S(I,J+1) + (V(I+1,J)+VNEW(I+1,J))
   *(S(I,J+1)+S(I+1,J+1)) - (V(I,J)+VNEW(I,J))*S(I,J+1))
   *(2 - S(I,J+1)*S(I+1,J+1)) X(0.25)
DHTY = DIVX + DIVY

HNEW(I,J) = H(I,J) + DHTY

ENDJ ENDJ ENDJ

FOR II=1MAX STEP =1 UNTIL 2 DD BEGIN

FOR J=1MAX STEP =1 UNTIL 2 DD BEGIN

IF 3[I,J] EQ 0 THEN HNEW[I,J]=HNEW[I,J]+0.5*FININTAVG(H[I,J])


Z[I,J]=DI[I,J]+H[I,J]

SL[I,J]=IF Z[I,J] GTR THEN 0 ELSE 1

UL[I,J]=IF NEW[I,J] S[I,J]

VL[I,J]=VNEW[I,J]*S[I,J]

DRI[I,J]=SWAT(U[I,J]+2*V[I,J]*2)

ENDJ

% ADVANCE MODEL BY 1 TIME STEP

T=T+DT

NTIME=NTIME+1

L501: IF MODE EQ 1 OR MOD2 EQ 3 THEN WRITE(TW,/*"4MAP="))

% ENTER KOEF VALUES FOR EACH MAP FOR TIME UP TO 10 VALUES PER CARD OR TW

% VALUE OF ZERI (0) FLAG END OF MAP RUN, NEGATIVE VALUE OF KOE

% CALLS HAVE-HINGOCASE ROUTINE "SURF SUP"

READ(TMP1,F1,1 FOR I=0 STEP 1 UNTIL 9 DO MAPCODE[I])

FOR I=0 STEP 1 UNTIL 9 DO BEGIN

IF MAPCODE[I] LESS 0 THEN SGEIN SURF SUP GO TO L501; ENDJ

IF MAPCODE[I] EQL THEN GO TO DNO; ELSE MAPS(MAPCODE[I]); ENDJ

% PROGRAM LOOKS FOR ADDITIONAL MAP REQUESTS

GO TO L501

DNO: IF NTIME EQL 0 THEN GO TO L1000;

IF NTIME EQ 0 THEN GO TO L200;

IF (MODE EQ 1 OR MODE EQ 2) AND NTIME MOD 3 EQL 0 THEN BEGIN

% MOVE THE MODEL STORM EVERY 3RD TIME STEP TO SAVE RUNNING TIME

XH = XH + VX*3

YH = YH + VY*3

ENDFR

STORMCRAJ HBORDERJ ENDJ

GO TO L1000J

ENDDOFJOBEND.

* * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * *
DATA FOR RUN FROM T=4HRS=6HRS, INCLUDING WAVE HINDCASTS
PAGE 1

MODE FOR RESTART
RESTARTING TO, NOT AT 4 HRS (2100CDT)

DO WAVES
DOWNTOWN NEW ORLEANS
SW OF GRAND ISLE
JUST SW EAST JETTY
PORT EADS
HEAD OF ST LOUIS BAY
10 KM E OF GULFPORT (EDGEWATER)
BILOXI

DO WAVES
DOWNTOWN NEW ORLEANS
SW OF GRAND ISLE
JUST SW EAST JETTY
PORT EADS
HEAD OF ST LOUIS BAY
10 KM E OF GULFPORT (EDGEWATER)
BILOXI
DAUPHIN ISLAND
NEAR IRISH BAYOU, LAKE PONCHARTRAIN
INLAND OF CLERMONT HARBOR
RIGS (DESTROYED)
RIGS (DESTROYED)
GULF RIG (DESTROYED)
RIGS (DESTROYED)

RETURN

DO WAVES
DOWNTOWN NEW ORLEANS
SW OF GRAND ISLE
JUST SW EAST JETTY
PORT EADS
HEAD OF ST LOUIS BAY
10 KM E OF GULFPORT (EDGEWATER)
BILOXI
DAUPHIN ISLAND
NEAR IRISH BAYOU, LAKE PONCHARTRAIN
INLAND OF CLERMONT HARBOR
RIGS (DESTROYED)
RIGS (DESTROYED)
GULF RIG (DESTROYED)
RIGS (DESTROYED)

RETURN
DATA FOR RUN FROM T=0HRS-6HRS, INCLUDING WAVE HINDCASTS

PAGE 2

INLAND OF CLERMONT HARBOR
RIGS DESTROYED
RIGS (DESTROYED)
RIG (NOT DESTROYED, 172MPH MEAS)
GULF RIG (DESTROYED)
RIGS (DESTROYED)

RETURN *
0* 60 MIN
0* 63 MIN
0* 66 MIN
0* 69 MIN
0* 72 MIN
0* 75 MIN
0* 78 MIN
0* 81 MIN
0* 84 MIN
0* 87 MIN
0* 90 MIN
0*
3*12*13*5*6*7*R*1*
8 19
8 13
12 11
14 12
13 21
15 22
16 22
16 21
19 21
12 21
15 12
16 15
16 16
12 15
13 17
RETURN *
0* 90 MIN
0* 93 MIN
0* 96 MIN
0* 99 MIN
0* 102 MIN
0* 105 MIN
0* 108 MIN
0* 111 MIN
0* 114 MIN
0* 117 MIN
0* 120 MIN

DO WAVES
DOWNTOWN NEW ORLEANS
SW OF GRAND ISLE
JUST SW EAST JETTY
PORT EADS
HEAD OF ST LOUIS BAY
10 KM E OF GULFPORT (EDGEWATER)
BILoxi
DAUPHIN ISLAND
NEAR IRISH HAYOU, LAKE PONCHARTRAIN
INLAND OF CLERMONT HARBOR
RIGS DESTROYED
RIGS (DESTROYED)
RIG (NOT DESTROYED, 172MPH MEAS)
GULF RIG (DESTROYED)
RIGS (DESTROYED)

RETURN *
0* 90 MIN
0* 93 MIN
0* 96 MIN
0* 99 MIN
0* 102 MIN
0* 105 MIN
0* 108 MIN
0* 111 MIN
0* 114 MIN
0* 117 MIN
0* 120 MIN
0*
3*12*13*5*6*7*R*1*
8 19
8 13
12 11
14 12
13 21
15 22
16 22
21 21
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20 20
12 21
DATA FOR RUN FROM T=8HRS TO 6HRS, INCLUDING WAVE HINDCASTS
PAGE 3

15 12 RIGS (DESTROYED)
15 14 RIGS (DESTROYED)
16 16 RIG (NOT DESTROYED, 172 MPH MEAS)
12 15 GULF RIG (DESTROYED)
13 17 RIGS (DESTROYED)
RETURN
0,* 120 MIN
0,0,* FINAL CARD -- SAVE INITIAL CONDITIONS

* * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * * *
APPENDIX D

SAMPLE OUTPUT FROM COMPUTER

1. MSL depth, D. Data from U.S. Hydrographic Office maps 1115, 1116, and 1264 to 1275. Depths in meters.

2. Free surface, H. MSL height, meters.

3. UBAR current, m sec$^{-1}$.

4. VBAR current, m sec$^{-1}$.

5. Pressure, P, millibars.

6. Wind speed, WIND, m sec$^{-1}$.

7. Output of PROCEDURE SURFSUP--wave hindcast data.
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POINT (XP,YP) = (105,000, 270,000) IS DRY.

*ALLOWABLE BREAKER HEIGHT AT POINT (XP,YP) = (105,000, 180,000) IS 0.01 METERS

WAVES COMPUTED FOR FETCH OF 205.391 KM,
WIND RMS = 23.75 M/SEC CALCULATED OVER 40 POINTS

SIGNIFICANT DEEPMAR WAVE:
SIGNIFICANT HEIGHT = 8.22 METERS PERIOD = 10.95 SECONDS
PEAK OSCILLATORY CURRENT = 0.15 M/SEC
SURGE DRIFT CURRENT + WAVE SWASH = 1.02 M/SEC

*ALLOWABLE BREAKER HEIGHT AT POINT (XP,YP) = (165,000, 150,000) IS 2.81 METERS

WAVES COMPUTED FOR FETCH OF 184.419 KM,
WIND RMS = 28.87 M/SEC CALCULATED OVER 38 POINTS

SIGNIFICANT DEEPMAR WAVE:
SIGNIFICANT HEIGHT = 8.81 METERS PERIOD = 11.33 SECONDS
PEAK OSCILLATORY CURRENT = 2.32 M/SEC
SURGE DRIFT CURRENT + WAVE SWASH = 3.25 M/SEC

*ALLOWABLE BREAKER HEIGHT AT POINT (XP,YP) = (195,000, 165,000) IS 0.04 METERS

WAVES COMPUTED FOR FETCH OF 176.872 KM,
WIND RMS = 27.76 M/SEC CALCULATED OVER 44 POINTS

SIGNIFICANT DEEPMAR WAVE:
SIGNIFICANT HEIGHT = 8.91 METERS PERIOD = 11.40 SECONDS
PEAK OSCILLATORY CURRENT = 0.26 M/SEC
SURGE DRIFT CURRENT + WAVE SWASH = 1.40 M/SEC
ALLOWABLE BREAKER HEIGHT AT POINT (XP,YP) = (180,000, 300,000) IS 2.58 METERS

WAVES COMPUTED FOR FETCH OF 54.537 KM.
WIND RMS = 21.47 M/SEC CALCULATED OVER 21 POINTS

SIGNIFICANT DEEPWATER WAVE
SIGNIFICANT HEIGHT = 3.83 METERS PERIOD = 7.47 SECONDS

PEAK OSCILLATORY CURRENT = 2.22 M/SEC

SURGE DRIFT CURRENT + WAVE SMASH = 3.61 M/SEC

ALLOWABLE BREAKER HEIGHT AT POINT (XP,YP) = (210,000, 315,000) IS 0.47 METERS

WAVES COMPUTED FOR FETCH OF 319.045 KM.
WIND RMS = 13.24 M/SEC CALCULATED OVER 89 POINTS

SIGNIFICANT DEEPWATER WAVE
SIGNIFICANT HEIGHT = 5.71 METERS PERIOD = 9.12 SECONDS

PEAK OSCILLATORY CURRENT = 0.89 M/SEC

SURGE DRIFT CURRENT + WAVE SMASH = 2.43 M/SEC

ALLOWABLE BREAKER HEIGHT AT POINT (XP,YP) = (225,000, 315,000) IS 0.27 METERS

WAVES COMPUTED FOR FETCH OF 336.294 KM.
WIND RMS = 15.37 M/SEC CALCULATED OVER 86 POINTS

SIGNIFICANT DEEPWATER WAVE
SIGNIFICANT HEIGHT = 6.80 METERS PERIOD = 9.96 SECONDS

PEAK OSCILLATORY CURRENT = 0.71 M/SEC

SURGE DRIFT CURRENT + WAVE SMASH = 2.12 M/SEC

ALLOWABLE BREAKER HEIGHT AT POINT (XP,YP) = (300,000, 300,000) IS 2.13 METERS

WAVES COMPUTED FOR FETCH OF 359.455 KM.
WIND RMS = 22.11 m/sec calculated over 57 points

SIGNIFICANT DEEPWATER WAVE:

SIGNIFICANT HEIGHT = 10.11 meters PERIOD = 12.15 SECONDS

PEAK OSCILLATORY CURRENT = 2.02 m/sec

SURGE DRIFT CURRENT + WAVE SWASH = 3.21 m/sec

ALLOWABLE BREAKER HEIGHT AT POINT (Xp, Yp) = (120,000, 285,000) is 1.53 METERS

WAVES COMPUTED FOR FETCH OF 77.691 KM
WIND RMS = 0.00 m/sec calculated over 21 POINTS

SIGNIFICANT DEEPWATER WAVE:

SIGNIFICANT HEIGHT = 0.00 meters PERIOD = 0.00 SECONDS

PEAK OSCILLATORY CURRENT = 0.00 m/sec

SURGE DRIFT CURRENT + WAVE SWASH = 1.43 m/sec

ALLOWABLE BREAKER HEIGHT AT POINT (Xp, Yp) = (165,000, 300,000) is 1.15 = 0.03 METERS

WAVES COMPUTED FOR FETCH OF 58.193 KM
WIND RMS = 24.07 m/sec calculated over 21 POINTS

SIGNIFICANT DEEPWATER WAVE:

SIGNIFICANT HEIGHT = 4.43 meters PERIOD = 8.09 SECONDS

PEAK OSCILLATORY CURRENT = 0.05 m/sec

SURGE DRIFT CURRENT + WAVE SWASH = 0.75 m/sec

ALLOWABLE BREAKER HEIGHT AT POINT (Xp, Yp) = (210,000, 185,000) is 35.52 METERS

WAVES COMPUTED FOR FETCH OF 184.111 KM
WIND RMS = 27.33 m/sec calculated over 45 POINTS

SIGNIFICANT DEEPWATER WAVE:
significant height = 9.05 meters p period = 11.69 seconds

peak oscillatory current = 7.10 m/sec

surge drift current + wave swash = 2.92 m/sec

*allowable breaker height at point (xp,yp) = (210,000, 195,000) is 14.64 meters

waves computed for fetch of 181,348 km
wind rms = 28.67 m/sec calculated over 66 points

significant deepwater wave

significant height = 9.32 meters p period = 11.66 seconds

peak oscillatory current = 3.39 m/sec

surge drift current + wave swash = 4.56 m/sec

*allowable breaker height at point (xp,yp) = (225,000, 225,000) is 9.74 meters

waves computed for fetch of 228,025 km
wind rms = 27.21 m/sec calculated over 107 points

significant deepwater wave

significant height = 9.93 meters p period = 19.03 seconds

peak oscillatory current = 4.31 m/sec

surge drift current + wave swash = 4.77 m/sec

*allowable breaker height at point (xp,yp) = (165,000, 210,000) is 1.74 meters

waves computed for fetch of 121,236 km
wind rms = 33.18 m/sec calculated over 63 points

significant deepwater wave

significant height = 6.82 meters p period = 11.34 seconds

peak oscillatory current = 1.54 m/sec

surge drift current + wave swash = 2.98 m/sec

point (xp,yp) = (180,000, 240,000) is inside eye of storm ... waves assumed calm.
process time = 1707,00 secs.
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