
Abstract

An unexpected run-up of the ocean along Daytona Beach, Florida, on 3–4 July 1992 was associated with at least one large ocean wave. The wave, which reached a height of about 3 m above normal tide, injured 75 people and damaged property along Daytona Beach. Analyses of meteorological and oceanographic observations are consistent with the hypothesis that a squall line generated a long water wave. The critical evidence is that the propagation speed of the squall line matched the shallow-water wave speed that prevailed along the direction of motion of the squall line. The squall line exerted force on the ocean for at least 3 h. The issues of recurrence and public safety are discussed.

1. Introduction

On 4 July 1992, at about 0330 UTC, the ocean rose rapidly to a height of about 3 m above normal tide at Daytona Beach, Florida. The unexpected run-up, which lasted only a minute, caused at least 75 minor injuries to individuals and damage to several dozen vehicles on the beach.

The run-up was apparently associated with one or more ocean waves of unusually large amplitude that broke upon the shore. The only apparent proximate cause of such a wave event was the presence of a southward-propagating squall line located north of the region. The National Data Buoy Center (NDBC) buoys (41009 and 41010 in Fig. 1), which were in the vicinity, did not measure large ocean waves that could have accounted for the sudden run-up. An additional buoy (number 41006) ~200 km farther east also showed no significant wave action. Speculation that an underwater seismic event or a mudslide along the continental shelf caused the wave has no observational support (Sallenger et al. 1994).

Thieke et al. (1993) documented the run-up by interviewing Volusia County Beach rangers who were patrolling the beach over a north–south distance of 24 km. The rangers, whose vehicles were swept by the wave, reported that either a single large crest or a large crest followed by two smaller ones occurred. The wave, which first arrived on shore at the north end of Daytona Beach and progressed southward at a speed of 13–14 m s$^{-1}$, affected 48 km of shoreline. The event lasted 40–60 s at each affected location, and eyewitnesses (Nickerson 1993) estimated the height of the primary crest to be 2.4–3.0 m. Field surveys summarized by Thieke et al. (1993) found a maximum run-up height of 1.8 m over a region about 8 km in width; the run-up outside this region was 0.7 m. The 1.8-m run-up, indicated by water marks, was roughly half of the wave height of 3 m.

Large, unexpected waves from nonseismic sources have been documented before. During 26 June 1954, for example, a large wave that formed on Lake Michigan hit the Chicago, Illinois, waterfront. Six people fishing on a breakwater were swept away and drowned. Ewing et al. (1954) attributed the “abrupt increase in lake level” to at least one wave that reached a height of 2.4 m and that varied in height along the shoreline due to local shoaling and tunneling effects. The wave was associated with the passage of a squall line that had passed through the region 2 h earlier. The squall line had winds of about 27 m s$^{-1}$, and an abrupt rise in air pressure of 3.4 mb occurred within 5 min as the squall line apparently propagated through the region. Analysis of barograph traces from Illinois and Ohio indicated that the pressure disturbance was 241 km in length.

The theoretical link between squall line passage and the generation of large waves has been discussed...
by several authors. Proudman (1953) showed that waves of high amplitude would be expected when the wind speed matched the phase speed of a shallow-water wave. For a semi-infinite channel of constant depth $H$, resonant coupling between the air and water occurs when $V^2 = gH$, where $V$ is the constant wind speed at the surface and $g$ is the acceleration of gravity. Harris (1957) used additional data from tide gauges on Lake Michigan to estimate the heights of the waves that could have been generated from such a squall line. He found generally good agreement between theoretical and observed values but noted that shoaling, wave reflection, and convergence due to shoreline contours contributed to the observed magnitude of wave run-up. He concluded that “the available data and theory support the reasonableness of the hypothesis, but are not sufficient to establish it beyond a reasonable doubt.”

Platzman (1958) conducted a two-dimensional linear integration of the primitive equations to predict the rise associated with the June 1954 wave run-up event in Chicago. He presumed that the energy supplied to Lake Michigan was the combined result of the atmospheric pressure gradient and strong winds associated with the squall line. His simulated squall line was presumed to be straight, with a fixed length and width. He noted that the first significant rise along the southeast coast of Lake Michigan coincided with the passage of the squall line, while the reflected wave was responsible for the casualties in Chicago. In a subsequent paper, Platzman (1965) simulated a series of run-up scenarios for various positions and directions of movement of squall lines over the southern basin of Lake Michigan. These data were used to successfully predict a wave-related run-up in the region during 1960 (Irish 1965). The prediction was fairly accurate, except that the wave arrived sooner than expected. Irish attributed the early arrival time to the bow-shaped squall lines associated with this event. The positive results from these simulations in Lake Michigan led to the implementation by the U.S. Weather Bureau of operational warnings when run-ups from large waves are expected (Hughes 1965).

Whitham (1974) conceptually discussed the behavior of waves on water of uniform depth. Although the phase speed and group velocities of long waves both increase with increasing wavelength, the group velocity always has magnitude that is smaller than the phase speed. If a pressure disturbance (e.g., a squall line) is translating with speed $U$, it will excite shallow-water waves of phase speed $c_\omega$. If $U < c_\omega$, the water waves will propagate outward with respect to the disturbance. If $U > c_\omega$, then only transient waves are generated, which decay far from the disturbance. When $U = c_\omega$, however, the energy transferred from the atmosphere to the water cannot be radiated away from the atmospheric disturbance because there is no difference between the group velocity of the excited waves and the wind velocity that is generating it. In this case (Stoker 1957), the linearized equations of motion used to describe the evolution of the waves become invalid because a significant amplitude develops and the waves become nonlinear.

Akylas (1984) analyzed the development of forced long water waves using linear analysis and neglecting viscosity. He derived analytical solutions to the shallow-water equations, in the special case of $U = c_\omega$. In the linear case, waves decay exponentially with respect to distance behind the disturbance and experience a sinusoidal decay with respect to distance ahead of the disturbance. The theory, however, predicted that all the waves will increase in amplitude with respect to time, eventually becoming unbounded. Actually, the initially negligible nonlinear terms become dominant and limit the unbounded growth of the waves.

Akylas then presented a nonlinear analysis of the forcing of waves, which resulted in a forced Korteweg-de Vries (KdV) equation. The KdV equation, solved numerically, generated solitons. The generation of the solitons depended on the total force acting on the water surface and not on the “precise details” of the forcing.

Radar, satellite, and surface meteorological data indicate that prior to the arrival of the Daytona Beach wave, a squall line moved southward off the coast of southern Georgia and northern Florida. Squall lines are characterized by a series of convective cells that propagate at $\sim 15 \text{ m s}^{-1}$ relative to the ambient wind field. In radar and satellite imagery, squall lines often are arc shaped. The leading edge of squall lines is characterized by large gradients of radar reflectivity associated with the convection. The back side of squall lines often has a large area of stratiform precipitation, characterized by horizontally homogeneous radar reflectivity patterns (Houze 1993, chapters 8–9).

The following sections describe the sources of data and the oceanographic and meteorological characteristics of this event. A scale analysis is presented relating coastal ocean measurements to the atmospheric forcing. The final section includes a discussion of the likelihood of this type of wave event recurring at Daytona Beach.

2. Sources of data

a. Radar data

Radar reflectivity measurements were manually recorded at National Weather Service (NWS) offices...
TABLE 1. Characteristics of the WSR-57 radar.

<table>
<thead>
<tr>
<th>Characteristic</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Beam width</td>
<td>1 deg</td>
</tr>
<tr>
<td>Gain</td>
<td>38.5 dB</td>
</tr>
<tr>
<td>Wavelength</td>
<td>9.99 cm</td>
</tr>
<tr>
<td>Peak power</td>
<td>500 kW</td>
</tr>
<tr>
<td>Pulse length</td>
<td>4 µs</td>
</tr>
<tr>
<td>PRF</td>
<td>200–800 Hz</td>
</tr>
</tbody>
</table>

at Daytona Beach, Florida (DAB), and Waycross, Georgia (WYS). Both sites operated WSR-57 weather surveillance radars, the characteristics of which are listed in Table 1. The 10-cm wavelength radar provided useful values of radar reflectivity out to a range of 278 km (150 nm). For this study, hourly summaries of the two radars were examined. The summaries, known as radar logs, were manually prepared analyses of the radar echo intensities.

The radar logs consisted of manually gridded values of radar reflectivity from the video integrator processor (VIP), which yielded digital values in a range of 1–6. The radar reflectivity values associated with these VIP levels are given in Table 2. Analysts normally ignore echoes associated with anomalous propagation and ground clutter then record values of radar reflectivity on square grids, 46 km wide, covering the 278-km range from the radar. When an echo was present beyond that range, the analysts marked the appropriate grid box to indicate that an echo was present. No attempt was made to determine the VIP level at such ranges, since the radar beam spreads and samples higher altitudes with increasing range. In the analyses presented below, radar data were not analyzed beyond 278 km.

Though the resolution of the gridded radar data was coarse, a well-defined squall line having a distinct leading edge was seen for several hours as it propagated within range of the radars, prior to the wave run-up event. The two sets of radar logs have the same map projection and scale. Hence, there is unique and complete overlap among a subset of the 46-km analysis boxes sampled by the two radars. The area of overlapping radar coverage is indicated in Fig. 1. Fortuitously, the highest reflectivities during the lifetime of the squall line occurred within the region of overlapping coverage.

A composite radar echo map was generated each hour by taking the maximum radar reflectivity value in each of the boxes. Thus, the composite comprised single-radar samples within 278 km of each radar among the nonoverlapping areas, plus the maximum reflectivity value from either radar in the overlapping areas. The composite radar maps, discussed in subsection 2b, permitted analysis of the position and structure of a mesoscale squall line system.

During this period, a WSR-88D Doppler weather radar was also operating at the NWS office in Melbourne, Florida. Though radar reflectivity and radial winds measurements from this radar were not recorded, verbal reports from the meteorologist on duty, who watched the radar displays, are included.

b. Satellite imagery

Geostationary satellite images were examined on an hourly basis from 1800 UTC 3 July through 0800 UTC 4 July 1992. These images were provided by the National Hurricane Center (NHC). Visible imagery is useful until sunset between 2200 and 2300 UTC 3

**Fig. 1.** Daytona Beach and the South Atlantic Bight (SAB). The locations of the Daytona Beach, Florida, and Waycross, Georgia, NWS WSR-57 surveillance radars are indicated. Range rings are located at radius 278 km (150 nm). Bottom topography of the SAB is contoured in equivalent shallow-water phase speeds (m s⁻¹). Values exceeding 25 m s⁻¹, associated with deeper water to the east of the 25 m s⁻¹ contour, are not shown. The C-MAN station at Savannah, Georgia (SVLS1), and St. Augustine, Florida (SAUF1), are marked.

**TABLE 2.** Radar reflectivity values (dBZ) associated with VIP levels on the WSR-57 radar.

<table>
<thead>
<tr>
<th>VIP</th>
<th>dBZ</th>
</tr>
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<tbody>
<tr>
<td>1</td>
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<td>5</td>
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<tr>
<td>6</td>
<td>57</td>
</tr>
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</table>

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July. An additional infrared image from the Advanced Very High Resolution Radiometer (AVHRR) polar-orbiting satellite was obtained at 2336 UTC 3 July, as the satellite flew over the region.

**c. Oceanographic observations**

Two NDBC Coastal-Marine Automated Network (C-MAN) stations were operating in the region. The St. Augustine, Florida, C-MAN station (SAUF1 in Fig. 1) is located on St. Johns County Pier, 8 m above mean sea level (MSL). The other is located on an offshore platform near Savannah, Georgia (SVLS1 in Fig. 1), located at MSL (anemometer height is 30 m MSL). C-MAN stations provide 2-min averages of wind speed and direction, peak 5-s wind gusts, air temperature, and barometric pressure on an hourly basis. In addition to these measurements, some C-MAN stations, such as SAUF1 and SVLS1, measure 10-min mean wind speed and direction six times each hour, as well as the peak 5-s gust during the hour.

Tide gauge measurements were obtained by the NOAA National Ocean Service at St. Augustine, Florida. Wave measurements from NDBC moored buoys 41006, 41009, and 41010 were examined. These buoys (two of which are located in Fig. 1; the other is east of the domain) showed no sign of large wave motion during the period of interest.

**d. Other sources of data**

Additional data were examined with respect to this case study, not all of which are presented in section 3. Surface and 500-mb National Meteorological Center (NMC) analyses for time periods preceding the event were examined, along with upper-air soundings at Waycross, Georgia, and Charleston, South Carolina; wind records from the Savannah and Jacksonville International Airports; and pressure traces from Savannah, Sapelo Island, and Brunswick, Georgia, and Jacksonville, Florida (see Fig. 1 for locations).

3. Atmospheric and oceanographic analysis

**a. Observations**

Figure 2 shows the surface and 500-mb NMC analyses for 0000 UTC 4 July 1992. A surface pressure ridge extended from the Bahama Islands westward into the Gulf of Mexico. Satellite images indicated clear skies occurred during the entire day over the Florida peninsula, which is an unusual event during the summer (Burpee 1989). A surface trough extended over the Appalachian Mountains from Virginia through Alabama. A weak cold front was located about 200 km east of Daytona Beach and extended eastward to an extratropical low over the Atlantic Ocean. The 500-mb chart (Fig. 2b) indicates the east-west ridge above the surface ridge over Florida. A region of depressed heights and relatively low tem-
FIG. 3. GOES infrared satellite imagery for (a) 2200 UTC 3 July 1992 and (b) 0100 UTC 4 July 1992. SAV is Savannah, Georgia; JAX is Jacksonville, Florida; DAB is Daytona Beach, Florida.

Temperatures at Waycross, Georgia, was likely associated with a short wave.

The satellite image for 2200 UTC 3 July (Fig. 3a) shows extensive cloudiness over Georgia and South Carolina, which was associated with a mesoscale convective system. Figure 3b shows the 0100 UTC 4 July image, which indicates the southward reposition of the cloud mass over Georgia. The cloudiness east of Florida appeared to be dissipating.

A series of radar analyses, at hourly intervals, from 2000 UTC 3 July through 0325 UTC 4 July 1992 was examined. Between 2000 and 2130 UTC, a squall line was present over central Georgia, oriented east to west, near 33°N, and was moving southward (not shown). Satellite imagery (not shown) indicated that the line extended from the coast of South Carolina west through Georgia and into Alabama. By 2228 UTC (Fig. 4a), the line of convection had moved southeastward across Georgia. Satellite imagery (not shown) indicated that the line extended from the coast of South Carolina west through Georgia and into Alabama. By 2228 UTC (Fig. 4a), the line of convection had moved southeastward across Georgia. Satellite imagery indicated that a well-defined line of convection was present, with a trailing stratiform region typical of squall line structure (Houze 1993). A time series of wind speed from Savannah International Airport (Fig. 5) indicated maximum wind gusts of 18 m s⁻¹ at 2125 UTC, with a 15-min average wind speed of ~10 m s⁻¹. This corresponded closely with the passage of the squall line according to the radar reflectivities (Fig. 4a).

The squall line extended over the South Atlantic Bight (SAB), reaching SLVS1 by 2228 UTC (Fig. 4b). SLVS1 showed a rapid increase in the 10-min mean wind speed, from less than 4 m s⁻¹ during 2000 to 2210 UTC to greater than 14 m s⁻¹ at 2225 UTC. During this period, the wind direction shifted from 200° to 290°, as the squall passed (Fig. 6). Pressure rises of ~1.5 mb were associated with the squall passage at Sapelo Island, Georgia, and Brunswick, Georgia (not shown).

By 2330 UTC, radar echoes with VIP levels 4 and 5 were present over the coastal waters of northern Florida (Fig. 4b). The leading edge of the convection appeared to be moving southward over the water, while the leading edge over land was moving southward more slowly. Satellite imagery 30 min later indicated that the convection near 31°N, 81°W was the southernmost point of an arc-shaped line of convection pointing southward.

A wind trace from Jacksonville (Fig. 7) indicated the passage of the squall, beginning at 2325 UTC, accompanied by gusts to 15 m s⁻¹. Sustained winds of ~10 m s⁻¹ lasted for ~15 min, followed by ~30 min of steadily decreasing winds. The surface pressure rose 1.0 mb at Jacksonville (Fig. 8) over a period of 20 min, beginning at 2330 UTC as the squall passed. The higher pressure lasted ~1 h.

The squall had advanced farther southward, maintaining the bow-shaped echo at 0025 UTC 4 July (Fig. 4c). The leading edge of the line was now in the region of dual-radar coverage and showed level 5 radar echoes. The bow shape of the convective line was apparent in the satellite imagery (Fig. 3b), and clouds extended east to 78°W, well beyond the 278-km range of the radar analysis.
The leading edge of the squall line had moved southward, to the vicinity of SAUF1 at 0130 UTC (Fig. 4d). Between 0125 and 0145 UTC (Fig. 9), the 10-min mean wind gust during this period was 18.7 m s$^{-1}$. Figure 4e shows that the squall line continued to advance farther south to halfway between St. Augustine and Daytona Beach. Finally, an hour later (Fig. 4f), the convection was dissipating and had advanced no farther southward.

Dennis Decker, the warning coordination meteorologist on duty at the NWS office in Melbourne during this period, observed the southward-moving squall...
He issued standard advisories to marine interests since this appeared to be a squall line of typical intensity (D. Decker 1993, personal communication).

Figure 10 shows isochrones of the leading edge of the convective line over the region as indicated by VIP level 2 in the radar reflectivity composites (Fig. 4). The southward advancement of the squall line is clearly evident during the period 2000 UTC 3 through 0230 UTC 4 July. Between 0025 and 0130 UTC, as denoted by the isochrones ~30 km offshore of Florida, the squall line moved ~50 km southward, yielding a translation speed of ~13 m s⁻¹ parallel to the coastline. During the next hour, the squall line continued to advance southward at 13 m s⁻¹. This speed of translation matched the peak 10-min average wind at St. Augustine (Fig. 9), indicating that the wind surge was collocated with the leading edge of the squall line. Such wind surges, or gust fronts, frequently accompany squall lines. Since the squall line propagated at this speed for at least 3 h, it is likely that the gust front associated with the squall line lasted at least 3 h. The tide gauge at St. Augustine (Fig. 9) measured a peak surge (observed tide minus predicted tide) of ~0.3 m relative to National Geodetic Vertical Datum (NGVD) of 1929, beginning at 0120 UTC, coinciding (plus or minus ~5 min) with the increase in wind speed.

Isochrones of the leading edge of cloudiness, as inferred from the visible and infrared satellite imagery, were examined on a hourly basis (not shown). These isochrones indicated that the leading line of convection extended eastward ~300 km over the ocean (and off the edge of the map domain). Though portions of the squall line were located east of the radar coverage, the translation toward the south was generally much slower than near land (consistent with the bow echo in the radar pattern) and, in any case, was located over deeper water, where the shallow-water wave phase speed was much higher than the propagation speed of the squall line. Hence, it is unlikely that the portion of the squall line east of 80°W could have generated the wave that hit Daytona Beach. The satellite imagery confirmed that the squall never actually reached Daytona Beach, and that there was no other deep convection in the region. Hence, the wave run-up was not associated with atmospheric convection in the immediate vicinity of Daytona Beach.

A small pressure rise of ~0.3 mb occurred at Daytona Beach between 0200 and 0300 UTC. It is not apparent whether this pressure rise was related to the squall line.

b. Time of arrival analysis
The values of shallow-water phase speed over the SAB, beneath the convection, ranged from 5 m s⁻¹ just off the coast to 20 m s⁻¹ at the radar range rings. The bottom topography at St. Augustine results in shallow-water phase speeds that range from 10 m s⁻¹ at 20 km offshore to 15 m s⁻¹ at ~50 km offshore. The location of the 13 m s⁻¹ phase speed contour thus lies ~80 km due north of Daytona Beach. If the 13 m s⁻¹
winds excited a shallow-water wave with a phase speed of 13 m s\(^{-1}\), the travel time to Daytona Beach would be 1.7 h. With an initial time of 0130 UTC (based on the radar analysis in Fig. 10), the estimated time of arrival at Daytona Beach would be ~0312 UTC.

Thieke et al. (1993) reported that the wave reached the northern end of the Daytona Beach area at 0256 UTC and proceeded southward 24 km during the following 30 min. The estimated time of arrival of the wave based on analysis of the radar data and the bathymetry lies in the middle of the time interval during which the wave was observed to arrive. These reports are thus consistent with the hypothesis that the wind surge associated with squall line excited the rogue wave.

4. Scaling the atmospheric forcing

Section 3 showed that a squall line propagated along the coast at approximately the speed needed to be in resonance with a shallow-water gravity wave in the ocean. Our hypothesis that this wave was generated by the squall line is supported further here by estimating the magnitude of the atmospheric forcing. An idealized momentum equation for a slab of water and reasonable if not well-observed values for the forcing terms in this equation are used. Since the available data are sketchy, this case does not warrant a rigorous and detailed treatment of the atmospheric forcing. The expression that is evaluated is quite simple, and it would be straightforward to apply to other events of a similar nature in either a forecast or hindcast mode.

In a derivation outlined in the appendix, it is shown that the time rate of change of the height of the wave, in a frame of reference moving with the wave, can be estimated as

\[
\frac{dh}{dt} = \frac{1}{2c}(HF + R),
\]

where \(h\) is the wave height, \(c\) is the shallow-water gravity wave speed, \(H\) is the mean depth, and \(F\) and \(R\) are terms representing the horizontal atmospheric pressure gradient and surface stress, respectively. This indicates that the wave height responds to the sum of the forcing from atmospheric pressure gradient and wind stress. The forcing terms in (1) can be roughly estimated from the available data, with sufficient accuracy for this analysis. The mean depth \(H\) is 17 m, consistent with a wave speed of 13 m s\(^{-1}\). The atmospheric pressure difference across the crest of the wave is 2 mb over a spatial scale of 10 km. The former value is based on the pressure observed at Brunswick Naval Air Station, Georgia, as the edge of the squall line passed; the spatial scale is based on previous observation of squall lines (e.g., Charba...
1974) and is equivalent to a temporal scale of 10 min at a fixed point.

With a water density of $10^3$ kg m$^{-3}$, the pressure gradient force term [the first term on the rhs of (1)] is $= 1.3 \times 10^{-5}$ m s$^{-1}$. The stress term is estimated using the bulk aerodynamic method with a wind speed of 15 m s$^{-1}$, a drag coefficient of $1.5 \times 10^{-3}$, and an air density of $1.2$ kg m$^{-3}$, which yields a magnitude of $= 1.6 \times 10^{-5}$ m s$^{-1}$. Assuming the pressure gradient and surface stress terms are additive and constant with time, the height (m) of the crest of the wave at time $t$ (s) is

$$h(t) = 3 \times 10^{-5}t.$$  \hspace{1cm} (2)

The atmospheric forcing is assumed to have been exerted for $3 \times 10^4$ s, which implies a wave-scale height over the open water of $= 0.3$ m. This estimate (0.3 m) for the height of the wave over the open ocean is considerably lower than the wave heights and run-up elevations observed at Daytona Beach.

The difference between the observed wave and the scale analysis can at least partly be accounted for following the analysis by Thieke et al. They suggest that a squall line—generated edge wave (e.g., Munk et al. 1956; Greenspan 1956) along the Florida coast was responsible for the hour-long sea level rise observed at St. Augustine (Fig. 9b). This edge wave would have served to locally raise the sea level for the primary wave at Daytona Beach. More importantly, Thieke et al. also discussed the relationship between wave run-up elevation on the beach and the wave height offshore. Using the empirical relationship of Synolakis (1987), they estimated for Daytona Beach that the observed maximum run-up of 1.8 m implies an offshore wave height of 0.7 m. The treatment by Synolakis (1987) is two-dimensional and is therefore appropriate for a beach with a constant slope—that is, without alongshore variations in bathymetry. For the present case, a ridge of relatively shallow depth extends northward offshore of Daytona Beach. An ocean wave propagating from the north would likely refract as it encountered this feature, and the wave energy would be focused at Daytona Beach. As indicated earlier, the extreme wave heights were observed at the beach over a region only 8 km wide. Our estimate of an open-ocean wave height of 0.3 m is consistent with the observed run-up elevations of 0.7 m outside this narrow region, given a typical ratio between the run-up and the wave height offshore.

In summary, our scaling yields an open-ocean wave amplitude that is broadly consistent with the observed wave run-up along the Florida coast. There is a factor of approximately 2 between the estimated open-ocean height of 0.3 m and the height of 0.7 m needed to account for the maximum run-up observed at Daytona Beach. This factor is hypothesized to be due to the combination of three effects: errors associated with the assumptions used in deriving (1), uncertainties in the magnitudes of the forcing terms in (1) perhaps due to alongline variations in squall intensity, and bathymetric focusing. Additional cases are needed to determine if (1) can be reliably used to estimate open-ocean (or lake) wave amplitudes.

5. Conclusions

Analysis of weather surveillance radar, coastal surface observations, and geostationary satellite imagery indicated that a squall line propagated southward over the SAB during the hours preceding the arrival of at least one large wave at Daytona Beach, Florida. Analysis of tide gauge and atmospheric data
The growth of the wave is assessed in a frame of reference moving at the shallow-water gravity wave speed. This method separates the effects of the rapidly varying component associated with the free wave from the slowly varying component associated with the growth of the wave.

Platzman (1958) described the equation of motion associated with a water wave as

\[ \frac{\partial \mathbf{V}}{\partial t} = -g \nabla h - \rho_w \nabla p + \rho_w \frac{\partial \tau}{\partial z}, \]  
(A1)

where \( \mathbf{V} \) is the water velocity, \( h \) is the elevation of the free surface above mean level, \( p \) is the atmospheric pressure, \( \tau \) is the horizontal frictional stress, \( g \) is gravity, and \( \rho_w \) is the density of water. He neglected nonlinear advective terms, nonhydrostatic forces, density variations in the ocean, the Coriolis force, and stress at bottom.

We extend Platzman's analysis by integrating (A1) over the depth of the water, to obtain

\[ \frac{\partial \mathbf{M}}{\partial t} = H(F - g \nabla h) + \mathbf{R}, \]  
(A2)

where

\[ \mathbf{M} = \int_{-H}^{h} \mathbf{V} \, dz \]  
(A3)

is the depth-integrated transport,

\[ F = -\rho_w \nabla p \]  
(A4)

is the pressure gradient term, and

\[ \mathbf{R} = \rho_w \nabla \tau \]  
(A5)

is the surface stress term. In these equations, \( H \) is the mean water depth (assumed constant). It is assumed that the wave moves along an isobath at the speed of a shallow-water gravity wave, \( c = (gH)^{1/2} \), and that variations in the \( y \) direction are negligible. Figure A1 illustrates the hypothesized relationship between the squall line and the water wave.

Next, we express the time derivative in a frame of reference moving with the wave:

\[ \frac{d}{dt} = \frac{\partial}{\partial t} + c \frac{\partial}{\partial x}. \]  
(A6)
Using (A6) in (A2), the transport equation can be recast as

\[ \frac{dM}{dt} - c \frac{\partial M}{\partial x} = H \left[ F - \frac{g}{c} \left( \frac{dh}{dt} - \frac{\partial h}{\partial t} \right) \right] + R, \quad (A7) \]

assuming that the atmospheric forcing terms are constant in time and space moving with the wave. Collecting terms, (A7) can be rewritten as

\[ \frac{dM}{dt} + c \frac{dh}{dt} = (HF + R) + c \left( \frac{\partial M}{\partial x} + \frac{\partial h}{\partial t} \right), \quad (A8) \]

using \( c = (gH)^{1/2} \). The second of the bracketed terms on the rhs of (A8) is zero for an ideal shallow-water gravity wave. Individually, each of these two terms is large, but they cancel each other by way of the continuity equation (e.g., Holton 1979, p. 158). Therefore, the change in transport and height following the wave can be estimated considering just the forcing following the wave.

The mass transport can be related to the height by assuming an equipartition of energy between kinetic and geopotential, with the crest of the wave coincident with the maximum in transport, as for an ideal shallow-water wave. The kinetic energy of the wave is thus related to the geopotential energy by

\[ \frac{1}{2} \int_{-H}^{h} v^2 dz = \frac{gh^2}{2}. \quad (A9) \]

Two additional approximations are made:

\[ \frac{1}{2} \int_{-H}^{h} v^2 dz = \frac{H}{2} \bar{v}^2, \quad (A10) \]

where \( \bar{v} \) is the mean velocity of the water over the layer, and

\[ \bar{v} = \frac{M}{H}. \quad (A11) \]

This implies that

\[ M = ch. \quad (A13) \]

Substitution in (A8) gives

\[ \frac{dh}{dt} = \frac{1}{2c} (HF + R). \quad (A14) \]

This indicates that the wave height responds to the sum of the forcing from atmospheric pressure gradient and wind stress.

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References

With the development of meteorological science and the continual refinement of the technologies used in its practical application, the need to produce a new edition of the International Meteorological Vocabulary (IMV) became evident (the original edition was published in 1966). This volume is made up of a multilingual list of over 3500 terms arranged in English alphabetical order, accompanied by definitions in each of the languages (English, French, Russian, and Spanish) and an index for each language. This new edition has been augmented with numerous concepts relating to new meteorological knowledge, techniques, and concerns. It should help to standardize the terminology used in this field, facilitate communication between specialists speaking different languages, and aid translators in their work.