# An Example of North Atlantic Deep-Ocean Swell Impacting Ascension and St. Helena Islands in the Central South Atlantic

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#### ABSTRACT

During 1999, the dataloggers of the pressure transducer–based tide gauges at Ascension and St. Helena Islands were upgraded in order to enable the monitoring of wave conditions in addition to the measurement of still water levels. Within a few months, the gauges had recorded an example of unusually large deep-ocean swell, which, from the inspection of numerical wave model output, appears to have been generated by the remains of Hurricane Irene in the North Atlantic almost 1 week earlier. This fortuitous event serves to remind us of the potential importance of swells to communities on distant, low-lying coasts, particularly if the climatology of swells is modified under future climate change, and of the importance of in situ wave recording to wave model development. It is suggested that global ocean monitoring programs should place greater emphasis than hitherto on swell monitoring and prediction, with one component of the monitoring being provided by island tide gauges.

#### 1. Introduction

A deep-ocean swell can travel many thousands of kilometers from the storm that produces it and can cause considerable damage on distant coasts. A major "swell event" was observed around 26 October 1999 at Ascension and St. Helena Islands in the central South Atlantic (Fig. 1). The swell had propagated from the remains of Hurricane Irene, which occurred about 1 week earlier in the central North Atlantic, holding approximately the same position for 2 days, thereby providing the sustained, intense directional wind forcing required for the generation of a swell (Tricker 1964; Tucker and Pitt 2001). The increasing wave heights and lengthened wave periods characteristic of swells were observed in the first set of 1-Hz-sampled subsurface pressure (SSP) data obtained from the coastal tide gauges at English Bay (Ascension) and Jamestown (St. Helena). Both gauge sites are on the northwestern side of the islands in the lee of the southeast trade winds, and are, consequently, exposed to swells originating from the north-west.

Cartwright et al. (1977) gave a historical background to the study of "rollers" at St. Helena, and explained the various issues to be considered when undertaking measurements of swells in shallow water. The paper also described a first set of measurements at the island using a wave recorder (pressure sensor) deployed in 12 m of water in James Bay, approximately 180-m offshore of Jamestown, for 2 months in 1970–71. However, its conclusions were limited by the short duration of the measurements, by the fact that there were no substantial swell events in that period, and by the difficulty of assigning individual events to particular distant storms. Deep-ocean wave modeling at that time was in its infancy.

The present paper can be considered as continuing the work of Cartwright (1971) and Cartwright et al. (1977), by making use of modern technical developments to make measurements right at the coast, and by employing a wave model to identify the origin of swells. It describes some of the recent data obtained, which includes evidence for swells significantly larger than that observed by Cartwright et al. (1977), and explains why such a technical capability is important to global ocean observing.

#### 2. The tide gauges

The Ascension and St. Helena tide gauges form part of a United Kingdom contribution to the Global Sea

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FIG. 1. Locations of Ascension and St.Helena Islands in the central South Atlantic.

Level Observing System (GLOSS; IOC 1998), and have been operated in their present form since 1993 (Spencer et al. 1993; Axe 2000). They are based on measurements of SSP by Paroscientific Digiquartz transducers located below low tide, together with an efficient method for data control that involves two additional transducersone in the open air (a barometer) and one located at approximately mean sea level (Woodworth et al. 1996). The three transducers are denoted in technical documents as "C" (the SSP-measuring transducer), "B" (at approximate mean sea level), and "A" (the barometer) (e.g., IOC 2002). All three are enclosed within a vertical container attached to the coastal rock face at Ascension, and to the harbor wall at St. Helena. Only the C sensor is relevant to the present study. This is located at the bottom of the vertical container above the seabed. The gauge dataloggers provide measurements of SSP (and of the A and B pressures used for data control, and of derived sea level) averaged over 15-min intervals. Such a sampling of "still water levels" is normally adequate for the study of all oceanographic processes from storm surges and tides through to long-term sea level change, and provides a relatively small dataset that can be transmitted to data centers in near-real time.

Following successful short tests in 1997, the loggers at Ascension and St. Helena were adapted in June and March 1999, respectively, in order to provide a second dataset with 1-Hz sampling suitable for wave studies. This was an interesting technical development because it is unusual, in our experience, for tide gauges to have a wave measurement capability, although the use of pressure sensors to record waves in shallow water is itself by no means new (e.g., Munk and Snodgrass 1957; Seymour et al. 1994; Tucker and Pitt 2001). Continuous recording of wave data was maintained until November and December 1999, respectively, and provided the datasets used in this paper. In view of the considerable amount of data needing to be stored in the loggers, which is as yet not routinely transmitted, a new mode of operation has since been installed that is based on a local, continuously overwritten buffer that stores the most recent year of wave data. The buffer can be downloaded to a portable computer whenever possible.

SSP is measured in millibars (mb) of pressure, which

is almost equivalent (to within approximately half a percent) to centimeters of water; we have used centimeter units below. In our analysis, the 1-Hz values of SSP are first studied within moving 25-h windows, with the tidal contribution to the SSP time series computed and removed, leaving a residual SSP series. The tide is predominantly semidiurnal at the islands, with an amplitude of M2 (the main semidiurnal tide due to the moon) of approximately 33 cm at both places. Wave properties are then estimated from the residual series using shorter windows of 1 h, with the root-mean-square (rms) of the 1-Hz values within the window used to define significant wave height  $(H_s)$  via the relation  $H_s = 4 * \text{rms}$ , with  $H_s$  corresponding to the average height of the highest one-third of the waves (Cartwright and Longuet-Higgins 1956). Wave period is represented by a simple "mean zero-up-crossing period  $(T_z)$ ," which is the average difference between times in the 1-Hz series at which residual SSP increases from less than to more than its mean value in the window.

Cartwright et al. (1977, their section 3) explained the necessary corrections to convert measurements by shallow-water pressure sensors into inferred values of  $H_s$  in the neighboring deep ocean. These corrections include terms for the attenuation of the wave signal with depth, and for the amplification of deep-ocean waves on entering a small island shelf, such as that at St. Helena. They explained that the bottom descends abruptly from the shore (their Fig. 1) and, consequently, dissipation of wave energy will be small. They concluded that, with their sensors deployed at a 12-m depth, such corrections for  $H_s$  are at the 5%–10% level for swells in the  $T_z$  range of 10–20 s (their Fig. 2). Similar corrections would also apply for measurements at the same depth at Ascension, where the bottom descends even more abruptly.

In the present work, we have employed even shallower sensors located approximately 1 m below low water, which essentially removes any signal attenuation. The most important remaining issue is, then, consideration of the amplification of the deep-ocean waves in the shallow waters adjacent to the gauge. From inspection of hydrographic charts, we estimate the water depth to be about 6 m within a few 10s of meters of both gauges (within approximately 50 m at St. Helena and 20 m at Ascension), which is a depth for which the corrections would be approximately 15% for swells in the  $T_z$  range of 10–20 s. This is sufficiently small that we have not applied them in the analysis below. Similarly, we have not adjusted the observed  $T_{z}$  for a small overestimation due to the relatively greater attenuation of signals from shorter waves at depth, and possible modifications of wave spectra due to interactions with wind or wave-induced currents have also not been taken into account. We realize that this simplified analysis could be potentially improved upon by a numerical modeling exercise that employs a complete knowledge of shallow water bathymetry, and includes a consider-



ation of wave breaking in the few meters before the gauge.

#### 3. The 26 October 1999 swell event

Figure 2 shows North Atlantic meteorological surface analysis charts at 1200 UTC for the week of 15-21 October 1999. The depression in the Labrador Sea on 15-16 October originated from several smaller lows in the area of the Great Lakes and the St. Lawrence Valley, which coalesced around 13 October and moved northeast. The Mariners Weather Log (NOAA 2000) lists ship reports of local wave heights of 9 m on 15 October, and describes the depression as a "48 hour storm ending at around 0000 UTC 16 October, which then stalled and drifted east and weakened." Figure 2 shows the depression to have been considerably weaker on the 17th. By 19 October, the central part of the North Atlantic was occupied by the remains of Hurricane Irene, which had taken a northerly route parallel to the North American coast, reintensifying after passing Cape Hatteras. Ship reports from different points around Irene refer to wave heights ranging from 6 to (an unconfirmed) 23 m on 19 October (NOAA 2000). Irene deepened to 949 mb at 0000 UTC 20 October before continuing east and weakening.

The Met Office routinely runs global and regional spectral wave forecast models to predict the surface wave energy spectrum. The global model is a secondgeneration model based on that described by Golding (1983), with a continuous program of development to the present (Holt 1998). Figure 3 shows model analyses for  $H_s$  in the wave frequency bin centerd on an 18-s period, which will be seen as the most appropriate period for discussion here. The total wave energy at this wave period is included in  $H_s$ , which is for the most part remotely generated swell, although in some locations and at certain times the values will contain significant wave energy generated directly under the strong winds of the storm. The model is run at the same resolution (5/9 by 5/6 deg) as the numerical weather prediction model providing the forcing surface winds, and gives an estimate of the wave energy spectrum at each model grid point. The wave energy spectrum is discretized into 16 equally spaced directions and 13 frequency components, spaced logarithmically between 0.04 and 0.324 Hz.

Figure 3 (on 17 October) shows the origin of the "first pulse" of the swell in the North Atlantic in this week arising from the Labrador Sea storm. That pulse can be tracked southward by the model, and its arrival was detected by the gauge at Ascension around 23 October (yearday 296). It was less evident at St. Helena. It was followed 2–3 days later by a much larger "second pulse" originating from Irene (Fig. 3; on 21 and 23 October), which arrived at Ascension around the 26th. According to the model,  $H_s$  at the islands will have been on the order of 250 cm or more during the second pulse,

but most of the energy in both pulses passed to the west of the islands.

From Fig. 3, one can also note that there should have been significant energy at Ascension around 21 October (yearday 294) stemming from a northward-moving major swell system from the southwest Atlantic arriving some 2 days ahead of the North Atlantic first pulse, with most energy once again passing to the west of the islands. It will be seen that no large  $H_s$  was observed by either gauge at this time, indicating that their locations are probably unsuitable for monitoring swells from the south.

Parts of the time series of  $H_s$  recorded at the gauges are shown in Fig. 4. From June to late October, the "ambient"  $H_s$  at both Ascension and St. Helena had been approximately 50 cm. Only rarely did  $H_s$  exceed 80 cm, and was larger than 100 cm on only three or four occasions, comparable to the magnitude of swells experienced in the Cartwright et al. (1977) data. The longest sustained period over 100 cm was for several days in late September at Ascension, which from the Met Office model we learn may have contained a small contribution from the early stages of Hurricane Gert. Gert crossed the Grand Banks a few days later, contributing to significant higher-latitude North Atlantic swells (Met Office 2002).

The main features of the  $H_s$  time series are the very large values in late October at both islands, especially on yeardays 298–300 (25–27 October), but extending for several days further at a reduced level (Fig. 4). This event was significantly larger than any other in either our or the Cartwright et al. (1977) data, and took place at a time of spring tides. Respectively,  $H_s$  peaked at approximately 280 and 160 cm at the Ascension and St. Helena gauges. These are comparable to the wave model values of Fig. 3 (although values are not directly comparable, owing to the gauge values representing wave heights across the spectrum, neither gauge nor model values being corrected for depth effects).

The large (by Ascension and St. Helena standards) tides at this time have some instrumental and, possibly, oceanographic consequences within a discussion of the observed swell. Inspection of the original 1-Hz SSP time series for Ascension shows evidence for the swell at low tide on 26 October to have been so large that the wave troughs must have fallen occasionally below the pressure transducer. This can be seen by the recorded SSP having at times its minimum possible value (i.e., atmospheric pressure), the transducer being for a short time in the open air. (The transducer does not lose calibration when this occurs.) There is little that any gauge operator can do to prevent this from happening if the tides and swell combined are large enough to produce deep troughs. Although this is not a large effect within the overall dataset, it means that  $H_s$  values computed at low tide for these few days will have been slightly underestimated.

Figure 4 also suggests a reduction of  $H_s$  during low



FIG. 3. Significant wave height  $(H_s)$  for waves of an 18-s period from the Met Office global wave model, units are meters with contours every 0.25 m: (a) 1200 UTC 17 Oct, (b) 0000 UTC 21 Oct, (c) 1200 UTC 23 Oct, (d) 1200 UTC 26 Oct. The dots indicate the two islands.



FIG. 4. Time series of  $H_s$  computed using a 1-h window for (a), (b) Ascension and (c), (d) St. Helena. The short straight line in (c) indicates missing data. (Yearday 299 = 26 Oct.)

spring tides at St. Helena on day 299 when  $H_s$  was generally at its largest. However, there is no evidence in the St. Helena data that the wave troughs ever fell below the sensor as at Ascension. Consequently, we have concluded that this could be a real oceanographic, and not instrumental effect, possibly owing to the longer swell waves breaking farther from shore at low spring tides (cf. Thornton and Guza 1983). It can be seen that during the following days, as the tides moved away from springs, changes in  $H_s$  seem less correlated with the tide.

The corresponding time series for  $T_z$  are shown in Fig. 5. The ambient  $T_z$  is approximately 10 and 12 s at Ascension and St. Helena, respectively, until the swell event. Then  $T_z$  increases to approximately 18 s at both islands, at about the same times as maximum  $H_s$ , indicative of a swell rather than wind waves. (Hence, our use of the 18-s model output in Fig. 3.) Figure 5 shows that for several intervals at Ascension, there is a clear demonstration of the well-known process of longer-period swells arriving at a coast before the shorter periods.

This can also be seen for the event of yeardays 299–300 in the St. Helena data.

The large-swell event at Ascension shown in Fig. 4 can be used in a variant of the "ridge line" method first developed by Barber and Ursell (1948), later improved upon by Munk et al. (1963), and described more recently by Cartwright et al. (1977) and Tucker and Pitt (2001). It employs the equation

$$t = t_0 + 4\pi f \Delta/g_s$$

where *t* is the arrival time of swell of frequency *f*, which was produced at time  $t_0$  by a storm distance  $\Delta$  away, and *g* is the acceleration due to gravity. For example, the distance between Irene at approximately 50°N, 40°W and Ascension is 6900 km, which implies that the longest recorded swell (18 s) observed at around yearday 298.5 (Fig. 5c) must have originated approximately 5.7 days previously (i.e., on late 19 October or early 20 October), consistent with the meteorological observations. Alternatively, if one assigns days 298.5 and 302.0



FIG. 5. As in Fig. 4, but for  $T_z$ .

to  $T_z$  values of 18 and 12 s, respectively (Fig. 5c), one can eliminate  $t_0$  from the equation and compute a storm distance  $\Delta$  of 8500 km. Such  $T_z/t$  assignments from the time series are inherently less precise than the computation of  $t - t_0$  from  $\Delta$ . A similar exercise can be made from studies of the time difference of the occurrence of 18-s swells at the two islands from detailed inspection of the residual SSP power spectra. One estimates a 19-h time difference from the spectra, consistent with the above equation, with a distance between the islands of 1300 km. (One estimates a time difference of approximately half a day from the peaks of  $H_s$  and  $T_z$  in the individual time series of Figs. 4 and 5, but this procedure is less precise.)

Ascension - Wave Period T\_(1 hr)

The impact of the swell event on the normal 15-min SSP data is small but distinctive. At Ascension, the 15min "tidal residuals" (measured SSP minus SSP predicted from previous knowledge of the ocean tide) for the period show small (several centimeters) positive fluctuations at low tides, which are consistent with the wave troughs falling below the transducer as mentioned above. At St. Helena, where there is no evidence for a similar wave–trough effect, the 15-min residuals for 26–30 October present a seichelike record, with a typical amplitude of 3 cm and periods between 0.5 and several hours. We have concluded that this was caused by fluctuations in real physical processes during the swell event, either in wave setup in the corner of James Bay in which the gauge is located, or in seichelike activity caused by the swell.

Ascension - Wave Period T\_(1 hr)

We have searched for other examples of these distinctive signals in the 15-min residuals because the present gauges were installed a decade ago, without discovering any, although a small number of examples exist in both records of high-frequency seichelike variability similar to that at St. Helena discussed above, but with a small amplitude. Consequently, although we know from the Met Office wave model and from a variety of observations that the islands are subject to a number of swell events each year, it does suggest that the event of 26 October was unusual. 1102

# 4. Importance of wave observations to model development

Techniques for modeling deep-ocean waves have progressed greatly in recent years (Komen et al. 1994). An overview of operational numerical wave modeling has been given by Bidlot and Holt (1999). Observations for validation of wave models have also improved, in particular through routine measurements of wave heights by satellite remote sensing, with other wave properties being theoretically observable but less accessible (e.g., Davies et al. 1997). However, developments in both model physics, including that related to swell, and in observations for validation are still required.

Model validation benefits from in situ recording because time series measurements from ocean buoys, light vessels, weather ships, and tide gauges at particular locations are clearly complementary to the data acquired from the moving ground track of a satellite, although the in situ data are as yet largely restricted to Northern Hemisphere coastal waters (e.g., Bacon and Carter 1991; Mettlach et al. 1994). Ocean buoy data have been used intensively in wave model validation (Bidlot et al. 2002) and in the intercalibration of wave height measurements from space (Cotton and Carter 1994). Offshore wave recorder data have also been used to study deep-ocean swell propagation across a wide shelf to the coast (e.g., Draper and Bownass 1983). However, there are relatively few detailed studies of swell events using information right at the coast where the swell impacts, largely because tide gauges are not usually designed to record such high-frequency sea level changes.

There are good reasons to make wave measurements at gauge sites for model validation, as long as the gauges are located such that useful wave information will be acquired. Therefore, it is reasonable to suggest that at least some of the gauges in the global network be equipped to record wave conditions using pressure transducers capable of 1 Hz or similar sampling. This would apply particularly to locations where swell events have sometimes been misreported as storm surges, or even mean sea level changes (Harangozo 1992). Such higher-frequency information may also be of use to studies of seismic events and tsunamis.

### 5. Some conclusions

It is to be hoped that technical developments, such as those described here, will lead to the compilation of enhanced databases of wind waves and swells for climate studies and local coastal engineering applications, as well as in aiding wave model development and, thereby, the capability for more accurate swell forecasting.

In performing this study, two important recommendations occurred to us. The first recommendation can be made to global ocean monitoring programs to consider investment in the instrumentation of tide gauges so that wave recording is also possible (tsunami recording would be automatically accommodated as a byproduct). This need not be a requirement at all sites, but a subset of the GLOSS network might be considered, including sites in the "swell pools" on the eastern sides of the tropical oceans (Chen et al. 2002). The present study has demonstrated that the existing Ascension and St. Helena gauges are ideally located to study swells propagating on great circle paths from the North Atlantic. On the other hand, they appear to be insensitive to swells from the south. With a second set of gauges installed on the southeast coasts of the islands, swells originating from most directions should be monitored efficiently. It would be interesting to use global and local wave models to determine which gauge sites in other parts of the ocean would be optimal for investment.

The second recommendation can be made to scientists engaged in climate change impacts studies. Animations made from Met Office model output and studies of previous swell events elsewhere (e.g., at the Maldive Islands; Harangozo 1992) demonstrate strongly just how far swells can travel and still have an impact on a distant, low-lying coast. With an increasing number of people living close to the coast (Nicholls and Small 2002), deep-ocean swell generation, and its potential modification as a consequence of climate change, is clearly an issue that needs attention, alongside the more intensively studied topics of changes in mean sea level and storm surges.

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