A Numerical Study of the Storm Surge Generated by Tropical Cyclone Jane

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ABSTRACT

In 1983 Tropical Cyclone Jane crossed the North West Coast of Australia generating a storm surge. Currents associated with this storm surge were recorded at two offshore moorings south of the cyclone track. The data from these moorings are suggestive of the propagation of a continental shelf wave between the two stations. This hypothesis is tested by carrying out a numerical simulation of this storm surge based on the depth-integrated shallow-water equations, with wind-wave-enhanced bottom friction. Analysis of the numerical results shows that the storm surge can be interpreted as due to continental shelf waves.

1. Introduction

The last few decades have seen much effort directed at the modeling of storm surges generated by intense localized atmospheric systems, such as tropical cyclones or midlatitude depressions (see, for instance, the reviews by Heaps 1983; or Das 1994). However, while much attention has been paid to the various aspects of the numerical formulation, there has been comparatively less attention paid to understanding the dynamical processes involved. Here we are particularly concerned with the nature of the ocean response in the region outside the area of intense wind stress forcing. For instance, such a focus is relevant in the relaxation phase of a storm surge after a tropical cyclone has made landfall, or for the case of a tropical cyclone traveling parallel to the coast in the regions upstream and downstream of the tropical cyclone center.

From theoretical considerations it is clear that the ocean response in the regions outside the forcing center should consist of longshore propagating coastally trapped waves. Further, Grimshaw (1988) used linearized long-wave theory to establish that most of the response should consist of low-mode continental shelf waves. This conjecture was subsequently supported in a numerical model study by Tang and Grimshaw (1995). Similar conclusions have been reached by Martinsen et al. (1979) in a numerical study of the storm surges on the western coast of Norway, by Beardsley and Haidvogel (1981) in a numerical study of the response of Middle Atlantic Bight to a translating storm, by Liu (1987) in a study of storm surges in the East China Sea, and by Fandry and Steedman (1994) in a numerical model study of storm surges on the North West Shelf of Australia. Further support can be found in the study by Fandry et al. (1984) of the sea-level data associated with storm surges generated by tropical cyclones on the West Australia Coast, although their theoretical analysis was for a flat-bottom ocean that allows only for Kelvin waves.

It is pertinent to recall here that the coastally trapped wave spectrum generally consists of superinertial edge waves, Kelvin waves, and subinertial shelf waves (see, for instance, Le Blond and Mysak 1978; or Allen 1980). However, the time and space scales of tropical cyclones are usually such that the continental shelf wave portion of the spectrum is preferentially excited (Tang 1994). Further, continental shelf waves are vorticity waves whose kinetic energy component is considerably larger than their potential energy component. Thus, they can be effectively generated by the tropical cyclone wind.
stress field, since the predominant response in the ocean due to strong wind stress forcing is a cyclonic eddy (see, for instance, Tang 1994, or Figs. 7b, 8b, and 9b). Although the generation of continental shelf waves should thus be taken as the norm, we note that there are circumstances where the other components of the spectrum could also be excited (Fandry and Steedman 1994; Tang 1994). For instance, Johns and Lighthill (1992) show that, over a uniform shelf slope with no deep-water rigid boundary, tropical cyclones traveling parallel to the coast in the opposite direction to shelf waves can generate edge waves.

Since continental shelf waves are vorticity waves with predominant kinetic energy compared to potential energy, they are more readily identified from current meter observations than sea-level observations (e.g., Allen 1980). But, in contrast, most observational data relating to storm surges are derived from coastal sea-level measurements. Clearly, this can cause some difficulties in correctly determining where in the coastally trapped spectrum an observed storm surge signal should lie. Hence, in seeking to establish the dynamical processes constituting storm surges it is desirable to obtain current meter data, preferably offshore. Thus, the current meter data obtained at the offshore moorings North Rankin and North Gorgon when Tropical Cyclone Jane approached the North West Coast of Australia (see Fig. 1 for location) are particularly useful in this context and form the basis for this study. Particularly noteworthy is the phase lag between the storm surge arrival at these two sites, which suggests the interpretation of the data in terms of continental shelf waves; indeed, this is the main purpose of this paper. This is achieved by carrying out a numerical simulation of the storm surge generated by Tropical Cyclone Jane and analyzing the numerical model results in terms of coastally trapped wave processes. In section 2 we describe the observations in more detail. Then in section 3 we describe our numerical results. We note here that our numerical model is based on the vertically integrated shallow-water equations since these are sufficient for the analysis of the barotropic coastally trapped wave field. Aspects of the vertical structure of the currents generated due to internal frictional processes requires a three-dimensional model, and the application of such a model to storm surges in the North West Shelf of Australia is described by Hearn and Holloway (1990). In section 4 we provide a concluding summary and discussion.
2. Observations

Tropical Cyclone Jane developed well off the coast northwest of Western Australia, and after meandering for several days moved toward the southeast and the coastline. Tropical Cyclone (TC) Jane reached peak intensity a short distance from the coast on 9 January 1983 with an estimated central pressure of 947 mb and a maximum wind speed of 47 m s$^{-1}$. The cyclone crossed the coast approximately 100 km northeast of Port Hedland, as shown in Fig. 1, and then weakened as it moved inland. A more detailed description is given by Bate (1983).

Meteorological observations were obtained from coastal stations at Port Hedland and King Bay. Time series of wind speed and direction and atmospheric pressure are plotted in Fig. 2. Values were recorded every 10 minutes at King Bay and every 3 hours at Port Hedland. Maximum winds of 21 m s$^{-1}$ and minimum atmospheric pressure of 980 mb were recorded at Port Hedland at around 1400 UTC 9 January coinciding with the time at which the cyclone crossed the coast. At King Bay, approximately 250 km southwest of the cyclone track, the winds were weaker with a maximum of 15 m s$^{-1}$ and atmospheric pressure of 992 mb. Additional, but limited, observations were made twice daily at Barrow Island (see Fig. 1 for location), and the maximum recorded wind was 12 m s$^{-1}$ and minimum atmospheric pressure was 991 mb. These records demonstrate the decay in the forcing of the wind away from the cyclone center.

Two moorings, at North Rankin and North Gorgon (see Fig. 1 for locations), each held strings of three current meters and seabed pressure gauges for measuring sea level, and recorded the effects of TC Jane. The current meters at North Rankin were Neil Brown Acoustic current meters and recorded 5-min vector averages of currents at depths 23, 83, and 120 m in 123-m water depth. The current meters were deployed on 7 January 1983 under threat of the cyclone and recovered on 17 May 1983. At North Gorgon, EG&G electromagnetic current meters recorded 15-min vector averages at depths 23, 89, and 205 m in 208-m water depth. The current meters were deployed on 12 August 1982 and recovered on 22 January 1983. The moorings were separated along the shelf by a distance of 153 km.

In order to resolve the ocean response to the cyclone it is necessary to remove the strong tidal motions found in this region from the recorded signals. This is achieved by using low-pass filters. With the short record of observations at North Rankin prior to the cyclone, only a weak filter could be used so that not too much data were lost. The filter removed most of the semidiurnal current but allowed some diurnal motion to pass. For consistency, the same filter has been used on the North Gorgon data. The resulting time series are plotted in Figs. 3 and 4 where currents have been resolved into cross-shelf and alongshelf components with positive alongshore 65° east of north at North Rankin and 30° east of north at North Gorgon, the different orientations reflecting the local bathymetry.

At both locations strong alongshelf currents to the southwest are observed, particularly at the upper two
current meters. Currents reach 0.6 m s$^{-1}$ at depths 23 and 83 m at North Rankin and 1.0 and 0.75 m s$^{-1}$ at depths 23 and 89 m at North Gorgon. Weaker responses are at the seabed. The peak flows seen in the upper water column are sustained for approximately 36 hours. The very strong currents at North Gorgon are particularly surprising as the location is some 450 km from the cyclone track and at Barrow Island, only 55 km from North Gorgon, winds reached only 12 m s$^{-1}$. The local forcing was clearly insufficient to generate such strong currents. It is also interesting that the response was stronger at this deeper site than the shallower North Rankin location that was 153 km closer to the cyclone path.

The low-pass filtered sea level responses from the two moorings are plotted in Fig. 5. In this case, a weak filter has again been used but designed to remove semi-diurnal and most of the diurnal motion (weak diurnal oscillations are evident in the filtered records). Significant positive surges are seen at both locations, the largest of 0.3 m at North Rankin and the smaller of 0.1 m at North Gorgon. At each location, the times of maximum surge appear to occur at the same time as the maximum alongshelf current. In addition, coastal sea levels were recorded at Broome, Dampier, and King Bay (locations are shown in Fig. 7). Time series of water levels with predicted tides removed (tidal residuals) are plotted in Fig. 6 and show surges in response to Jane at each location. The surge at Broome, to the northeast of Dampier, lags those at Dampier and King Bay by approximately 20 h.

There is a significant phase lag between the responses at North Rankin and North Gorgon with North Rankin leading, suggesting the response propagates between the locations. The time lags have been estimated from the filtered time series by simply overlaying the plots and obtaining the best agreement in the shapes of the time series. However, because the time series have quite broad peaks, it is difficult to estimate the time lags to within 1–2 h. Comparing the currents at 23 m from the two locations gives a time lag of about 10 h, while the currents at 83 and 89 m give a time lag of about 7 h and the sea levels give a time lag of about 8 h. Further, in Fig. 11b we show a time series of the magnitude of the depth-integrated currents at each site, from which we infer that there is a time lag of about 9 h. Using an averaged value of 8.5 h, and with a separation distance of 153 km, the speed of propagation is found to be 5 m s$^{-1}$ with an estimated error of ±1.0 m s$^{-1}$.

### 3. Numerical simulations

In this section we describe some numerical simulations of the storm surge generated by TC Jane in order to demonstrate that the observations can be consistently interpreted as associated with the generation of coastally trapped waves. We use the depth-integrated shallow-water equations that are, in standard notation,

\[
\begin{align*}
\frac{\partial}{\partial t} &+ u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} \left( \frac{\tau^u}{\rho H} - \frac{\tau^v}{\rho H} - \frac{P_w}{\rho} \right) = 0, \\
\frac{\partial}{\partial t} &+ u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} \left( \frac{\tau^u}{\rho H} - \frac{\tau^v}{\rho H} - \frac{P_w}{\rho} \right) = 0,
\end{align*}
\]

\[
\zeta + (Hu)_x + ( Hv)_y = 0,
\]

where $H = h + \zeta$ is the total water depth, $\zeta$ is the surface elevation relative to the undisturbed water depth $h(x, y)$, $(u, v)$ are the velocity components in the along-
FIG. 4. Low-pass filtered current time series from the North Gorgon location. Cross-shelf and alongshelf (bold line) components at depths 23, 89, and 205 m are shown. The water depth is 208 m.
Fig. 5. Low-pass filtered time series of water levels from North Rankin and North Gorgon locations in 123-m and 208-m water depths respectively. Some diurnal tidal oscillations are also evident in the plots.

Fig. 6. Time series of tidal-residual sea levels at Broome, Dampier, and King Bay.

\[ P_A = P_c + (P_n - P_c) \exp[-(r_m/r)^b]. \] (3.4)

Here to model Tropical Cyclone Jane we let \( P_n = 1012 \) hPa, being the environmental pressure, \( P_c = 947 \) hPa, being the cyclone center pressure, and \( r_m = 65 \) km, being the radius of the maximum winds, while \( b \) is defined by

\[ b = 1.5 + (980 - P_c)/120, \] (3.5)

which provides a scaling on the profile shape. The symmetric, gradient-level azimuthal wind component at radius \( r \) is...
and $K$ denote the locations of Broome, Dampier, and King Bay, respectively.
The grid points denoted by R and G represent the North Rankin and North Gorgon
locations respectively. The computational domain shown in Fig. 7 has the
dimensions of 1500 km and 600 km in the longshore and offshore directions respectively.

The wind stresses are obtained from the wind velocity $U$, $V$ by using the quadratic drag law

$$\tau_{b}^{(x)} = c_{d} \rho \sqrt{U^2 + V^2}.$$  \hspace{1cm} (3.7)

where $W = (U^2 + V^2)^{1/2}$ is wind speed and $c_{d}$ is the drag coefficient. Following Smith and Banke (1975) and
Frank (1984), the drag coefficient is

$$c_{d} = \begin{cases} 0.63 + 0.066W, & \text{if } 0 \leq W < 25 \text{ m s}^{-1} \\ (2.28 + 0.033(W - 25)) \times 10^{-3}, & \text{if } W \geq 25 \text{ m s}^{-1} \end{cases}.$$  \hspace{1cm} (3.8)

The bottom stress is modeled by the usual quadratic friction law

$$\tau_{b}^{(w)} = C_{D} \sqrt{u'v'} + \mu(u, v).$$  \hspace{1cm} (3.9)

But importantly we allow here for a wind-wave-induced enhancement of the drag coefficient $C_{D}$ in the nearshore region. This is achieved by setting $C_{D} = F(X)C_{D0}$, where $C_{D0} = 1.5 \times 10^{-3}$ and $X$ is a variable that depends, inter alia, on the wind wave amplitude and the local water depth. The specific formulation for determining the enhancement function $F(X)$ is based on the theory of Grant and Madsen (1979), and our implementation of this theory is described in detail in Tang and Grimshaw (1996b). It is sufficient to note here that $F(X) \geq 1$ and in the absence of any wind waves, $X = 0$ and $F(X) = 1$. In deep water, $F(X) \approx 1$, but in the nearshore shallow water $F(X)$ can be considerably greater than 1. For instance, for a wind wave of amplitude 1.0 m and period 10 s, and a current magnitude of 0.8 m s$^{-1}$, $F(X)$ increases to 10 at a depth of 5.0 m. In our implementation here the wind wave amplitude is determined by assuming a quasi-steady-state balance between the wind stress and the wind wave dissipation (for further details, see Tang and Grimshaw 1996b).

The results of our numerical simulations are shown in Figs. 8, 9, and 10, where in each case we show contour plots of (a) sea-level elevation $\zeta$ and (b) the velocity field. In Fig. 8 we show the case when there is no wind wave enhancement of the bottom friction and $C_{D}$ in (3.10) is given by $C_{D} = C_{D0} = 1.5 \times 10^{-3}$, while Fig. 9 is similar but with $C_{D} = C_{D0} = 2.5 \times 10^{-3}$. Then, in Fig. 10 we show the case when the bottom friction is wind wave enhanced, and $C_{D} = F(X)C_{D0}$ with $C_{D0} = 1.5 \times 10^{-3}$.

In discussing these results we first compare the simulations shown in Figs. 8 and 9. Both these cases are for constant bottom drag coefficient $C_{D} = C_{D0}$, and the only difference is the greater friction in Fig. 9. We note that with the greater level of friction the coastal sea level is in broad agreement with the observations, but the offshore currents are considerably less. Note that coastal sea level was recorded at Broome, Dampier, and King Bay, denoted by the points B, D, and K respectively in Fig. 7, while the offshore currents were recorded at the sites North Rankin and North Gorgon denoted by the points R and G respectively in Fig. 7. Correspondingly when the friction is reduced to the value shown in Fig. 8, the numerical simulations overestimate the coastal sea level, but better agreement is ob-
Fig. 8. Results from the numerical simulation when the bottom-stress drag coefficient is set at the constant value $C_D = 1.5 \times 10^{-3}$. (a: top) Contours of the sea level elevation, with contour intervals of 10 cm and negative values shown with dashed contour lines. (b: bottom) The velocity field (in cm s$^{-1}$) with the scale being indicated by the maximum value, $V_{max}$. 
FIG. 9. As in Fig. 8 but the bottom-stress drag coefficient is set at the constant value $C_{D0} = 2.5 \times 10^{-3}$. 
FIG. 10. As in Fig. 8 but the bottom-stress drag coefficient is wind wave enhanced with an offshore value of $C_{D0} = 1.5 \times 10^{-3}$; (c) the wind wave amplitude, with contour intervals of 50 cm.
tained with the offshore currents. This dilemma is a common feature of many storm surge simulations (e.g., Hearn and Holloway 1990; Fandry and Steedman 1994). A possible resolution is offered here in Fig. 10, where the bottom friction is wind wave enhanced with the effect of increasing the friction in the nearshore region. The wind wave amplitudes used for this enhancement are shown in Fig. 10c. We now see that the coastal sea levels in the numerical simulations are at realistic levels, as also are the offshore currents. It is pertinent to point out here that the value of $C_{dp0} = 1.5 \times 10^{-3}$ here in Fig. 10 (and in Fig. 8) is a generally accepted offshore value, but the higher value of $C_{dp0} = 2.5 \times 10^{-3}$ is instead commonly used in storm surge models (see, e.g., the review by Das 1994).

Next, we turn to a more quantitative comparison with the current observations at North Rankin and North Gorgon. Here we use the numerical results shown in Fig. 10 and we select the points R and G shown in Fig. 7 to represent North Rankin and North Gorgon respectively. Each is chosen to lie on the grid point that most closely corresponds to the location of the observation site, with respect to longshore distance from the tropical cyclone path, offshore distance, and stillwater depth. Importantly, the points R and G are 150 km apart in the longshore direction. In Fig. 11a we show a time series for the total current magnitude at the points R and G. To compare these with the observational data shown in Figs. 3 and 4 we show in Fig. 11b the observed depth-integrated total current, from which we see that the peak depth-integrated currents at North Rankin and North Gorgon are 0.45 and 0.59 m s$^{-1}$ respectively. We see that the peak magnitudes are in good agreement. Further, we note that there is nearly a 10-h phase lag between these peaks in our numerical simulations, again in good agreement with the observations. It is pertinent to note here that analogous simulations with $r_m$, the radius of maximum wind, altered to 50 km (or 75 km) produced weaker (or stronger) currents respectively, but the phase lag was approximately unchanged, indicating the essentially linear nature of the flow field of coastally trapped waves. Similar time series for the sea-level elevation at points R and G (not displayed here) show a similar phase lag, even though the amplitudes are small and difficult to distinguish in the numerical output.

Because the observations sites at North Rankin and North Gorgon are both outside the region of direct wind forcing, we can infer that the response there is due to coastally trapped waves (see, e.g., Tang and Grimshaw 1995). The most direct evidence of this is the phase lag between the two sites, and to relate this to coastally trapped wave theory, it is necessary to determine the
group velocities for coastally trapped waves at these locations. Here, however, a difficulty arises due to the marked narrowing of the shelf between these sites (see Fig. 1). We choose a representative topography between points R and G in Fig. 7 to determine the dispersion relation for the frequency $\sigma$ as a function of the longshore wavenumber $k$ (see Fig. 12) but somewhat closer to the point R. Note that the group velocity can generally be expected to be an increasing function of $fL$ where $L$ is the shelf width but will also increase as the shelf slope increases. Thus, although the shelf narrows between the points R and G, the effect on the group velocity will be partially compensated by the corresponding increase in slope. Next we must estimate the appropriate wavenumber to use in calculating a representative group velocity. From an examination of the model results in Fig. 10, and bearing in mind that the radius of the maximum winds ($r_m = 65$ km) will be one of the factors determining the predominant length scale, we choose the wavenumber $k = 5 \times 10^{-6}$ m s$^{-1}$. Then from Fig. 12 we estimate that the group velocity for the first-mode shelf wave is about 6 m s$^{-1}$, while the group velocity for the second mode is about 4 m s$^{-1}$. The speed for either mode is consistent with the observed speed of the storm surge, namely, 5 m s$^{-1}$. To distinguish between the modes, it is necessary to examine the modal structures. However, we find that in the vicinity of the points R and G the first and second modes have somewhat similar structures. Thus, the best we can infer is that the storm surge is composed of either the first or the second mode, or possibly a mixture of both modes, and we note that this is consistent with the analysis of Tang and Grimshaw (1995). Further the relatively small signal found in the sea-level elevation at the points R and G in contrast to the strong signal in the current field is consistent with our interpretation of the storm surge at these locations being due to the propagation of a continental shelf wave.

While our main purpose is to compare the observations and simulations for the offshore observations at R and G since these sites are best suited for the detection of coastally trapped waves, it is also of interest to compare the observations and simulations for sea level at some coastal sites. Hence, in Fig. 13 we show the sea levels for the points B and K respectively, these being
these observation sites are located well away from the mode continental shelf waves. We reiterate that both interpretation of the storm surge at these sites as low-gon, and we have argued that this is evidence for the surge at the two sites of North Rankin and North Gor-
gand the simulation is the phase lag between the storm water. The most striking feature of both the observations the feature of being wind wave enhanced in shallow equations with a quadratic bottom stress law that has

storm surge based on the depth-integrated shallow-water

described the results of a numerical simulation of this

generated by TC Jane. Then in section 3 we have de-

4. Discussion

In this paper we have described in section 2 some offshore current meter observations of the storm surge generated by TC Jane. Then in section 3 we have described the results of a numerical simulation of this storm surge based on the depth-integrated shallow-water equations with a quadratic bottom stress law that has the feature of being wind wave enhanced in shallow water. The most striking feature of both the observations and the simulation is the phase lag between the storm surge at the two sites of North Rankin and North Gor-
gon, and we have argued that this is evidence for the interpretation of the storm surge at these sites as low-
mode continental shelf waves. We reiterate that both these observation sites are located well away from the

TC track and, hence, there is a reasonable expectation that the response there is due to the generation of low-
mode continental shelf waves (Grimshaw 1988; Tang

and Grimshaw 1995). Further, the predominance of the current signal at these sites over the sea-level signal, both in the observations and in the simulations, lends further support to the interpretation as a continental shelf wave. We also note again that a quantitative estimation of the theoretical group velocity of the first two continental shelf wave modes is consistent with the observed and simulated phase lag. It is also pertinent to comment here that the evidence for the presence of low-mode continental shelf waves is not so clear in the coastal sea-level data. This is mainly because the available data are from locations much closer to the TC forcing region, but is also due to the fact that continental shelf waves are better identified from current data.

It is noteworthy that we have been able to numerically simulate the storm surge and identify the presence of continental shelf waves using just the depth-integrated shallow-water equations, even though the observations show that the currents have significant vertical structure. We attribute this to the fact that this vertical structure is associated with internal frictional processes, as dem-

onstrated by Hearn and Holloway (1990), rather than with density stratification. If so, then a depth-integrated barotropic numerical model is appropriate for the ident-
ification of the continental shelf waves. We note that provided the friction is not too large, and indeed in our numerical model, the friction at these offshore sites is not large, then the structure of the continental shelf wave modes is largely preserved, and the main effect of fric-
tion is to induce a cross-shelf phase difference (Brink

and Allen 1978, 1983; Power et al. 1989). However, this phase difference is not readily apparent in our nu-

merical simulations, mainly because it is masked by the much more significant distortions of the flow field caused by the large topographic variability. We note again, for instance, that this topographic effect is presumably responsible for the larger signal at the more remote site due to the narrowing shelf there.

Temperature time series at North Rankin and North Gorgon show that the buoyancy frequency $N$ is about 1.4 $\times 10^{-2}$ $s^{-1}$ and 1.2 $\times 10^{-2}$ $s^{-1}$ at these two sites respectively for conditions before and after the storm surge. During the cyclone, there is some downwelling of the stratification, but no sign of significant vertical mixing. Although this is quite strong downwelling, we reiterate again that it does not appear to have a significant role in determining the structure of the storm surge. This is because the continental shelf waves generated here are long waves, for which the stratification parameter $S = N^2 H^2/\nu L^3$ is quite small (see, e.g., Huthnance 1978). Here $H$ and $L$ are typical depth and cross-shelf length scales. For instance, at North Rankin we set $H = 120$ m, $L = 100$ km, and find that $S = 0.11$, while for North Gorgon we set $H = 200$ m, $L = 100$ km, and then $S = 0.23$. However, it is useful to note that the
effect of even a small stratification parameter is to increase the phase speed slightly, and this would bring a slight improvement of our phase speed estimation in relation to the observations.

Next, we note that the North West Shelf is well known for its very large tides. Indeed, the barotropic tidal currents at North Rankin and North Gorgon have spring tidal flow of around 31 and 7 cm s\(^{-1}\) respectively. Thus, we can anticipate that there will be a strong nonlinear interaction between the storm surge and the tides, which in general can be expected to modify the numerical results obtained here. However, it has been shown by Tang et al. (1996) that the dominant mechanism in this nonlinear interaction is through the quadratic bottom stress term. In the present simulations, this term is quite small at offshore observation sites of North Rankin and North Gorgon, so, although we expect this tidal interaction to modify our numerical results somewhat at the coastal sites, we do not expect a very significant effect at the offshore sites. Thus, at these offshore observation sites, the tidal signal can effectively be linearly added to the storm surge signal, and indeed, this was anticipated in our filtering of the observational data to remove the tidal signal.

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