New Zealand Journal of Geology and Geophysics

ISSN: 0028-8306 (Print) 1175-8791 (Online) Journal homepage: http://www.tandfonline.com/loi/tnzg20

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To cite this article: R. Agnew (1966) Storm tides in the Tasman Sea, New Zealand Journal of Geology and Geophysics, 9:3, 239-242, DOI: 10.1080/00288306.1966.10422812

To link to this article: http://dx.doi.org/10.1080/00288306.1966.10422812

Published online: 12 Jan 2012.

STORM TIDES IN THE TASMAN SEA

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(Received for publication 5 October 1965)

Predicted spring tides in the days following new moon (perigee) on 28 July 1965 were considerably exceeded on the west coast of the North Island. The surface weather maps showed high pressures over south-east Australia, a south-westerly airstream covering New Zealand, and a deep depression over the Chatham Islands: the whole system moving slowly eastwards. Before examining meteorological factors which may have contributed to the record high-water levels experienced in both Manukau and Kaipara Harbours on 31 July, consider the automatic tide gauge data.

Successive low and high tide levels in feet above Auckland sounding datum are here tabulated for the port of Onehunga, Manukau Harbour.

——	30 July 1965			31 July 1965				1 Aug. 1965
Time (N.Z.S.T.)	0555	1202	1812	0012	0656	1251	1909	0050
Predicted height	2·2	14·3	2·5	14·6	2·3	14·2	2·7	14·5
Measured height	2·5	15·9	2·2	17·2	3·3	15·9	2·2	15·0
Difference (ft)	+0·3	+1·6	−0·3	+2·6	+1·0	+1·7	−0·5	+0·5

The first high water on 31 July equalled the Onehunga maximum of 21 June 1947, and may have exceeded it, for the tide gauge was near the limit of its mechanical range. Incomplete tidal records (H.W. only) from Ruawai, Kaipara Harbour, show a difference of 0·7 ft on 30 July, 1·7 and 1·1 ft on 31 July, and 0·8 and 0·2 ft on 1 August, all above the expected heights. In fact, the first high water on 31 July 1965, was 0·5 ft above previous maxima at Ruawai, recorded in 1939, 1950, and 1964.

Both the above stations are relatively far inland, but records are lacking from the open coast. There is no tide gauge permanently installed at the Manukau Heads, while the New Plymouth tide recorder was out of action at the end of July 1965.

Discounting fresh water and seismic disturbances, how far can present knowledge go in explaining these abnormal sea-levels? Let us first check for variations in atmospheric pressure. The synoptic weather chart for midnight on 30/31July 1965 gave $p = 1{,}003$ mb over the Manukau, and $p = 1{,}006$ mb north of the Kaipara entrance. Taking 1,012 mb to represent the seasonal mean pressure (cf. Gilmour, 1963, at Bluff), the deficit $\Delta p = 9$ mb at Onehunga and 6 mb at Ruawai. The corresponding rises in water level are 0·3 ft and 0·2 ft respectively.

If we had an atmospheric pressure disturbance travelling at velocity U over water of uniform depth h, the hydrodynamic factor $[1 - U^2/(gh)]^{-1}$ must be applied to the equilibrium elevation (Proudman, 1953), because the pressure disturbance is not stationary. This factor is practically unity for the Tasman Sea, but could treble the equilibrium elevation crossing the continental shelf, and double it again by total reflection at the coast, according to Abraham (1961). However, available records of our July storm show no evidence of a sudden rise in level or of the long wave motion usually accompanying such a surge; the only suspicious circumstance is the disappearance of a cold front straddling the South Island at 6 p.m. on 30 July and absent from the midnight weather map.

Typical depths for the Tasman Sea are 1,000–3,000 fathoms. The weather maps for 6 p.m. on 30 July and midnight on 30–31 July indicate an easterly displacement of about 180 nautical miles on corresponding isobars, thus:

$$U = \frac{180 \text{ nautical miles}}{6 \text{ hr}} = 30 \text{ knots} = 51 \text{ ft/sec}$$

when taking the smaller value for depth, h = 1,000 fathoms = 6,000 ft, and $g = 32$ ft/sec^2, the factor $[1 - U^2/(gh)]^{-1} = 1{\cdot}01 = 1$. There would still be only a 5% departure from unity if the disturbance travelled at 100 ft/sec. On approaching the coast decreasing depths increase the factor, thus there is a 15% departure from unity at the 100 fathom line, approximately 20 miles offshore. At the coast, $h = 0$ and the factor becomes infinite (*see* Abraham, 1961). More elaborate theoretical work for the case of a barrier summarised in Proudman (1953, pp. 297–300).

The most likely cause of "set-up" at the coast was the strong onshore wind. From the isobar spacing at midnight it is estimated that the gradient wind was SW 65 knots, less than 100 miles from the Manukau Heads. Allowing for frictional influence, the "surface" wind speed at midnight must have been about 46 knots at the standard elevation of 10 m. The ship *Karepo* reported 50 knot winds at midnight 250 miles south-west of the Manukau Heads. Land-based anemometer records suggest somewhat lower values: at noon and midnight on 30 July an average windspeed of 54 knots was measured at Cape Reinga, with gusts to 75 knots next morning; Kaitaia, Dargaville, and Manukau Heads registered 24 knots; at Whenuapai the windspeed was over 20 knots, with gusts to 45 knots, while 66 knots was reported on the Auckland Harbour Bridge. It will be assumed that windspeed normal to the Northland coast averaged 40 knots at standard elevation on the evening of 30 July 1965.

To estimate the "set-up" due to wind drag at the water surface, bottom profiles bearing 240° (true) from the entrances of both harbours were plotted from the North Island Bathymetric Chart (N.Z. Oceanographic Institute Miscellaneous Series 13) and these lines were continued into shallow water using Admiralty Chart No. 2,543, choosing channel soundings in the general direction of the wind. These composite profiles were then schematised by a number of "steps" up the continental slope and across the shelf into each harbour. Regarding any step as a basin of length Δx and uniform depth h, its surface slope

$$\Delta h/\Delta x = (\tau_s + \tau_0)/(\rho_w \cdot g \cdot h)$$

after sufficient time for the circulation (assumed two-dimensional) to become steady. This "set-up" formula is readily derived from first principles. Assume a steady state has been achieved and balance the horizontal forces on the liquid. The pressure difference $\rho_w\, g\,.\,\Delta h$ acting over depth h gives force from right to left $= \rho_w\, g\,.\,h\,.\,\Delta h$. Wind drag on surface and bottom friction opposing return circulation give force $(\tau_s + \tau_0)\,.\,\Delta x$ to the right. Therefore

$$\rho_w\,.\,gh\,.\,\Delta h = (\tau_s + \tau_0)\,.\,\Delta x$$

$$\therefore \frac{\Delta h}{\Delta x} = \frac{\tau_s + \tau_0}{\rho_w\, gh}$$

The surface shear stress $\tau_s = C\,.\,\rho_a.\,W^2$ where W = windspeed at 10 metres elevation, and the bottom shear stress τ_0 may be neglected if the return current is weak, giving the simple formula

$$\Delta h/\Delta x = C(\rho_a/\rho_w)(W^2/gh)\,.$$

Here the ratio of salt water density to air density $\rho_w/\rho_a = 830$ at 12°C and the drag coefficient $C = 0{\cdot}0023$ according to Francis (1959). The set-up, or difference in water levels between the ends of such a basin is readily deduced to be

$$\Delta h\ (\text{ft}) = 0{\cdot}40\ \frac{\Delta x\ (\text{nautical miles})}{h\ (\text{fathoms})}$$

at the steady wind speed $W = 40$ knots. Continuity of the water surface gives the necessary connection between each step. Starting with an unchanged water surface over 100 miles offshore, the increments Δh were added to get the cumulative rise of water level inside each harbour, viz 1·2 ft at Onehunga and 0·8 ft at Ruawai. The formula for surface shear stress in terms of wind speed can be taken as a definition of C, whose numerical value must be obtained from experiments, such as those summarised in Francis (1959). Definitions of drag coefficient and other measurements of set-up do not conflict with these conclusions.

In passing, we should note that the wind set-up would be greater if stratification in the New Caledonia Basin confined the wind-driven circulation to the upper layer, thus reducing the effective depth, and it would also be greater if the storm caused an unusually rough interface between air and water, thereby increasing the drag coefficient.

Not all the water carried forward by the wind drift would return as a bottom current. For example, water heaped against the Taranaki (Egmont) coast would tend to escape south-eastwards into Cook Strait and northwards as a longshore current superimposed on the regular tidal streams and ocean currents. The raising of sea level by this additional water off Northland is probably negligible, but what of the geostrophic effect? A northward flow at mean velocity v would develop a transverse slope $(2\omega v\,.\,\sin\phi)/g$ due to the Earth's rotation at angular velocity $\omega = 7{\cdot}29 \times 10^{-5}\ \text{sec}^{-1}$.

To find the order of magnitude, assume $v = 0{\cdot}5$ ft/sec over a 12 mile width of the continental shelf at latitude $\phi = 37°$ S. Then the calculated super-elevation at the coast is $-0{\cdot}1$ ft, which is relatively small and in the opposite sense to the observed tilt.

Adding the above estimates of probable rise in sea level, we get for Onehunga $0{\cdot}3 + 1{\cdot}2 - 0{\cdot}1 = 1{\cdot}4$ ft, and for Ruawai $0{\cdot}2 + 0{\cdot}8 - 0{\cdot}1 = 0{\cdot}9$ ft, to be compared with the measured rises of 2·6 ft and 1·7 ft respectively. Thus our estimates are little more than half the observed changes in water level, and we appear to have exhausted those "steady state" explanations at present capable of formal solution. Qualitatively, there is another contributing factor, the set-up due to coastal obstruction of mass transport in the direction of wave propagation. This would be large for short step waves, such as our 20 ft west coast breakers, but its magnitude can only be guessed.

The time-dependent factor remains to be investigated. During 30 July the wind veered from 230°–200° at Cape Reinga, and from 280°–240° at both Kaitaia and Whenuapai. This change in direction would hardly be sufficient to generate a seiche, but a study of simultaneous departures from normal sea level on both sides of the Tasman might reveal storm-induced oscillations of semi-diurnal period. Local oscillations having shorter period are readily detected on the tide curve, but long period motion is hard to isolate. The Onehunga data could be interpreted as a semi-diurnal wave of amplitude 1 ft (approximately) superimposed on a "steady" rise and fall of sea level over several days.

Are there storm tides in the Tasman Sea, undetected because of the coarse network of recording stations and failure to correlate their results? If so, can their height and frequency be predicted? This is not just an academic question, for abnormal sea levels threaten lives and property—the July storm this year caused thousands of pounds worth of damage to stopbanks and harbour facilities. How soon can we expect a disaster from coincidence of astronomical and meteorological or seismic forces?

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