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A Comparison Of The Collinear Mean And The Mean Sea Surface As Reference Levels For Altimeter Analysis In The Southern Pacific Ocean

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A COMPARISON OF THE COLLINEAR MEAN AND THE MEAN SEA SURFACE AS REFERENCE LEVELS FOR ALTIMETER ANALYSIS IN THE SOUTHERN PACIFIC OCEAN

by

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ABSTRACT

Satellites are used to study various oceanic phenomena, including sea surface temperature, color, and sea height variability (which is related to oceanic currents).

In observing sea height variability from radar altimeters, three methods of analysis can be applied. They are the collinear or repeat track method, the cross-over difference method, and the mean sea surface method. All three are designed to remove the effects of geoid undulations from the altimeter records. This paper compares the collinear method with the mean sea surface method in a small geographic area. Both methods analyze the tracks of collinear altimeter data by subtracting a mean pass and a best fit quadratic curve. Geoid variability is effectively removed by subtracting the mean pass, while the quadratic curve removal eliminates remaining long wavelength orbit error and tidal signals. The difference between the two methods is in the mean pass that is subtracted. The collinear mean is computed from a point by point average of the collinear data taken from one month of the SEASAT mission. The mean sea surface method subtracts a mean pass derived from a global mean sea surface developed from cross-over data from the 3.5 year GEOS-3 and the full 3.5 month SEASAT missions.

After this processing, the residual data from these methods are compared in the spatial and spectral (wavenumber) domains. In the region of 10^{-2} cycles per kilometer, a spectral analysis will yield energy primarily from oceanic variability and eddy currents. Over all wavenumbers, the two methods compared qualitatively. Quantitatively the mean sea surface residuals had an order of magnitude higher variance than the collinear mean residuals. This was true at all wavenumbers except in the region of 10^{-2} cycles per kilometer, where the mean sea surface residuals still had more power, but only by a factor of two or three. This is significant because this region is thought to contain the peak amount of power from oceanic signals.

With more data it may be possible to completely and accurately define a mean sea surface suitable for the detection of various oceanic phenomena with a single pass of a satellite altimeter. It is concluded from this study that this cannot be done with the presently available mean sea surface.

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SYMBOLS

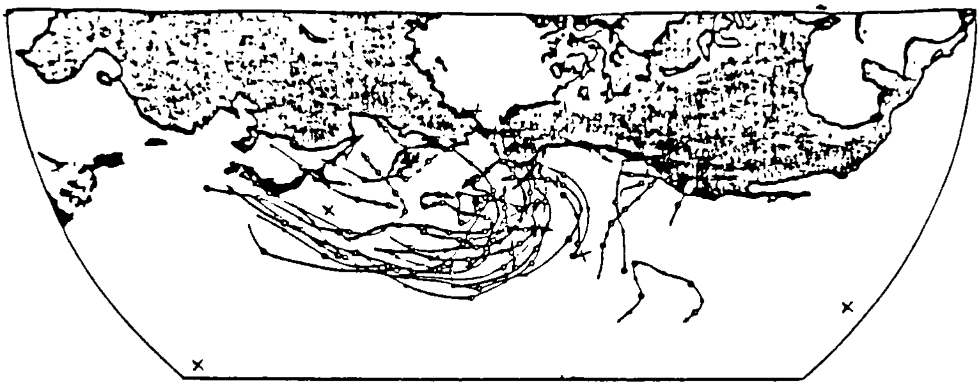
C	sum of corrections
f	Coriolis parameter
g	local gravitational acceleration
$h_e(X)$	computed satellite altitude at position X above the ellipsoid
h_o	observed altimeter measurement
h_{ss}	sea surface height above the reference ellipsoid
p	pressure
T_o	ocean tide
u	velocity component in the x direction
v	velocity component in the y direction
V_o	velocity at some position x for some Z_o
Z_o	arbitrary reference level
ρ	density of sea water
σ_m	measurement noise
Θ	latitude
Ω	Earth's rotation rate

INTRODUCTION

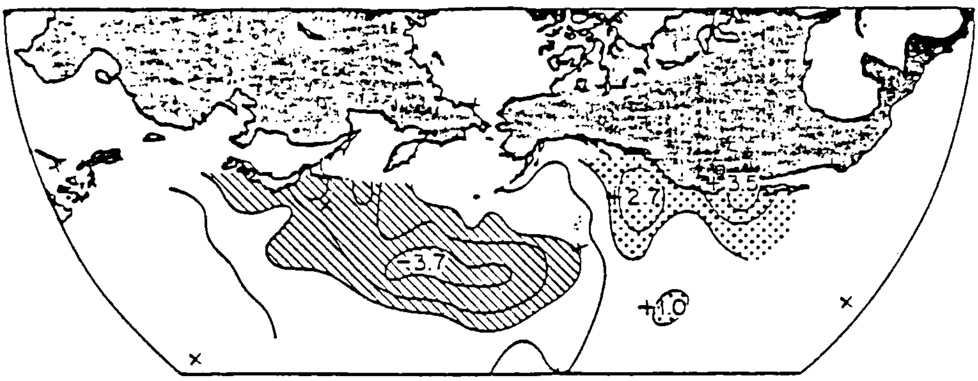
The oceans are vast turbulent bodies of water. The currents in them are continually changing in magnitude and direction. The study of ocean dynamics is paramount to our understanding of many oceanographic and atmospheric phenomena. Among other uses it would be possible to explain and follow ocean circulation to derive oceanic heat balances between polar, mid-latitude and equatorial regions, help ships travel more efficiently, and help the fishing industry to locate schools of fish. Global or basin-wide observations are important to understand the earth's climate and ocean-atmosphere interactions. (Stewart, 1985).

As an example of the importance of ocean dynamics, correlations can be drawn between the amount of energy stored regionally in the ocean's surface layers and terrestrial weather patterns. Climatologists have seen that surface winds and sea surface temperature affect weather over land masses. They can also draw direct conclusions from satellite data showing how storms follow changes in sea surface temperature. An example of this from Stewart (1985) is shown in Figure 1.

Physical oceanographers have recently spent considerable time observing fluctuating currents in the ocean rather than the mean circulation. This is because it has been found that a significantly greater amount of kinetic energy is in the variable flow. Included in this variability are fluxes of energy and momentum due to Rossby or planetary waves, deep



Storm Tracks - January 1977



Sea Surface Temperature - Winter 1977

Figure 1. The first picture shows centers of major sea-level cyclones over the north Pacific (Mariner's Weather Log, 1977).

The second shows departure of sea surface temperature from the seasonal mean for December, January and February in deg. F. Shaded areas are in excess of 1°F. (Namais, 1978).

ocean tides, and mesoscale eddies. Figure 2 illustrates the kinetic energy distribution between the mean flow and the fluctuating flow on a global scale. (Wrtki, Magaard and Hager, 1976).

There are many problems encountered in gathering oceanographic data. The main reasons are related to the size of the ocean, the access we have to it, and its turbulent nature. Mobility is hampered and, until recently, data collection has relied on ocean vessels. There is only a small fleet of oceanographic ships, so most of the data has been obtained from commercial and military vessels. Often this data is unreliable, and typically these ships avoid storms and the more turbulent sections of the ocean.

This is understandable, yet unfortunate, because dynamic areas of the ocean yield some of the most valuable information. Some of this data could be used to study heat, mass, and momentum exchange in a high wind. For example, energy exchange between the wind and ocean varies as the wind speed cubed, hence an 80 meter per second (ms^{-1}) wind transfers as much energy in one day as an 8 ms^{-1} trade wind blowing for almost three years. (Stewart, 1985). Figure 3 illustrates monthly distributions of surface observations made. It shows data concentrated in small areas.

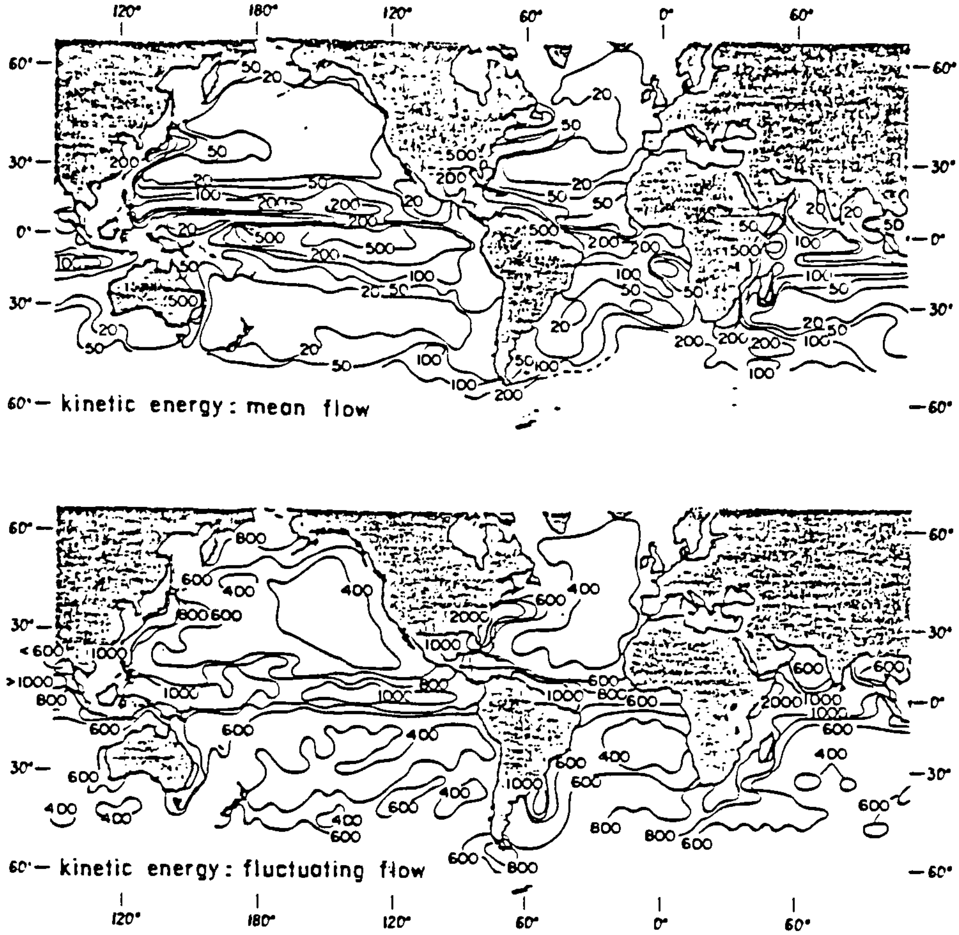


Figure 2. Kinetic energy per unit mass (cm^2/sec^2) of sea surface currents averaged over 5 degree squares. The first picture gives the mean flow kinetic energy, while the lower shows kinetic energy of fluctuating currents. (Wyrtki, Magaard and Hager, 1976).

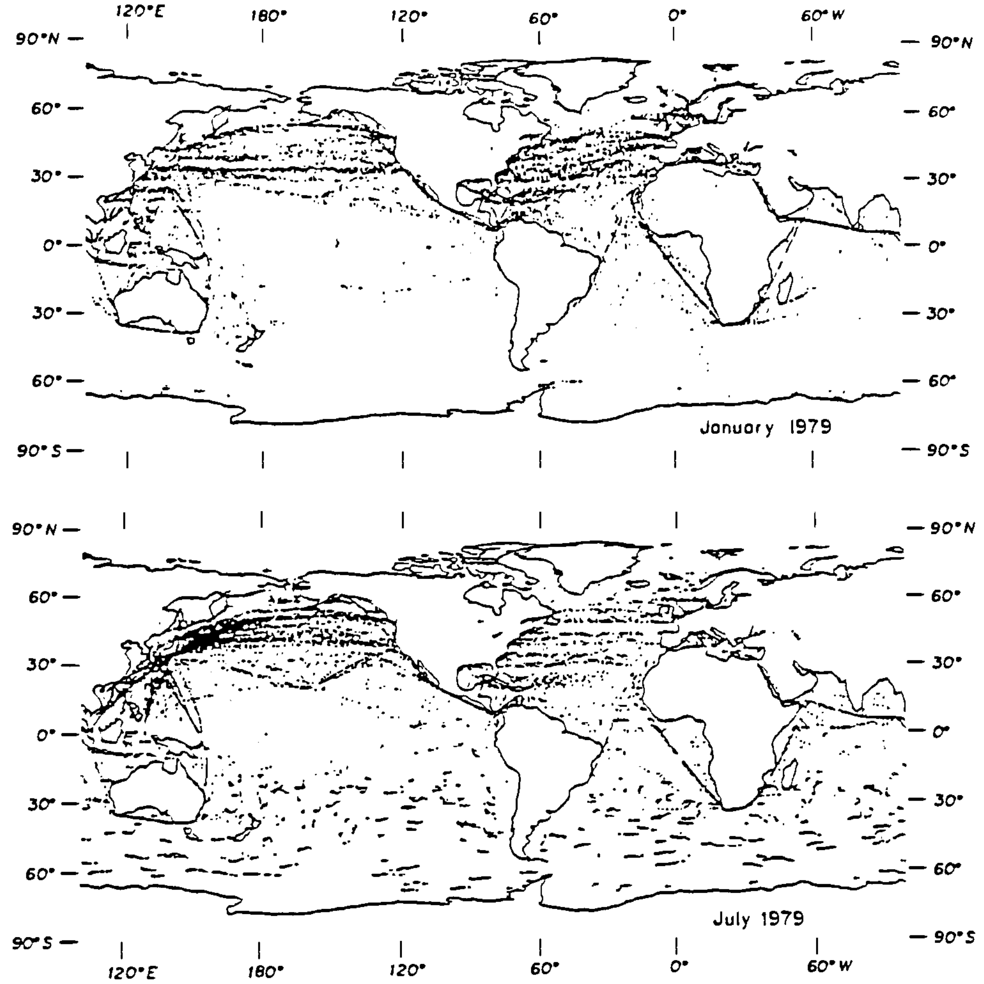


Figure 3. Monthly distribution of surface observations made by ships and buoys, received by the NOAA Pacific Marine Environmental Group in Monterey, California (plot by Douglas McClain). Note the increased observations in the Southern Ocean in July 1979, produced by drifting buoys deployed for the Global Weather Experiment.

Recent advancements in space technology and instrumentation have enabled NASA and other agencies to orbit satellites whose missions are to collect and transmit a wide range of oceanographic data. There are many advantages to using satellites in oceanography. Various types of data can be collected on a global scale, including color, temperature, and altitude measurements. All of the readings are possible in a relatively short period of time, and at very reduced operating expenses. Finally, it is beneficial to use satellites because the data can be monitored continually and reliably in near real-time.

SATELLITE ALTIMETRY

In recent missions, satellites have been using radar altimetry to observe sea height variations.

There are some unique terms used in satellite altimetry and oceanography that must be defined. One such term is the geoid. It can be described as the equipotential surface the ocean would assume if only gravitational effects of solid earth, water, and atmosphere were considered. To further define the geoid one must consider a non-homogeneous earth as it is today, but with no atmosphere. Also, if the water were to rotate at the same speed as solid earth, then it would flow until the whole ocean surface would be an equipotential surface. Given this, the surface would be smooth except for the mass concentrations in the solid earth, that would change the gravitational field. Then, adding a corotational atmosphere, the added weight would again slightly change the gravitational field, hence the equipotential surface. The geoid is the equipotential surface that corresponds to the mean sea level.

The geoid is very close to a biaxial, rotational ellipsoid determined by the mean mass of the earth and its rotation. The difference between these two surfaces is referred to as geoid undulation, and the difference in height from the geoid and the sea surface defines sea-surface topography. (Stewart, 1985). Figure 4 depicts the satellite altimeter process, and is specific to the SEASAT mission. (Tapley, 1982).

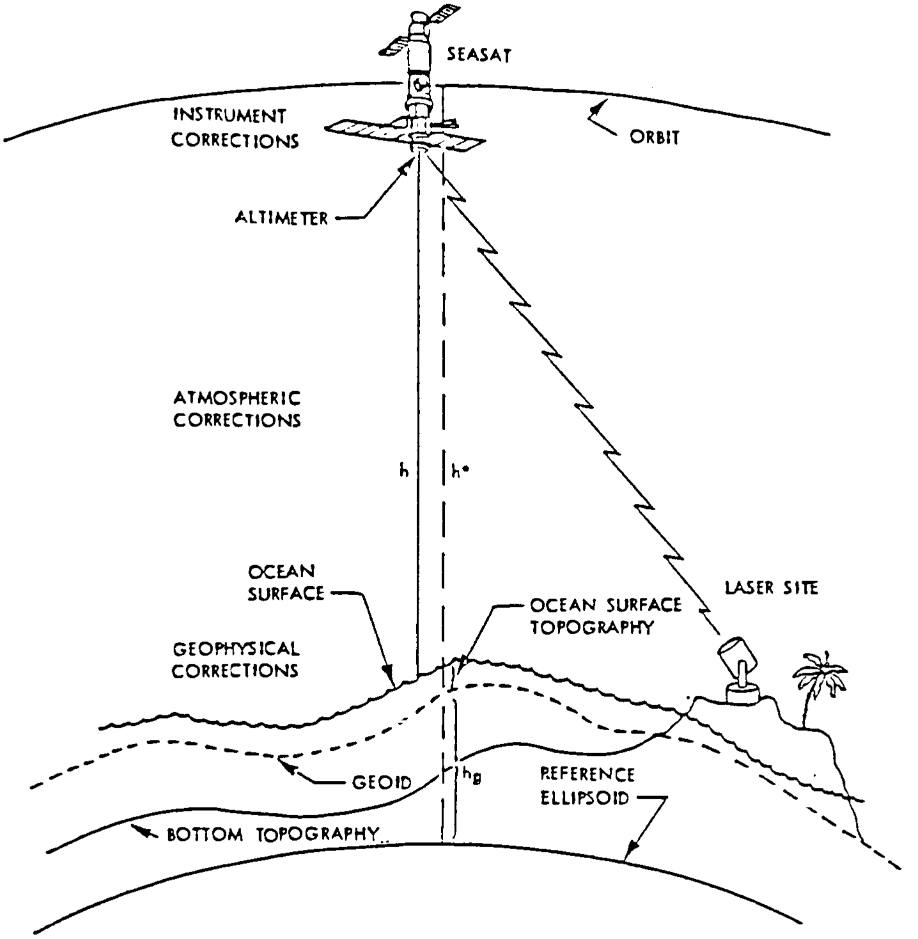


Figure 4. Schematic depicting SEASAT data collection, modeling and tracking system. Various heights and terms specific to this research are defined.

For the calculation of sea height variability, the height of the sea surface is referenced to the ellipsoid because the time invariant geoid is not yet known. The following equation holds:

$$h_{ss} = h_e(X) - h_o - T_o + \sigma_m + C \qquad (1)$$

where: h_{ss} = sea surface height above the reference ellipsoid.

$h_e(X)$ = computed satellite altitude at position X above the ellipsoid.

h_o = observed altimeter measurement.

T_o = ocean tide.

σ_m = measurement noise.

C = sum of all other corrections.

The main cause for error here is from radial orbit error, which in turn is primarily gravity field model errors. This can be coped with by taking orbital tracks of only a few thousand kilometers as discussed later, (Marsh, Cheney, et al., 1984), which restricts our observations to oceanic phenomena of smaller scale.

Elevation changes from the reference ellipsoid in sea surface topography are caused by several phenomenon. The geoid is the largest contributor with undulations that have variability on the order of 100m. The oceanic variability is on the order of 1m (Fu, 1983) and is largely due to surface current fluctuations. The remaining variability is primarily due to errors in radial orbit and tide removal. The orbit and tidal errors are both effectively removed by choosing an arc length near the principal frequency of the radial orbital error as discussed further in the methods section.

Surface currents can be approximated by a geostrophic equilibrium and hydrostatic balance. Using Cartesian coordinates with x positive to the East, y to the North, and z upward, the velocity components (u,v) in the (x,y) directions are related to pressure (p) by:

$$-fv = \frac{-1}{\rho} \frac{\partial p}{\partial x} \qquad (2)$$

$$fu = \frac{-1}{\rho} \frac{\partial p}{\partial y} \qquad (3)$$

$$0 = -\frac{\partial p}{\partial z} - \rho g \qquad (4)$$

where, $f = 2\,\Omega \sin(\theta)$ = Coriolis parameter

$\Omega = 7.272 \times 10^{-5}\ \mathrm{rad.s^{-1}}$ = Earth's rotation rate

θ = latitude

ρ = density of the sea water

g = local gravitational acceleration

p = pressure

Combining the hydrostatic equation (eqn. 4) with the equation for v (eqn. 2) yields (Fomin, 1964:4ff):

$$v(x,z) = \frac{g}{\rho f} \int_{z_0}^{z} \frac{\partial p}{\partial x}\, dz + v_0\,(x) \qquad (5)$$

where: z_0 = an arbitrary reference level

v_0 = an unknown constant of integration depending on z_0, representing the velocity at z_0.

With this relation, sea surface slope and geostrophic surface currents can be studied. The altimeter measures sea surface slopes ($\partial\zeta/\partial x$, $\partial\zeta/\partial y$), and these are directly related to geostrophic surface velocities (u_s, v_s) through:

$$v_s = \frac{g}{f} \frac{\partial\zeta}{\partial x} \tag{6}$$

The velocity at any depth would be:

$$v(x,z) = \frac{g}{\rho f} \int_z^0 \frac{\partial\rho}{\partial x}\, dz + v_s(x) \tag{7}$$

The expressions for u(y,z) are similar. To apply these equations a high degree of measurement accuracy is necessary. Oceanic topography measurement errors should be less than ±10 cm for general ocean circulation studies (Roemmich and Wunsch, 1982).

Strictly applied, geostrophic equilibrium necessitates a balance between pressure gradients and the Coriolis acceleration. However, small deviations in this balance exist and currents evolve. (For a further discussion see Stewart, 1985; Chapter 14).

Satellites measure the sea heights from which surface slopes can be determined. In this way current direction and magnitudes can be obtained. Kinetic energy can then be computed from the sea surface slope and the geostrophic relation. These parameters are more reliable today because, with each new mission, data acquisition has become increasingly accurate. For example, the altimeter on Skylab in 1973 had a precision of 60 cm, GEOS-3 had 25 cm in 1975, while SEASAT improved the precision to within 10 cm, in 1978. (Fu, 1983).

The SEASAT and GEOS-3 altimeter data have been applied in observing mesoscale ocean current variability (wavelengths from 10 km to 1000 km) and to the general oceanic circulation. The analysis of the data yields a variability spectrum over several wavelength bands. Variability with wavelengths less than 100 km is thought to be primarily instrument noise. (Fu, 1983). Variabilities with wavelengths longer than 100 km contain signals primarily from oceanic energies. However, it has been shown that the resolution of SEASAT and GEOS-3 in smaller geographic regions can go lower than 72 km in wavelength. Below 72 km, white noise begins to overpower the residual oceanic energy. (Marks and Sailor, 1986). The geoid energy is strong above this lower limit as well. Remembering that the geoid is time-invariant, subtracting a mean value from each data point will yield a "residual geoid energy" associated with errors in geoid resolution. This residual energy is then small compared to residual oceanic energy, and is generally negligible. The oceanic residuals in the larger ocean basins exhibit a spectral peak in the 250 km wavelength region (Fu, 1983), which corresponds well with the expected wavelength range representing oceanic mesoscale variability with dominant temporal scales of weeks to months. (Robinson, 1983).

METHODS OF ANALYSIS

As previously mentioned, geoid undulations are on the order of 100 m, whereas oceanic variability is only 1 m. In referencing only the sea height measurement on a point by point basis, the deviations would be hidden by the two order of magnitude difference from 1 m to 100 m. However, given that the geoid is time invariant, it is possible to observe variations from the geoid, using the method of altimetric differences at the same earth locations, at different times. Although the geoid has yet to be exactly defined for the entire earth, it can be effectively removed using this differencing method.

There are three methods that use altimetric differences. They are the "Repeat Track" or "Collinear Track" method, the "Cross-Over Difference" method and the "Mean Sea Surface" method.

To compare the mean sea surface method and collinear method, residuals were calculated only from the last 25 days of the SEASAT mission, when collinear data was gathered.

The Collinear Track method is exactly that. It utilizes data from when the satellite travels in the same path and the same direction it had before. The GEOS-3 mission lasted 3.5 years, but did not have defined repeat tracks. SEASAT began data collection in July 1978, and did so until August with an equatorial spacing of 165 km between tracks. On September 5, 1978, SEASAT was maneuvered into an orbit that would repeat its ground track, within 1 km, every three days. In this orbit, it had a

900 km equatorial spacing between adjacent tracks. SEASAT took measurements in this manner until it failed on October 10, 1978. (Tapley, et al., 1982). Between eight and nine sets of collinear data were taken in those last 25 days. Figure 5 gives the ground tracks for the 3 day repeat cycle.

For use in the collinear method, these ground tracks were divided into segments approximately 2000 km long. This distance was chosen to facilitate the removal of radial orbit and tidal errors. At the SEASAT altitude of 800 km the main sources of orbit error are gravity, atmospheric drag, and solar radiation pressure. Gravity dominates these and the principal frequency of the orbit error is once per revolution. To effectively remove this error, the principal frequency dictates a path length of a few thousand kilometers. (Cheney, et al., 1983). In carefully choosing arc lengths deep ocean tide modeling errors can be removed with the same adjustment used for orbit error removal. (Gordon and Baker, 1980 and Douglas and Cheney, 1981.)

Using a quadratic adjustment the arc length for each track was chosen to be 2000 km, and identified by a revolution number. These tracks were repeated every 43 revolutions. The rev. numbers used in this study were based on two separate, but intersecting ground tracks. Their original rev. numbers are 1157 and 1177. Rev. 1157 runs from 58 deg. south, 160 deg. west to 40 deg. south, 177 deg. west in an ascending path. Rev. 1177 descends from 40 deg. south, 164.5 west to 60 south, 184.4 west. They intersect at approximately 48.2 deg. south and 171 deg. west. The ground tracks are shown in Figure 6.

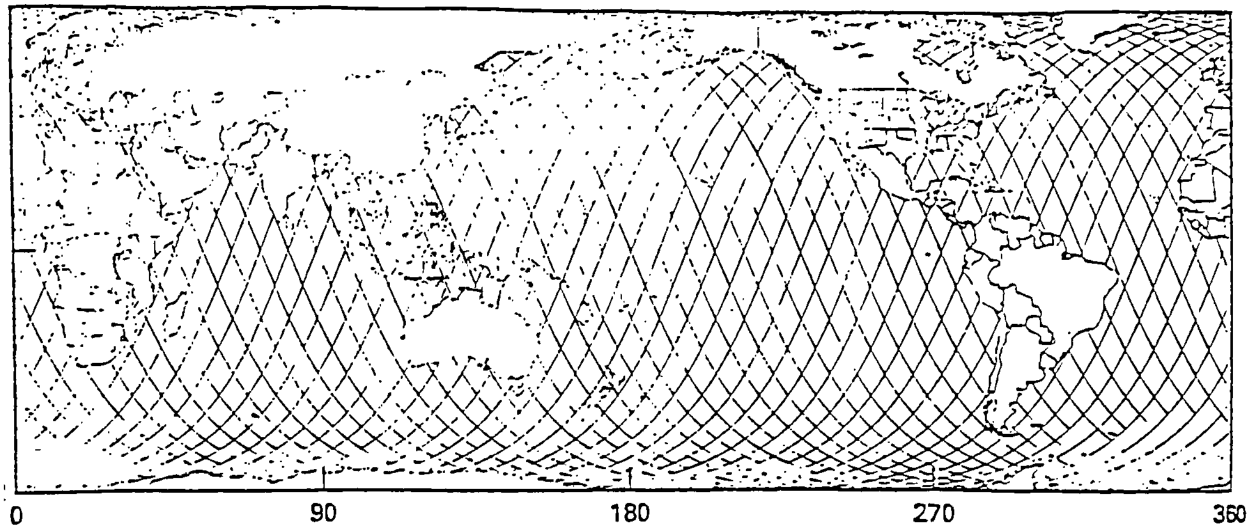


Figure 5. SEASAT ground track during the 3-day repeat cycle. Cross-track separation is approximately 900 km at the equator, 600 km at 45 deg. N. During the final 25 days of the mission (September 15 to October 10, 1978) 8 to 9 sets of altimeter data were obtained along these tracks.

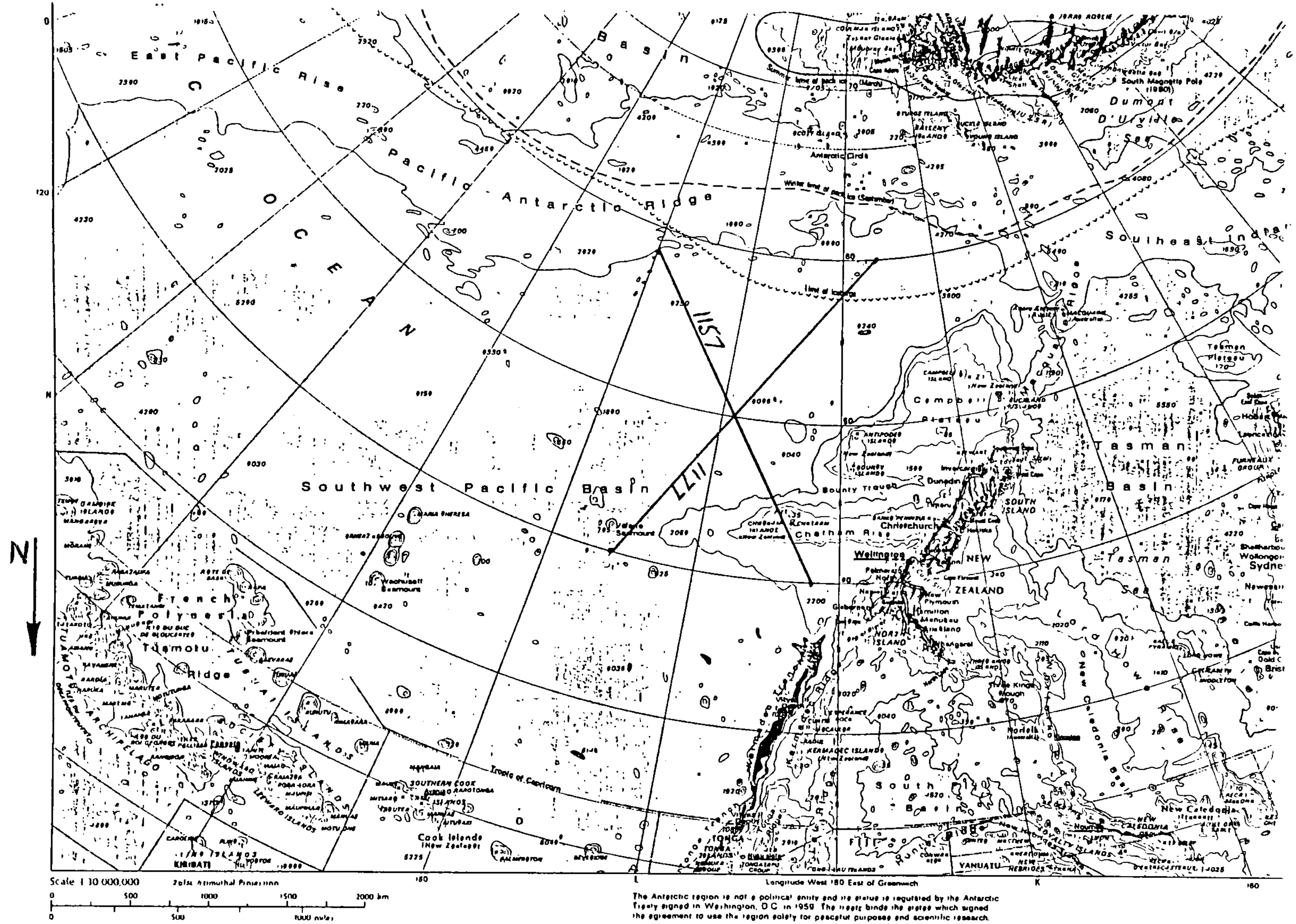


Figure 6. Geographic location of the tracks studied in this paper.

In comparing the collinear and mean sea surface methods, it is necessary to understand how each method reduces data to a set of residuals, which are then studied. The first step in the collinear method is to average the readings from each revolution at each location along the 2000 km ground tracks. This means that 8 points were averaged for each earth location, corresponding to the 8 repeat orbit cycles over the 25 day data collection period. (There were about 6.73 km between each of these locations along the ground track corresponding to a one Hertz sampling rate). This average is then subtracted from each point on each track to effectively remove the effects of the time-invariant geoid and non-varying oceanic signals. Then a quadratic curve was fit and subtracted from each track to quantify and subtract any bias from the remaining radial orbit error and tidal errors, as discussed earlier. The heights that remain are called residuals. They consist of instrument noise and any variations from the fitted curve and collinear mean due to oceanographic variability over the collinear sampling period.

Using the Mean Sea Surface method, the residuals are obtained by subtracting interpolated mean sea surface heights from each point. This surface has been previously defined for the ocean using a sea surface grid of altimetric differences. (Marsh, et al., 1984). After the mean sea surface removal a quadratic curve is found and subtracted, to remove long wavelength orbit and tidal errors, as in the collinear method.

The mean sea surface was obtained using data from the full SEASAT and GEOS-3 missions. SEASAT provided accurate information, but only lasted three months. Data from GEOS-3 was used not only to increase the sampling size, but because the mid-ocean eddy field is thought to have dominant periods of two to three months. This would appear time invariant to SEASAT, thus, biasing the geoid determination. (Cheney, et al., 1983).

In the geographic area of tracks 1157 and 1177, the eddy field should have a dominant period of less than one month (Bryden and Heath, 1985), therefore, the mean sea surface (Marsh, et al., 1984) which is derived from data collected over a period much longer than the 25 day collinear data set, should effectively average out the eddy fluctuations. The collinear analysis may still prove to be accurate because 25 days is sufficient to have detected most of the dominant eddy variability.

To derive the mean sea surface, GEOS-3 and SEASAT had several ground tracks that intersected at a common earth location. These crossover points were combined taking into consideration all possible orientations of the satellites, from both ascending and descending tracks. (Marsh, et al., 1984). One should note that an ascending (Asc) track went from southeast to northwest, and a descending (Desc) track traveled from northeast to southwest. This increased the number of data points compared to those available from the 8 sets of collinear tracks and allowed data from two separate missions to be used in calculating the mean sea surface.

In addition to the usual crossovers of (Asc-Desc) Geos and (Asc-Desc) SEASAT, the two satellites with different inclinations provided four new cross terms: (Asc Geos - Desc SEASAT), (Asc SEASAT - Desc Geos), (Asc Geos - Asc SEASAT), (Desc SEASAT - Desc Geos). A least-squares adjustment of the orbital bias was made to minimize the sum of the squares of the crossover differences. This reduced the relative errors in sea surface heights and defined the regional mean sea surface. (For further detail refer to Marsh, et al., 1984).

As shown by Figure 6 the ground tracks of Rev. 1157 and 1177 are located in a relatively calm section of the Southwest Pacific Basin, north of the Pacific-Antarctic Ridge. (Rand McNally, 1985). Figures 7-10 show the residuals for each ground track. Figure 7 and 8 show the collinear mean residuals, while Figure 9 and 10 show the mean sea surface residuals using the mean sea surface of Marsh et al., (1984) as a reference.

A spectral analysis was performed on both sets of residuals to describe the amount of power or kinetic energy in wavenumber space. This is similar to the analysis done by Fu (1983). Fu analyzed the characteristics of the ocean spectrum of mesoscale variability. He found residuals by removing a collinear mean and a best fit quadratic curve. He obtained wavenumber spectra by using a Fast Fourier Transform technique.

In this analysis the Fast Fourier Transform was also used. To smooth the data, and avoid power leakage from side-lobes, a modified Daniell window of half-width 2 was applied.

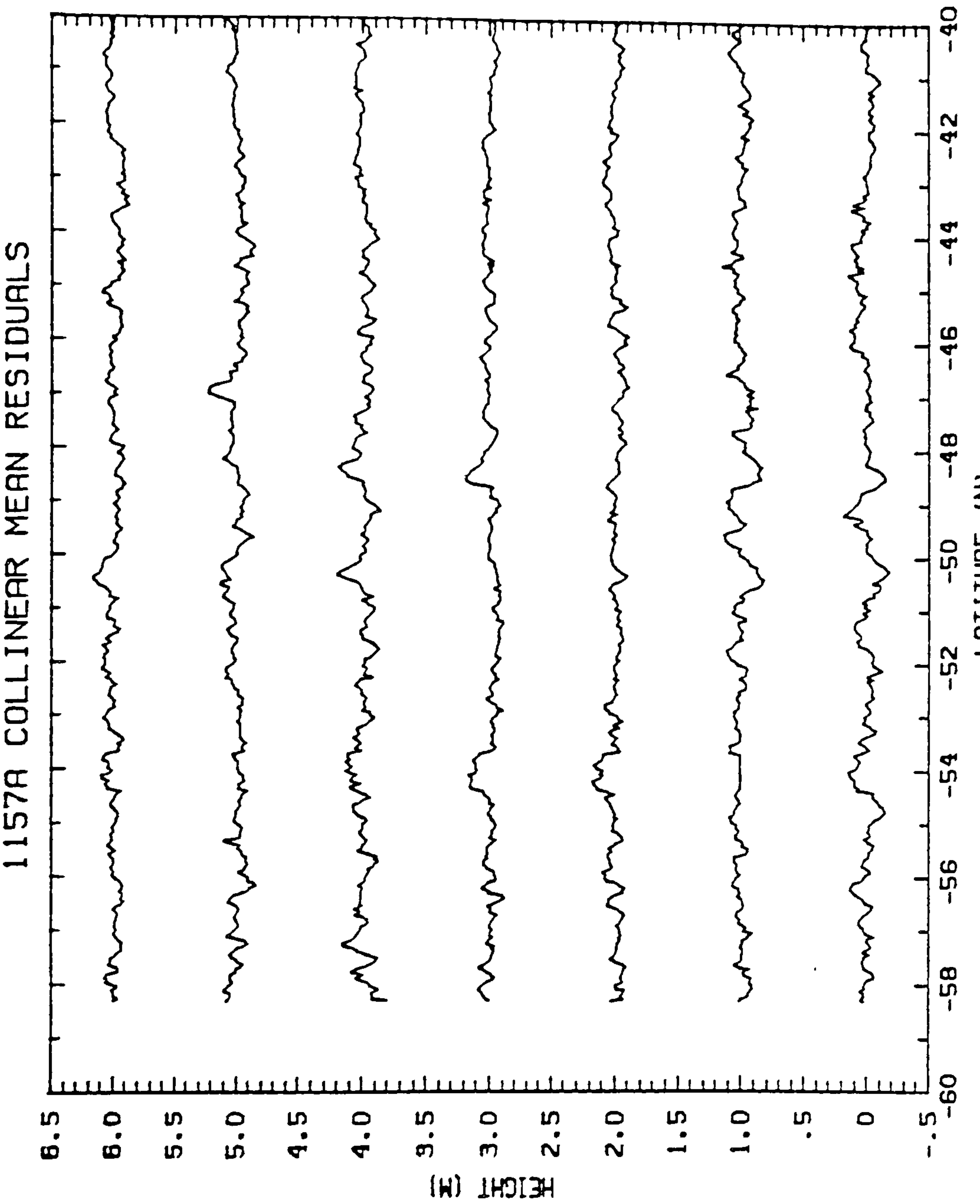

1157A COLLINEAR MEAN RESIDUALS
HEIGHT (M)
6.5
6.0
5.5
5.0
4.5
4.0
3.5
3.0
2.5
2.0
1.5
1.0
.5
0
-.5
LATITUDE (N)
-60
-58
-56
-54
-52
-50
-48
-46
-44
-42
-40

Figure 7

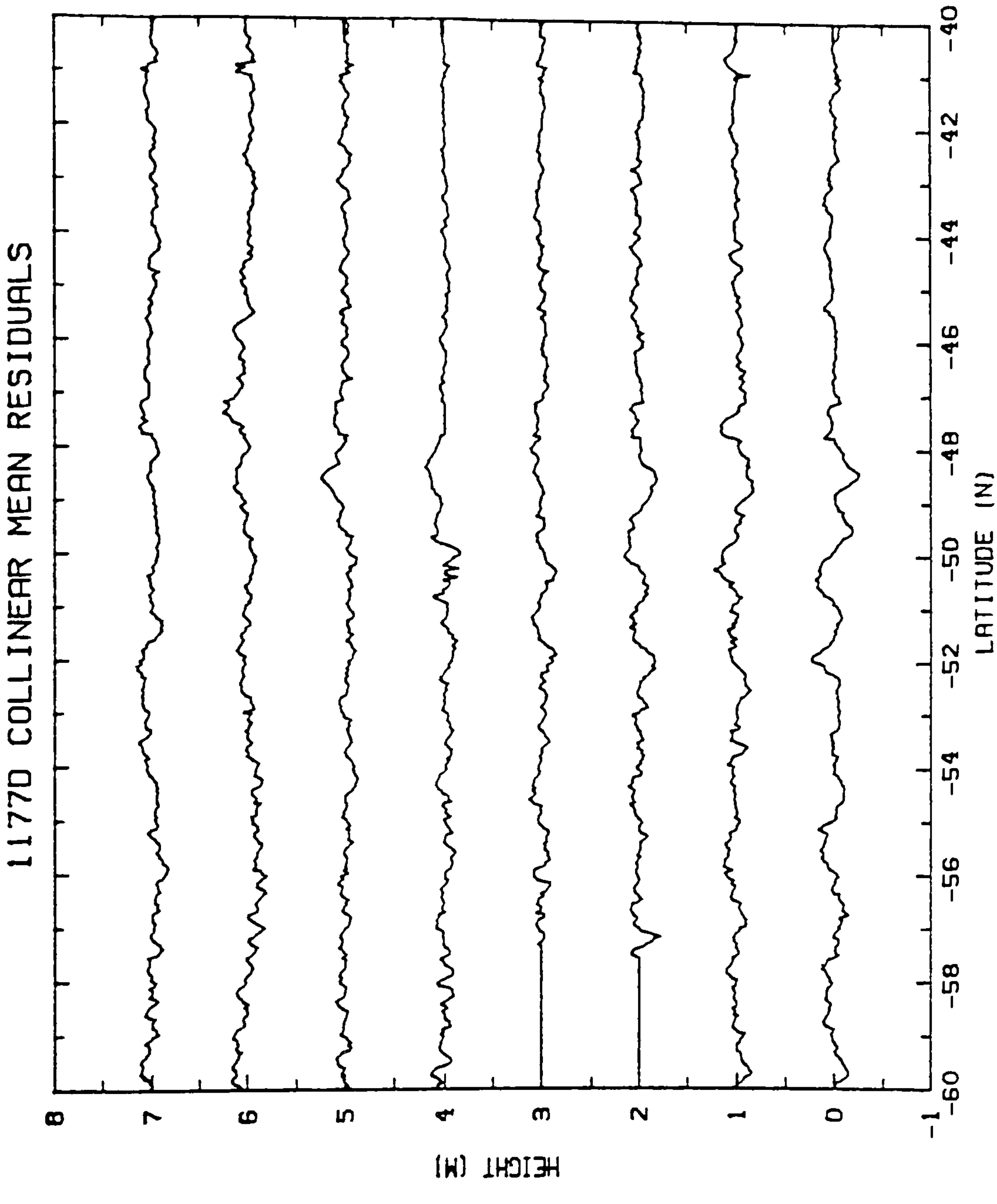

1177D COLLINEAR MEAN RESIDUALS
HEIGHT (M)
-1
0
1
2
3
4
5
6
7
8
LATITUDE (N)
-60
-58
-56
-54
-52
-50
-48
-46
-44
-42
-40

Figure 8

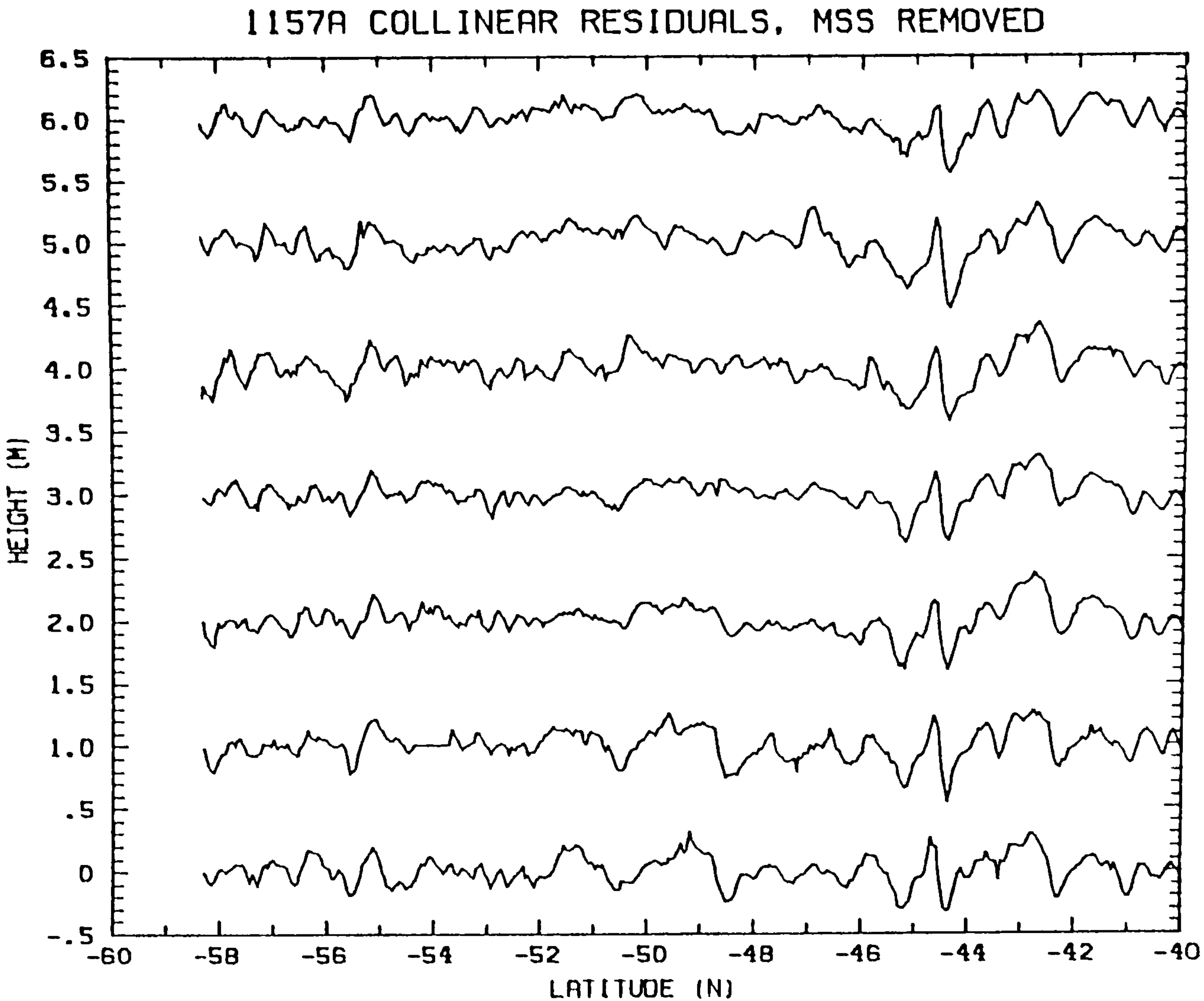

1157A COLLINEAR RESIDUALS, MSS REMOVED
HEIGHT (M)
6.5
6.0
5.5
5.0
4.5
4.0
3.5
3.0
2.5
2.0
1.5
1.0
.5
0
-.5
-60
-58
-56
-54
-52
-50
-48
-46
-44
-42
-40
LATITUDE (N)

Figure 9

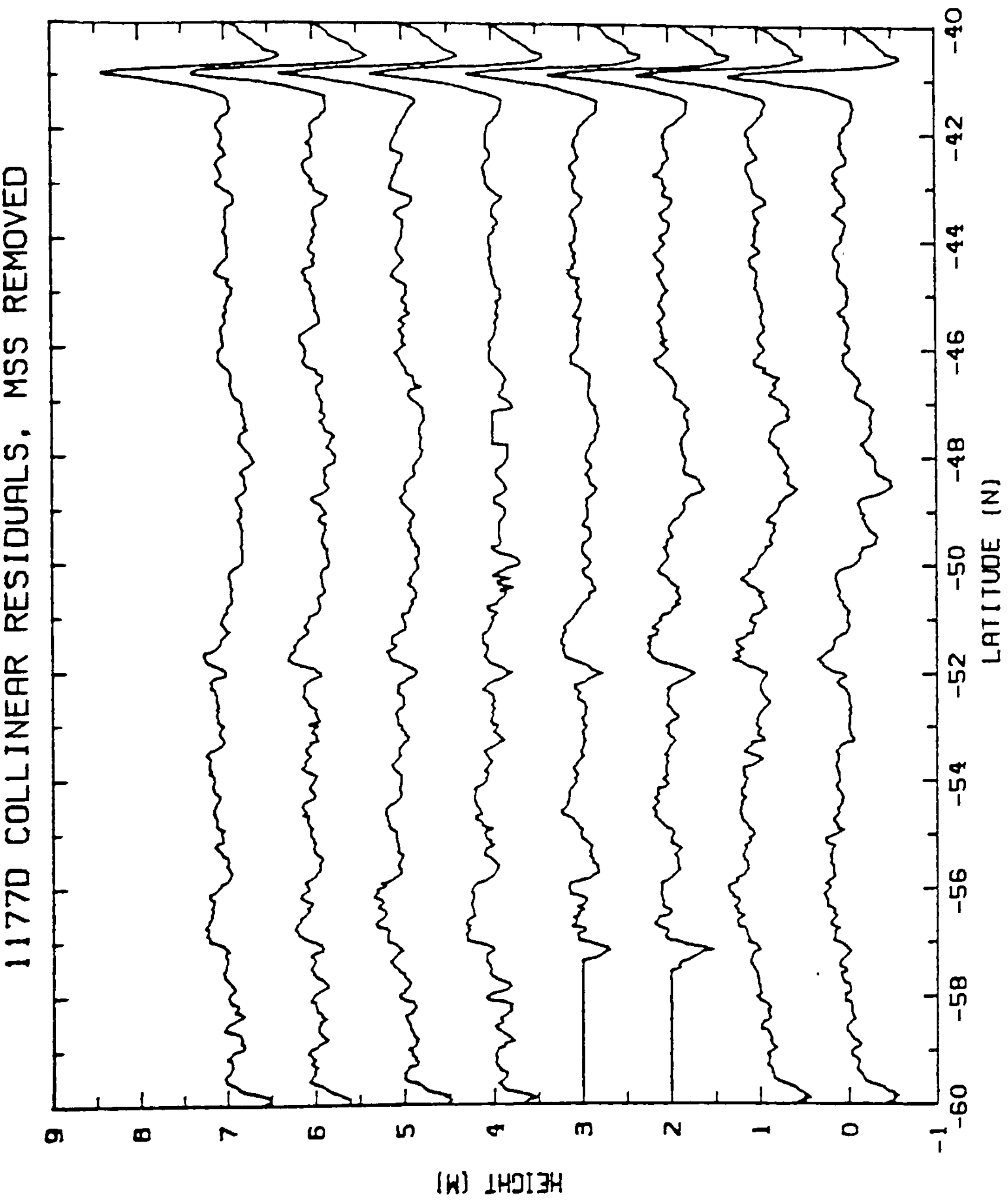

1177D COLLINEAR RESIDUALS, MSS REMOVED
HEIGHT (M)
-1
0
1
2
3
4
5
6
7
8
9
LATITUDE (N)
-60
-58
-56
-54
-52
-50
-48
-46
-44
-42
-40

Figure 10

The spectral computations were performed to provide a distribution of sea surface height variability as a function of wavenumber. The variability of the sea height is directly related to that of the geostrophic surface current, hence the spectra also represent ocean current variability.

The results of the analysis are contained in the next section.

RESULTS

The comparison of the residuals using both the collinear mean and mean sea surface methods showed that they were very different. They provided similar results in a qualitative sense. However, quantitatively the residuals differed by approximately an order of magnitude.

Observing the spatial domain, the computed collinear mean and the mean sea surface showed the same general features (Figures 11-16). Looking at track 1177, on Figures 12 and 14, there is a sharp rise at 41 degrees south latitude. Remembering that these two means are each an estimate of the geoid, then this rise must be some local change in the geoid. From the four or five meter magnitude, it is reasonable to suspect some topographical feature. In fact, an examination of bottom topography does show a large and sharp seamount on the ocean floor in this area.

Concentrating on this location, it can be seen that the peak in the collinear mean pass is much sharper and of greater magnitude than the peak in the mean sea surface. Figure 16 quantifies these differences with a range of 0.6 m to -1.4 m. The main reason for these discrepancies is that the mean sea surface uses a smoothing function too broad for this area. The effect of this is discussed in the conclusions.

For both methods, the spectra of the residuals have similar patterns. However, the power of the mean sea surface residuals were an order of magnitude higher than the residuals for the collinear mean. This occurred over all wavenumbers, and is shown in Figures 17-20.

The spectra of the residuals obtained from the mean sea surface were on the order of 10^{-1} m^2 at low wavenumbers and 10^{-2} m^2 at larger wavenumbers for rev. 1157 and its repeats, whereas the 1177 group had an energy on the order of 10^{-1} m^2 for all wavenumbers. The energy measured from the collinear mean residuals was respectively an order of magnitude less along both ground tracks.

One important observation to note is that in the region of 10^{-2} cycles per kilometer, the energy from the mean sea surface residuals was only two or three times that of the collinear mean residuals. This is important because it was hoped that the two methods would produce similar results when looking at oceanic variability in this wavenumber range. This is the range where most of the energy from the oceanic eddies exists for the southern ocean (Sciremammano, et. al., 1980). Figures 17-20 show that the concentration of energy is indeed the 90 to 100 km wavelength region.

Another way of presenting the differences between the methods is to calculate the variance of the residuals for each track. The variance of the residuals is equal to the area underneath the spectra. The variance of the residuals from the mean sea surface was an order of magnitude higher than that of the variance from the collinear mean residuals as shown in Table 1.

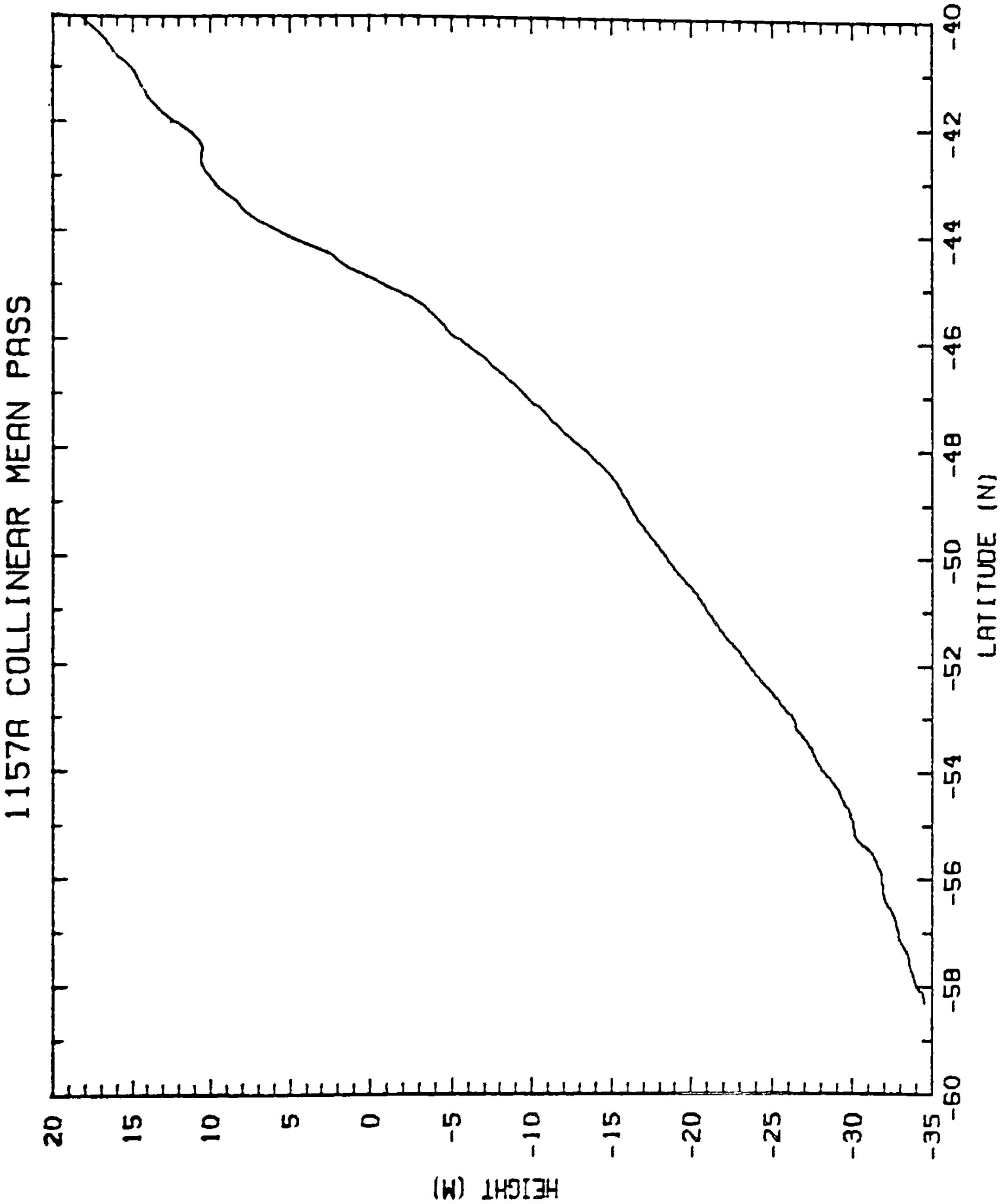

1157A COLLINEAR MEAN PASS
HEIGHT (M)
20
15
10
5
0
-5
-10
-15
-20
-25
-30
-35
LATITUDE (N)
-60
-58
-56
-54
-52
-50
-48
-46
-44
-42
-40

Figure 11

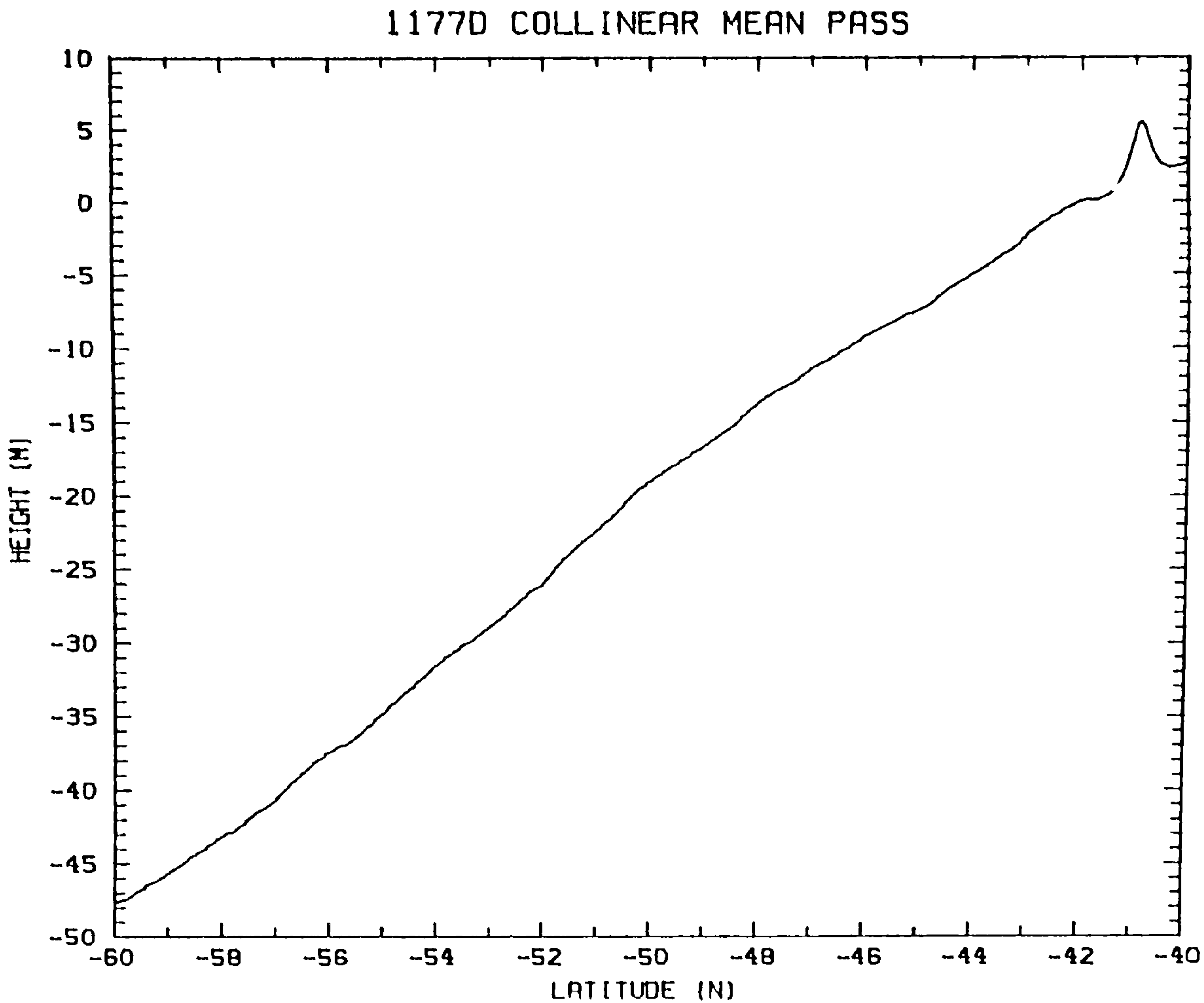

1177D COLLINEAR MEAN PASS
HEIGHT (M)
10
5
0
-5
-10
-15
-20
-25
-30
-35
-40
-45
-50
-60
-58
-56
-54
-52
-50
-48
-46
-44
-42
-40
LATITUDE (N)

Figure 12

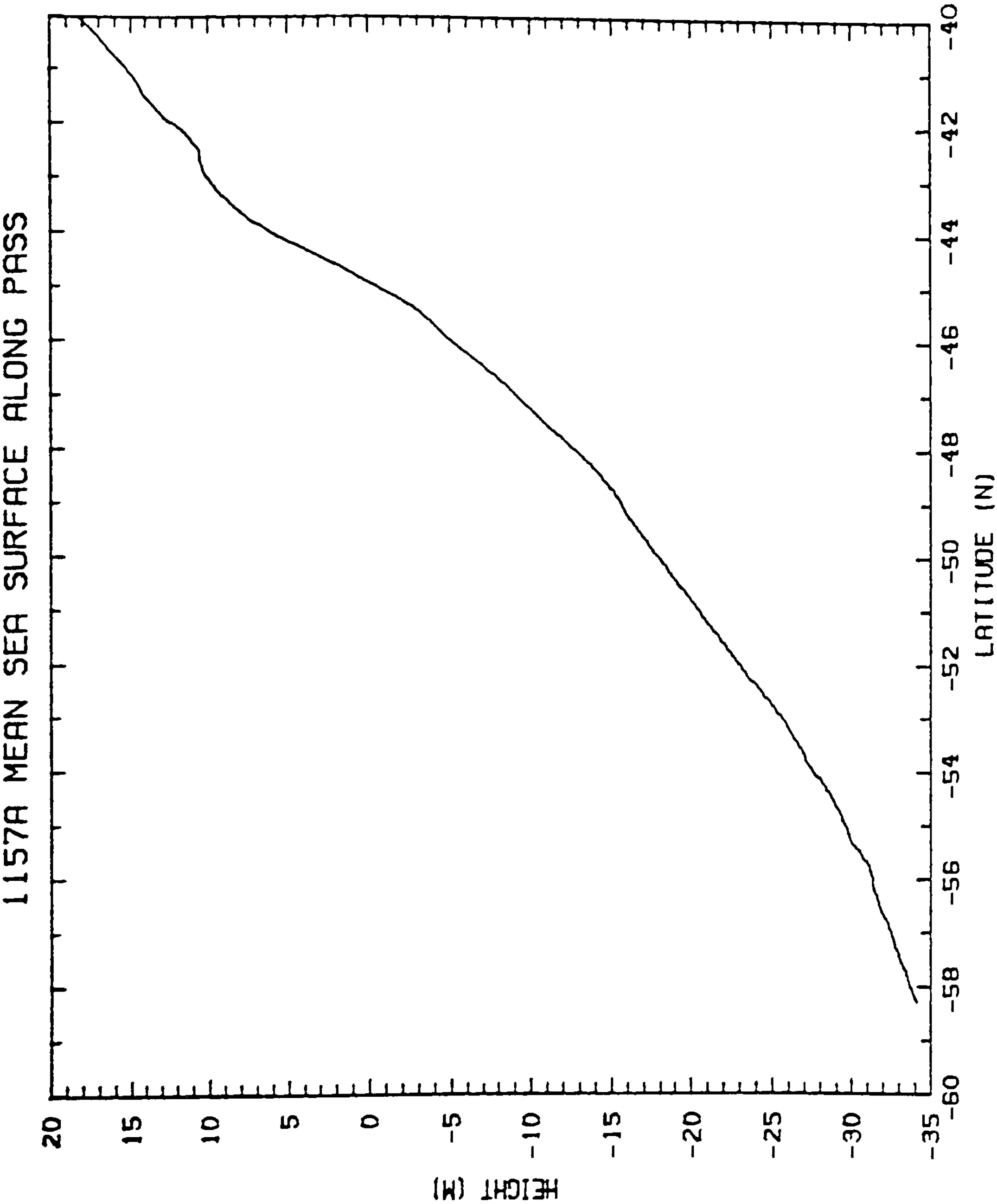

1157A MEAN SEA SURFACE ALONG PASS
HEIGHT (M)
20
15
10
5
0
-5
-10
-15
-20
-25
-30
-35
LATITUDE (N)
-60
-58
-56
-54
-52
-50
-48
-46
-44
-42
-40

Figure 13

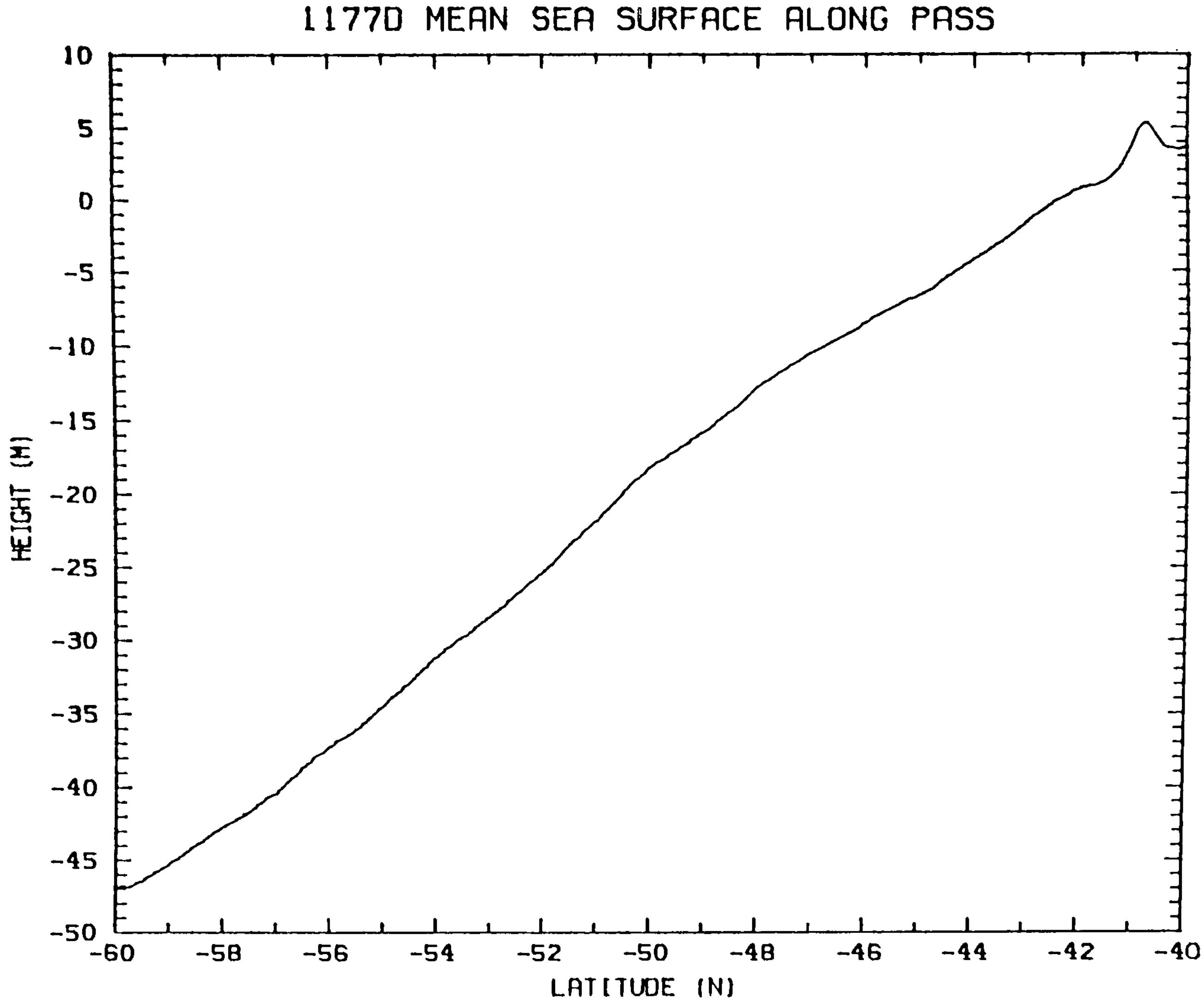

1177D MEAN SEA SURFACE ALONG PASS
HEIGHT (M)
10
5
0
-5
-10
-15
-20
-25
-30
-35
-40
-45
-50
-60
-58
-56
-54
-52
-50
-48
-46
-44
-42
-40
LATITUDE (N)

Figure 14

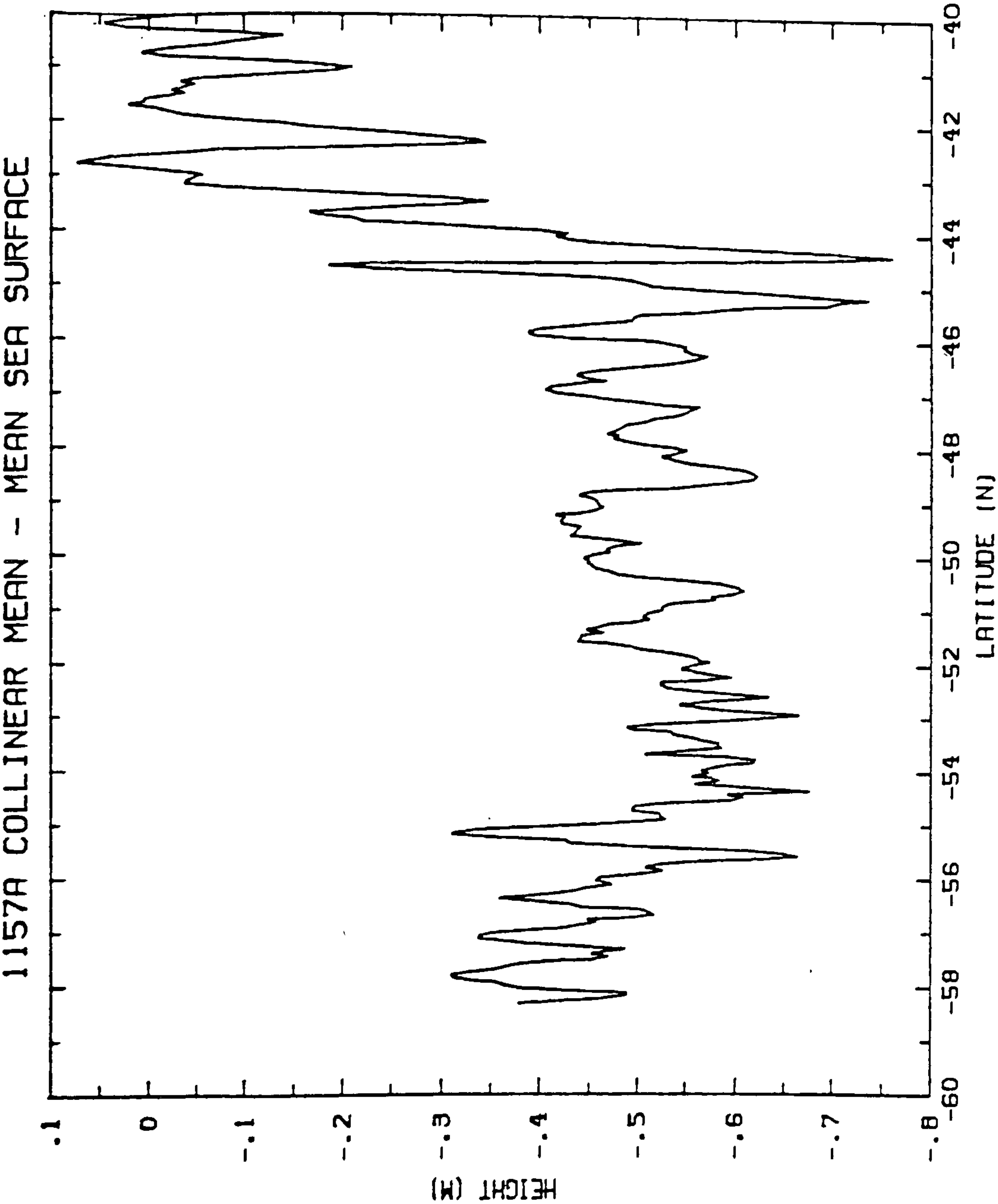

1157A COLLINEAR MEAN - MEAN SEA SURFACE
HEIGHT (M)
.1
0
-.1
-.2
-.3
-.4
-.5
-.6
-.7
-.8
LATITUDE (N)
-60
-58
-56
-54
-52
-50
-48
-46
-44
-42
-40

Figure 15

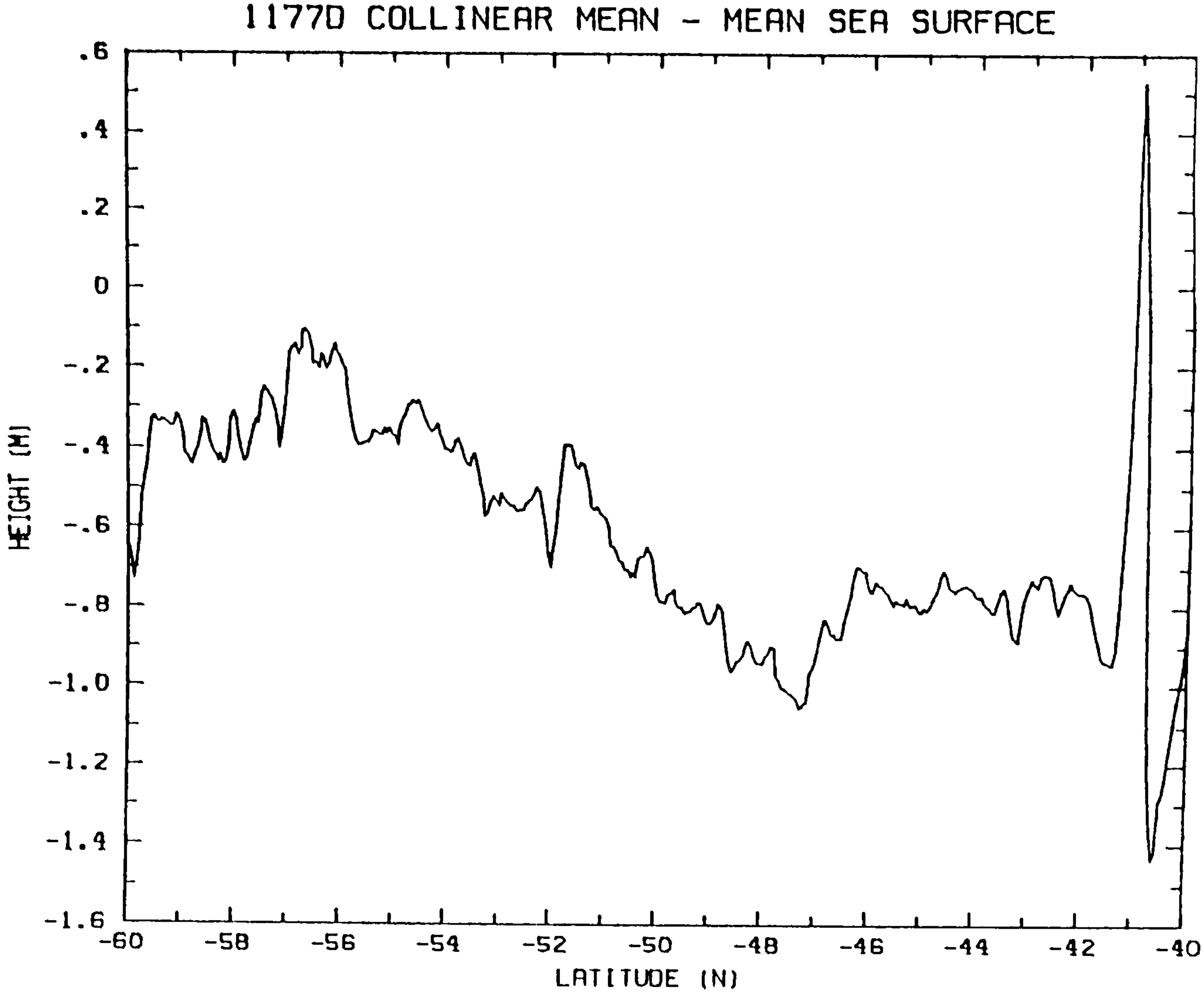


Figure 16

Note: The magnitude of spatial differences between the two means is larger for Ridge 77. Also, at 41 degrees South some oceanic phenomenon seems to be occurring.

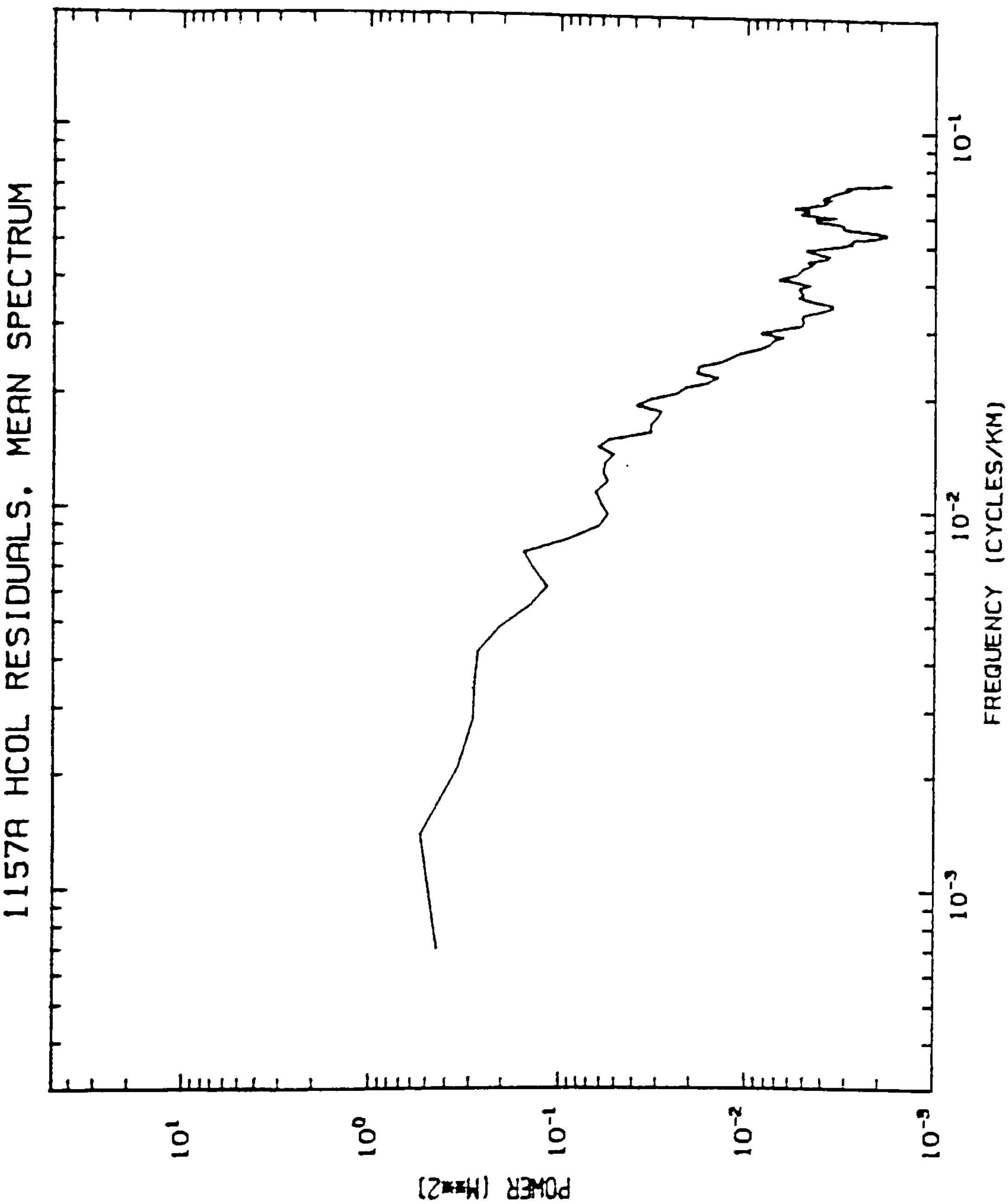

1157A HCOL RESIDUALS. MEAN SPECTRUM
POWER (M**2)
10^1
10^0
10^-1
10^-2
10^-3
FREQUENCY (CYCLES/KM)
10^-3
10^-2
10^-1

Figure 17

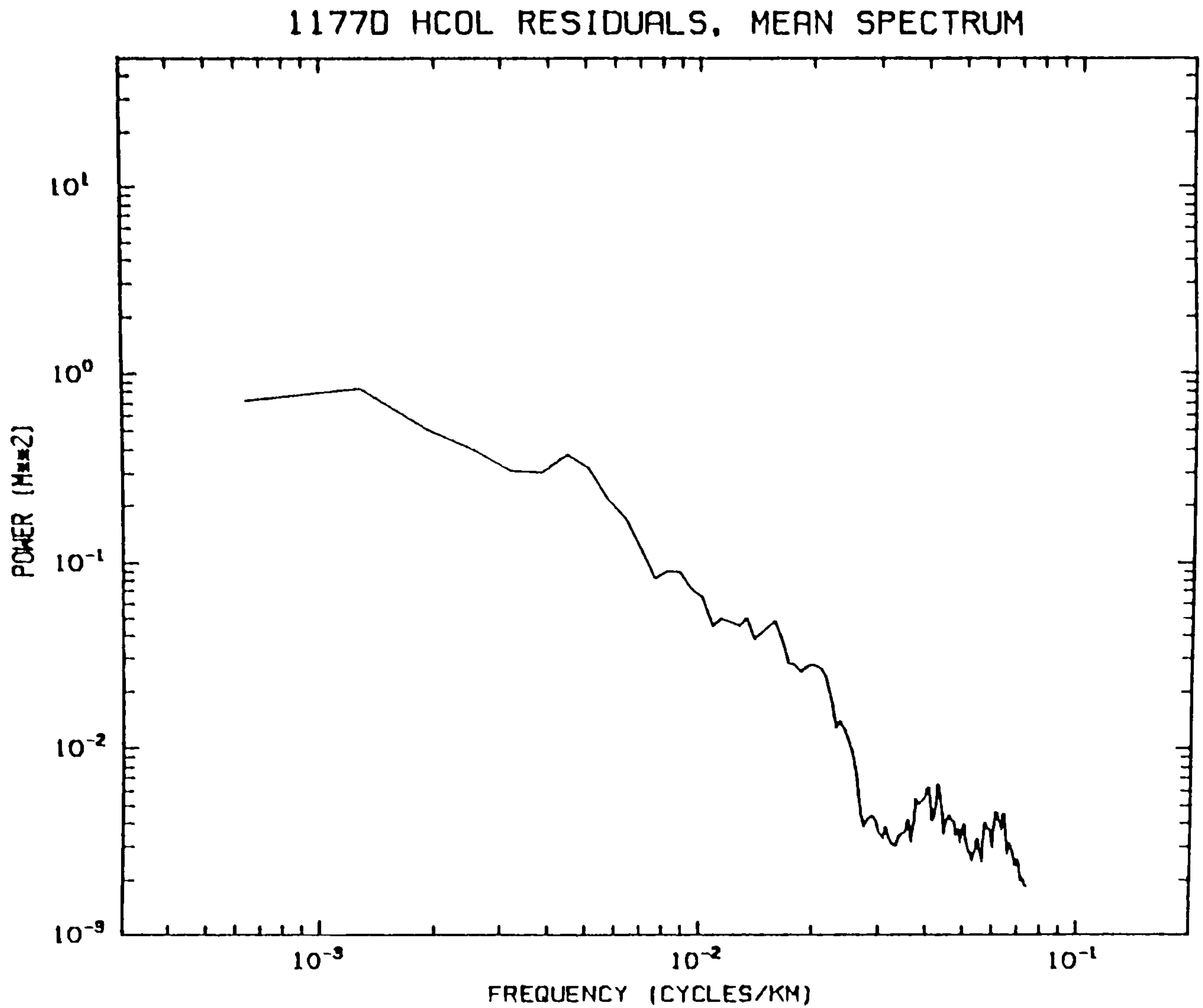

1177D HCOL RESIDUALS, MEAN SPECTRUM
POWER (M**2)
10^1
10^0
10^-1
10^-2
10^-3
10^-3
10^-2
10^-1
FREQUENCY (CYCLES/KM)

Figure 18

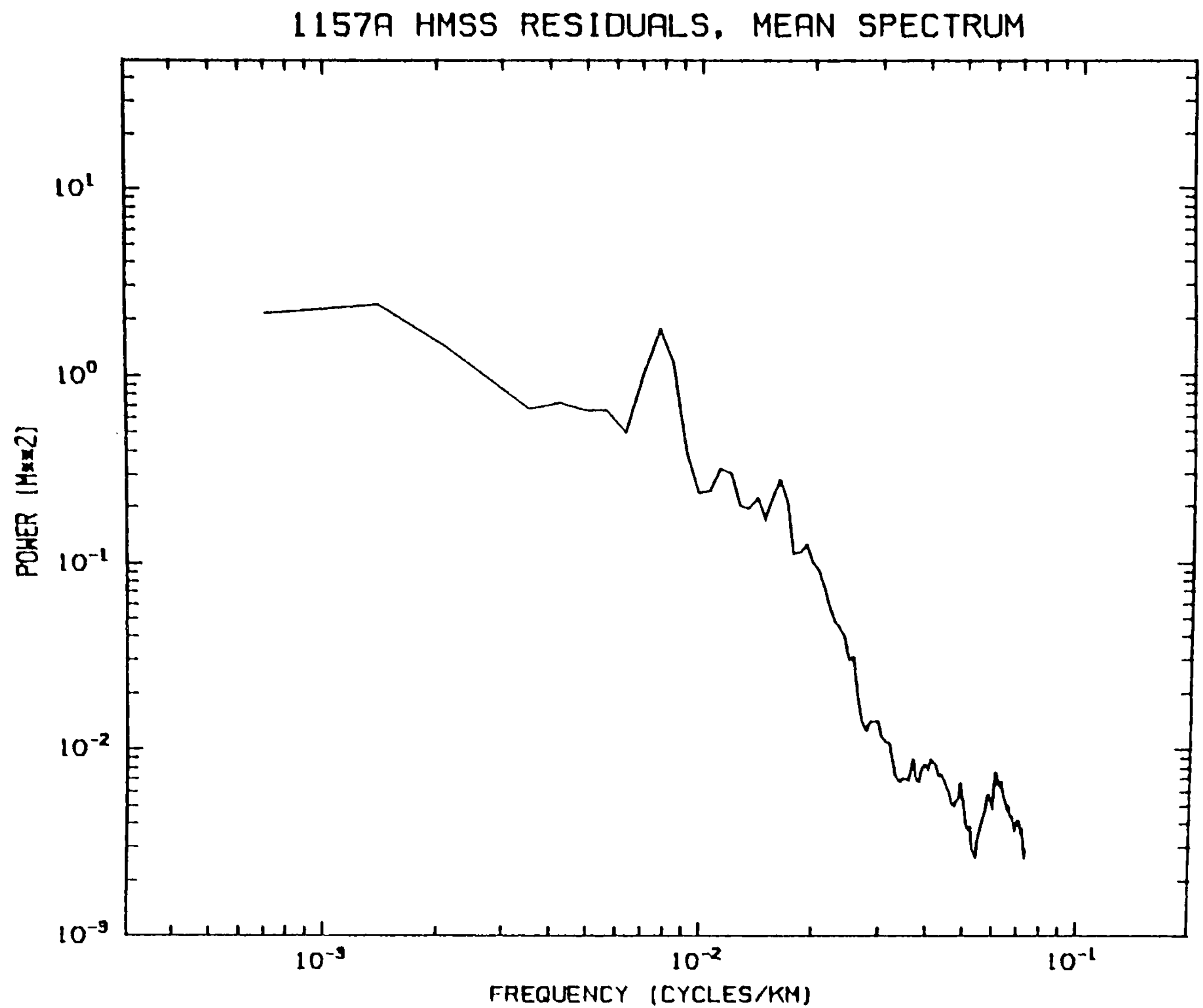

1157A HMSS RESIDUALS, MEAN SPECTRUM
POWER (M**2)
10^{1}
10^{0}
10^{-1}
10^{-2}
10^{-3}
10^{-3}
10^{-2}
10^{-1}
FREQUENCY (CYCLES/KM)

Figure 19

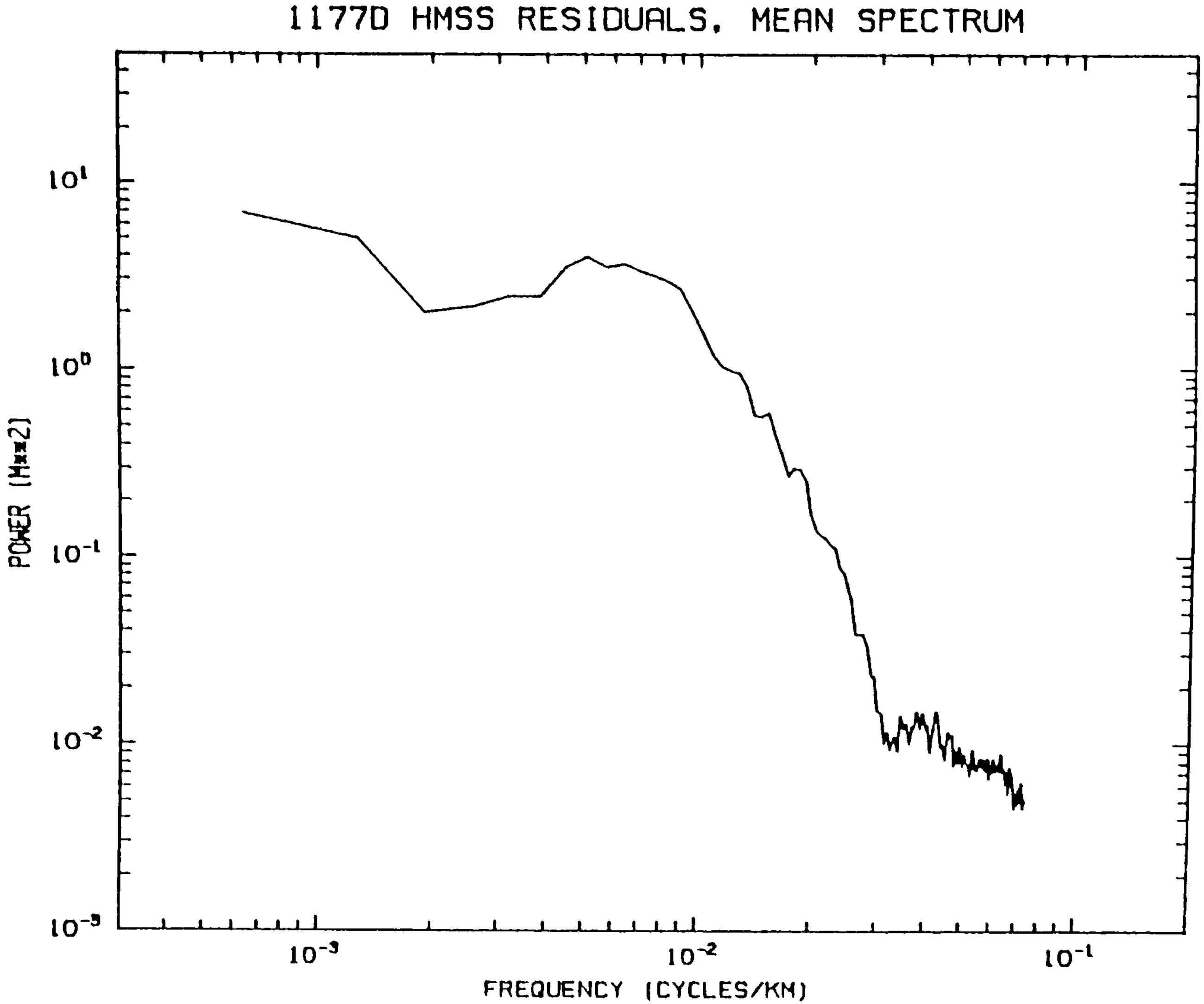

1177D HMSS RESIDUALS, MEAN SPECTRUM
POWER (M**2)
10¹
10⁰
10⁻¹
10⁻²
10⁻³
10⁻³
10⁻²
10⁻¹
FREQUENCY (CYCLES/KM)

Figure 20

Variance From the Mean (m**2)

Repeat Track Number	Original Track 1157 Mean Sea Surface	Original Track 1157 Collinear Mean	Original Track 1177 Mean Sea Surface	Original Track 1177 Collinear Mean
1	.011917	.002508	.041739	.003363
2	.015904	.003307	.035276	.005903
3	.015134	.003528	.038225	.002804
4	.010380	.002676	.036194	.002913
5	.012984	.002279	.031746	.001817
6	.013871	.003288	.036732	.003049
7	.012381	.003893	.047294	.004024
8	--	--	.047362	.005496
Average	.013224	.003068	.039321	.003671

Table 1. Comparison of Variance calculation from the mean using both methods.

Note that the variances from the mean sea surface method are an order of magnitude higher than those of the collinear mean.

CONCLUSIONS

It is expected that the Collinear mean method and the Mean Sea Surface method when applied to collinear data will yield the same or at least very similar results in the spatial and spectral domains. This was not the case for the two intersecting ground tracks analyzed here.

Both methods had similar gross features in the spectral domain, however, the mean sea surface residuals had an order of magnitude more power over almost all wavenumbers, than the residuals obtained from the collinear method. In the region of 90 to 100 km wavelengths, the two methods agreed more closely. This is advantageous because 90 to 100 km wavelengths contain the highest distribution of energy from oceanic variability in the geographic area under study here.

The collinear data was collected once every 3 days over a 25 day period. The mean sea surface of Marsh, et al. (1984) is derived from information gathered over the full 3.5 year GEOS-3 and the full 3 month SEASAT mission. Therefore, the mean sea surface had a larger data base. This was not an advantage in the Southwest Pacific Basin where the comparison was performed. When the surface was compiled by Marsh, et al. they used smoothing functions on various size grids of latitude and longitude, and a cap that covered the area of about three grid intervals, which were on the order of 1/8 degree each. (Marsh, et al., 1984). As shown by the data, there is a sharp sea-mount at the beginning of track 1177. This was detected and sharply defined by the collinear analysis, but smoothed out over a larger area in the mean sea surface analysis by the cap smoothing performed. The smoothing caused the spatial disturbance to start prematurely and travel farther along the track than the collinear method. The magnitude of this peak was also decreased in the mean sea surface method. Therefore, the smoothing functions utilized here cause a distortion in the topography, hence the geoid, contaminating the residuals formed by subtracting the mean sea surface.

In order to accurately and globally define a mean sea surface, new smoothing functions must be developed that do not wash out or distort either oceanic signals in smaller regions or sharp topographical features. When this is completed the mean sea surface method will prove to be more useful and can be used to facilitate data analysis. It will then be possible to analyze satellite altimeter data with a single pass and understand what oceanic phenomenon is occurring.

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